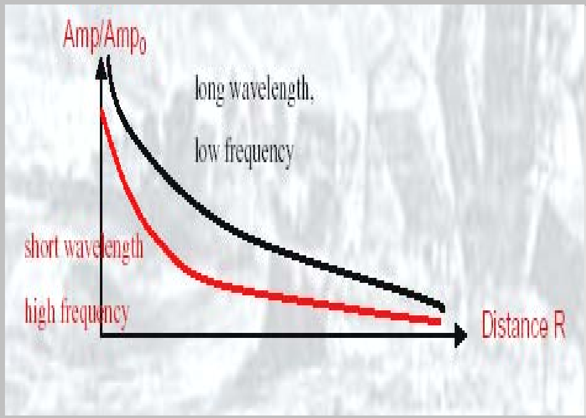
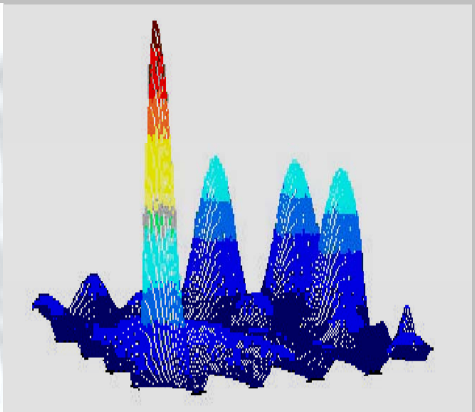
ADDIS ABABA UNIVERSITY
SCHOOL OF GRADUATE STUDIES

FACULITY OF SCIENCE
DEPARTMENT OF EARTH SCIENCES
FIELD OF STUDY GEOPHYSICS

SEISMIC WAVE ATTENUATION MEASUREMENTS ACROSS THE MAIN ETHIOPIAN RIFT AND ADJACENT PLATEAUS.

This thesis is submitted to the School of Graduate studies of Addis Ababa University in partial fulfillment of the Degree of Masters of Science in geophysics





BY GEREMEW LAMESSA

Addis Ababa University

July 2006

Acknowledgment

First of all, nothing is impossible with God! I thank my God Jesus Christ who is my savior and helper in every aspect of my life. Next to God, I have very much been gratified by the positive approaches and responses that I received from my advisor, Dr. Tilahun Mammo. His critical comments, technical help, and advice are important factors starting from the beginning up to the completion of this study.

I would also like to thank strongly my brother Engineer Abdissa Dessa for his financial help and encouragement and especially his regular hard work has been my energizer during my study.

I also express my hearty thanks to all my families for their encouragement and supporting me during my learning period, particularly my sister Hinkoshe Lemessa and my brother Keba Lemessa.

It is my pleasure to acknowledge and extend a special thanks to my brother Berhanu Beyene for his constant motivation and enormous help by producing conducive environment and providing me with advanced advice for the difficulties I faced while I was studying 24 hours at AAU faculty of science.

My thanks also goes to all members of Earth science department in general and particularly my respected

professors; Dr. Lackemariam, Dr. Atalay Ayele, Dr. Abera Alemu, Dr. Tigistu Haile and Dr. Tilahun Mammo who have Expended their unreserved efforts to make me acquainted with the skills of the field. I also mention the name of W/ro Ayenalem and the rest secretaries of our department for the over all supports they had provided during my stay as a postgraduate student.
I also like to extend my thanks to the following individuals for their cooperation in sharing their views for my study: - Ato Amenu Tolera (Msc), His brother Bikila Tolera (stu) and both his sisters (stu), all members of higher 23 S.S.S, specially the secretary of the school W/ro Engudy for her strong help in printing some parts of the manuscript, and the rest of my friends and colleagues whose name I haven't mentioned here.
I would like to thank also my colleague Fessah Ayele for giving me strong ideas on the seismic wave attenuation in general and his technical support for the completion of this study.

I would like to thank also registrar secretary of Adama teachers' college my sister Kassech Asfaw for printing some parts of the manuscript. My thank still goes to my colleagues Esthete Damte and my friend Abdulfetah Abdi for their strong motivation and advanced help in every difficulties I encountered during my study.

Geremew Lamessa

July 2006 A.A.U

ABSTRACT

As seismic wave propagates through earth materials, wave amplitude attenuates as a result of several factors. From measurement of amplitude of seismic data, attenuation measurement can be computed which can provide a viable tool for lithology discrimination and peak ground motion prediction for the region under study.

In this research, computation of attenuation coefficient is made from amplitude measurements done on the seismic data collected during the Ethiopian Afar Geoscientific lithopheric Experiment (EAGLE) which was undertaken in 2001-1003 project along profile I crossing the Main Ethiopian Rift, Western & Eastern Plateaus. For the generation of seismic waves, explosives were detonated inside boreholes and programmable seismic energy recorders called ‘‘ GURALP 6TD’’ were deployed on the surface at about 4-5 kilometer spacing across profile I. This study particularly focuses on surface wave attenuation. The recorded surface wave is processed so that the filtered amplitude is read appropriately.

1) In this study, we discuss the effect of the medium on seismic wave propagations (particularly seismic wave attenuation) from which we model seismic wave attenuation mathematically and analyze the 2D- graphical model of the region under study. In doing so, we see the interrelationship between the attenuation coefficient and distance, amplitude an distance, attenuation coefficient and frequency content of the wave, and dissipation factors Q^{-1} and the frequency. Finally, the relationship between attenuation coefficient and depth are also seen graphically and analytically. From the model, the attenuating behaviors of the region are discussed from the attenuation coefficient curve versus depth and evaluate the effects of the geology of the region on wave propagation for the Main Ethiopian Rift, Western and Eastern Ethiopian plateaus. The attenuation coefficient determined for the studied area varies as a function of both offset and frequency of the wave. Depending on the attenuating behaviors of the regions attempts have been made to determine the various lithologies.

)

Table of contents

Page

List of Tables

List of figures

Chapter 1
Introduction

As seismic wave propagates through earth materials, wave amplitude attenuates as a result of several factors. From measurement of amplitude of seismic data, attenuation measurement can be computed which can provide a viable tool for lithology discrimination and weak ground motion prediction for the region under study.

In this study we are concerned with seismic wave attenuation measurement: the loss of energy of a seismic wave as it propagates through the earth. Specifically, attenuation is defined as the loss of energy of a seismic wave as it travels though the earth, which is not caused by geometric spreading, but depends on the characteristics of the transmitting media. As an exploration tool, attenuation effects have only recently attracted attention. These effects can prove useful in two ways: as a means of correcting seismic data to enhance resolution of standard imaging techniques, and as a direct hydrocarbon indicator.

Many physical processes can lead to the attenuation of a seismic trace. These can be divided into two main categories: scattering and absorption, depending on the way energy is transformed. Scattering attenuation occurs because particles in the earth redirect the sound wave in other directions and happens when the scale of heterogeneities is smaller than the characteristic wavelength of the seismic wave. The energy is not dissipated, merely redirected. On the other hand, intrinsic attenuation is caused by absorption, or the conversion of acoustic energy into thermal energy. In other words, intrinsic attenuation is mainly due to friction losses. Absorption and scattering effects are difficult to separate; we assume most of the loss in energy is caused by intrinsic attenuation and focus on these effects exclusively. Intrinsic attenuation is caused by friction, particularly in porous rocks between fluid and solid particles.

Seismic attenuation is a fundamental property of subsurface media (Futterman, 1962; Strick, 1970; Kjartansson, 1979) that has a considerable impact on amplitude of recorded seismic data. As seismic wave propagates through earth materials, wave amplitude attenuates, because of the fact that the Earth is not perfectly elastic, i.e. propagating seismic waves attenuate with time due to various energy-loss mechanisms, such as movements along mineral dislocations or shear heating at grain boundaries (an elastic attenuation) as a dominating factor, geometrical spreading, scattering and others.

In this study, we discuss the effect of the medium on seismic wave propagations (particularly seismic wave attenuation) from which we model seismic wave attenuation mathematically and analyze the 2D- graphical model of the region under study. In doing so, we see the interrelationship b/n the attenuation coefficient, offset and frequency content of the wave and the relation ship between amplitude and offset. Finally, the relation ship between dissipation factors Q^{-1}, the frequency, and the attenuating behavior of the regions are discussed graphically and analytically.

Knowledge of attenuation can be very useful in lithology discriminations from the computed attenuation coefficient versus depth.

Seismic attenuation analysis is used to distinguish changes in the spectral and amplitude characteristics of the seismic signal with the purpose of identifying the presence of different rock types in the regions. The analysis of such changes can be used as a tool for characterizing rock properties and for understanding the behavior of wave propagation in the medium.

1.1 Objectives of the study

The major objectives of this work are:

1. To investigate the relation ship between the amplitudes of the waves, the distance from the shot to the detector and the attenuation coefficient α.
2. To investigate the relation ship between attenuation coefficient $\alpha(\omega)$ and the frequency content of the waves.
3. To make attenuation measurement contrasts across main Ethiopian Rift, Western and Eastern Plateaus by describing the attenuating behavior of layers under each region at various depths.
4. To develop 2D-graphical model of seismic wave attenuation (absorption attenuation) of the area under study.
5. To compute seismic dissipation factor Q^{-1} structure of the region.

1.2 Location of the study area.

The main Ethiopian rift (MER) is part of the East African Rift system (EARS) and comprise of Rift zones extending over a distance of 1000km from Afar triangle junction at the Red Sea Gulf of Aden to the Kenya rift.

The main Ethiopian Rift system strikes NNE from its southern extreme at (40 45") at lake chamo (which is among Ethiopian Rift valley lakes) and continues so for 650km until it reaches Ayellu volcano (at 90, 45’’) and change its orientation from NNW with out any discontinuity. (Mohor 1967, wolde 1989).

The West Ethiopian plateau is one of part of Nubian plate (Main African plate) and located at East side of it and situated to the west of East African rift system, demarcated by elongated fault lines .Is confined b/n the main Ethiopian Rift and the Sudan plain.

The East Ethiopian Plateau is found b/n The Main Ethiopian Rift and the Somali low land being separated   from the rift by sharp fault lines extended to its northern end.

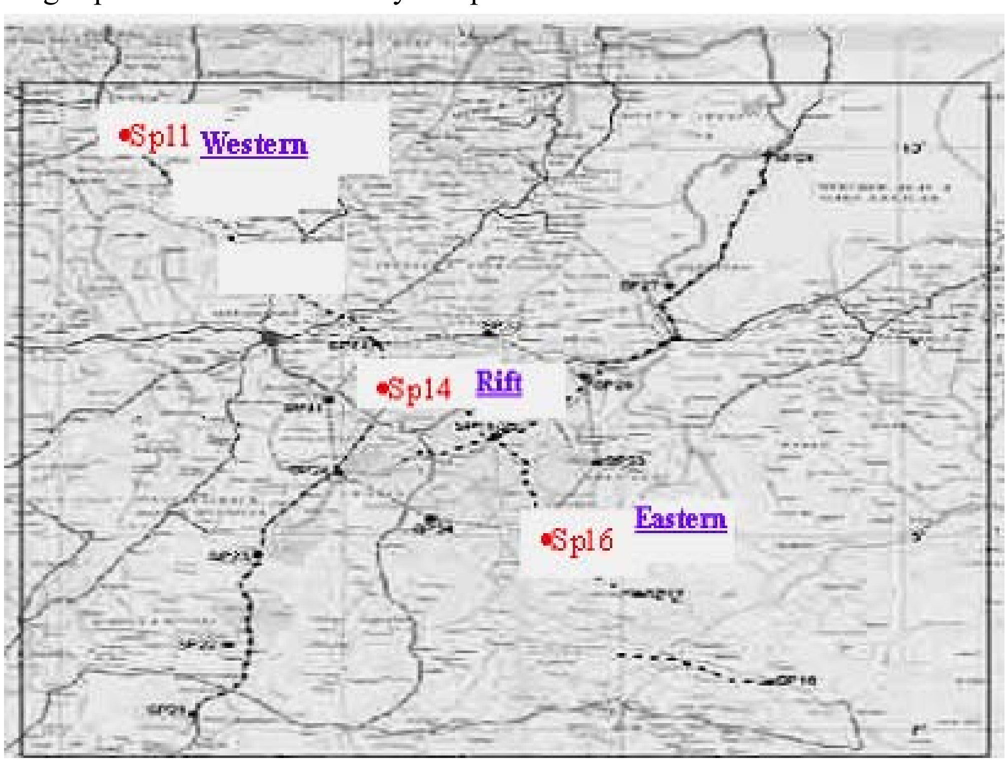


Fig 1.1 Location of the study area

1.3 Methodology

The Ethiopian Afar Geoscientific Lithospheric Experiment (EAGLE) discussed under chapter 3 was an international multidisciplinary study of the Ethiopian rift in particular the transitional process from continental to oceanic rifting along the Ethiopia rift at the horn of Africa in 2001-2003.The prime scientific objective of the EAGLE was to determine at what stage & in what way the process of controlling magma supply come to dominate Lithospheric faulting as continental rifting proceeds to sea floor spreading.

The controlled source seismic study of EAGLE project includes three profiles (EAGLE profile 1, 2, 3). This research is then based on seismic data collected during the EAGLE project along profile1.

In the project power gel C+ explosive was used for the detonation of charges in drilled holes of 50m deep &50-inch diameter were used. Seismic waves generated by explosive charges detonated in boreholes travel in the earth with their amplitudes attenuate and refracted from geological interface back to the earth surface. These refracted seismic waves were recorded by the instrument Gurlap 6TD that deployed at a distance of (4-5) km across profile I

The seismic pulses at various Gurlap 6TD were Fourier analyzed and waveform data collection and amplitude measurement can be done using already developed computer software (Sack) for data processing from which attenuation measurement of the area can be computed using spectral ratio method (amplitude ratio with offset and frequency variation). The technique requires that we first chose a reference geophone at which initial elastic wave is assumed to be generated at a position x_o with reference amplitude reading Ao(xo) or Ao(f) this is taken as the first geophone with A1 amplitude reading (as it is verified by mathematical derivation under subtopic 5.5(b). The Amplitude A1(xo) or A1(f) as the initial elastic wave amplitude Ao and any other geophones from the corresponding shots with amplitude reading A(x)or A(f) are compared to get regional attenuation. Variation of amplitude b/n successive geophone is taken for local attenuation measurements.

1.4 Significance of the study

The result of the research can be used to the following area.

1. The evaluation of the effects of the geology of the region under the study on wave propagation (i.e. attenuation) can help geo-technical engineer to design earth quake resistant structure and seismologist to understand strong ground motion prediction in the area.
2. It will be a reference for mapping of seismic attenuation model for the whole countrythat may help Ethiopian seismologist to make seismic hazard analysis of the area using their own attenuation relation, since we now have no attenuation model for the country.
3. It will give us a nice understanding of how attenuation coefficient varies with Frequency and offset.
4. The computed attenuation curves are used as the input for an inversion problem to determine the dynamic engineering properties of the layered soil profile by first determining layered shear modulus and damping ratio profiles.

1.5 The scope of the study

The seismic attenuation measurements that is undertaken for the region under study starts from the general model of seismic wave attenuation of any wave types and the principles and laws derived for the study are equally be applied to all waves with emphasizes are given to surface waves. The focus of this study is on surface waves, because we think that surface waves give us reliable results for the objective we started with and these wave types are strongly related with earthquake hazard analysis and engineering works in which our interests lie to give a reference study for further research on the modeling of ground motion attenuation relation and to model site- specific near surface earth conditions to control the response of foundations and the structure to the earthquake in engineering works. This can be done by measuring attenuation coefficient of surface waves as a function of frequency and offset. The relation ship between the attenuation coefficient as a function of frequency and offset are modeled graphically and analytically for the region under study and as well as the measurement of the dissipation factor Q^{-1} of the region which is in actual sense the intrinsic property of the rock medium by which we can make litho logy discrimination.

1.6 Thesis organizations

The structure of the thesis is as follows. In chapter two we see geological setting of the study area: Main Ethiopian rift and adjacent plateaus (East and West). In chapter3 we review the Eagle Experiment on which this thesis is based and data acquisition.

In chapter 4 we analyze the some basic theories of seismic wave propagations and types of seismic waves. Seismic wave attenuation and attenuation mechanisms as well as the mathematical foundation on which this thesis based is analyzed and modeled in chapter 5. In chapter 6 data processing and 2D-graphical modeling of attenuation coefficient α structure as a function of frequency, depth, and offset of the region are investigated and interpreted. Conclusions and recommendations are given in chapter 7.

Chapter 2 Regional geological setting

The western Ethiopian plateau (WEP) is uniquely identified from other parts of the main African plat by its topographic nature and litho logical units because it is a place of high rise above sea level and mostly natured by volcanic origin rock units while most other parts of the main African plate are characterized by there low altitude and sedimentary origin rock units. It is characterized by successive lithologic units with an older age of 250 million years and Archaean creation with less folded and metamorphosed younger verities and classified in to three major categories (Paul A, major 1971) according to their age and kind. The *pre Cambrian orogenic mountain ranges* formed at the early times. Geologic history were believed to denudated resulting nearly penuplnated land because the Arabic - Ethiopian massive was a stable land mass, even if some local occurrence of Paleozoic stratum are occurring at some places (mother 1971). The above denudalation was regionally true. The most distinguishable litho logic units of late time is edaga glacial and enticho sand stones, which are, believed to be records of the Paleozoic era. The existence of unconformity is between Adegrat sand stone and under lying clasties is not greater significant. The Paleozoic time clasties and the Mesozoic Adgrat sand stones are with barely possible different station.

In Mesozoic time the horn of Africa including WEP had been exposed to the transgression of the sea. At this time the preciously existing Gondwanaland was over passed by large-scale massive transgression and its sinking epeorogery haooened to be in Triassic with its climax being during early year Jurassic. It was thus later than the end of Mesozoic that the rising of the entire horn of Africa was re-attained and become above the oceanic level (Vkazmin AA 1972).

The transgression of the sea is believed to be executed from South – East to Northwest direction because the transgress Faces vary from upper Triassic age in Ogaden (SE) to lower even middle Jurassic in Tigray (NW). The Adegrat sand stone formed during that geological time is presently found all over the country with fairly constant thickness and Small-scale variations are observed with irregularities in basement complex peneplaneation. In most central parts of Ethiopia, the Adegrat sand stone is found with average thickness of 500m while at some places it is as thick as 1000m while at others it is few meters.

The other rock unit abundant through out the western plateau is the Antalo limes stone. It is observed to be lying over the basement complex in northern and central Ethiopia and Every time under lain by the adigrat sand stone.

As the west Ethiopian plateau (WAP) is separated by a sharp fault line demarcation from the main Ethiopian rift (MER), it's genesis and evolution is closely bounded to NER –

MER is characrixed by a series of major rift zones (. etal l986) having an extended existence ranging from AFAR-TRIPPEL Junction (as its extreme northern limitation) at gulf of Eden – Red sea intersection the extreme south Ethio – Kenya boundary. The EAR system of which MER is part is elongating further to the south to Zambezi – River basin in Mozambique with total North to south stretching of 3200 km.

The MER system is striking NNE from its southern extreme at (40 45") at lake chamo (which is among Ethiopian Rift valley lakes) and continues so for 650km until it reaches AYELU volcano (at 90, 45'') and change its orientation from NNW with out any discontinuity. (Mohor 1967, Woldegebriel 1989).

The Genesis of North East Africa stated about 30 million years ago due to lethothermic forces induced from stationery and upper mantel originated very high thermal & post that made the upper mantel execute series of up lifts. As consequent of some mantel induced activates, three radial up lifts had developed. These three radial up lifts are identified as the RED SEA RIFT SYSTEM, The GULF OF EDEN RIFT SYSTEM and the most North African system. Among these three systems, the last one i.e. NAR system is relatively in active while RSR and GER systems are extremely active to be the diverging boundary of the Arabian pens Lula from North east Africa. These three rift systems form about 120^0 triple junction at AFAR Depression and made that point become a unique site of shallow tectonically active crust (Plummer 1998) and with several surface high thermal manifestations and Active Volcanism.

EAR complex has four major subs – parts which are the Lake Rudolf, Lake Stefanla, the main Ethiopian Rift (MER) and the Afar triple junction depression. The fact that ERA system is very young and active rift system and its being an origin of oceanic ridge system (E wing and Heazen 1956) made it to be characterized by various active volcanisms, weak and strong earth quakes unlike the west African rift complexes which is relatively older, and non – oceanic ridge system (mc cannel 1967) the EAR system is in a process of changing Geological picture of the African content.

The natural development and its distinguishing tectonics are forwarded to be in various manners including power tension (Di palola 1972,Woldegebriel et at 1990), transnational (Boccalettlet et at 1992) and oblique (Bonini et at 1997). The nature of active tectionism at Afar of MER is not limited .The characterization of active tectonics at afar triple junction is not limited to the tripped junction but also extended to the south being manifested by the occurrence of Ten major volcanic centers as well as some major earth quake distributions. Recent scientific studies have confirmed that specific point just south of Ankober (south shewa) on the western rift margin experienced intensive seismic (E Daly, A. Ayele EAGLE 2004) and the region may serve as accommodation zone where the north ward propagations MER links with southern red sea lifts.

The MER separates the western and the Eastern plateaus of Ethiopian and much believed to be widening in its cross over distance as it becoming ocean meat of the fitter Geologic scenario of the planet. As several scientific predications are confirming that MER (and mainly at MER) will develop to be the splitting apart of the two African plates and the genesis of new ocean along the strike like of the Ethiopian dome.

The Eastern Ethiopian plateau is part of East African plate (Somali an plate) that extend to Indian ocean and is a place of high elevation as compared to Somali an plain and is mainly characterized by sedimentary rocks and WEP is characterized by igneous rocks.

Chapter 3 The Eagle project

3.1 Overview

The Ethiopian Afar Geoscientific Lithospheric Experiment (EAGLE) was an international multidisciplinary study of the Ethiopian rift in particular the transitional process from continental to oceanic rifting along the Ethiopia rift at the horn of Africa in 2001-2003.The prime scientific objective of the EAGLE was to determine at what stage & in what way the process of controlling magma supply come to dominate Lithospheric faulting as continental rifting proceeds to sea floor spreading.

The controlled source seismic study of EAGLE project includes three profiles (EAGLE profile 1, 2, 3). The first profile 1, across the Ethiopian rift in the vicinity of Nazareth extending to the Blue Nile in NW and to Ginnir (in Bale region) in the south East and measured to be 400 km long striking from NW to SE Profile two extended along the rift from Awassa in the south to Gewane in the North. A dense network of instrument profile 3) was also able to be deployed around the intersection of the two profiles to provide a 3-D topographic image of the subsurface immediately beneath the new volcanic segment of Bosetti in the vicinity of Nazareth (fig 3.1).

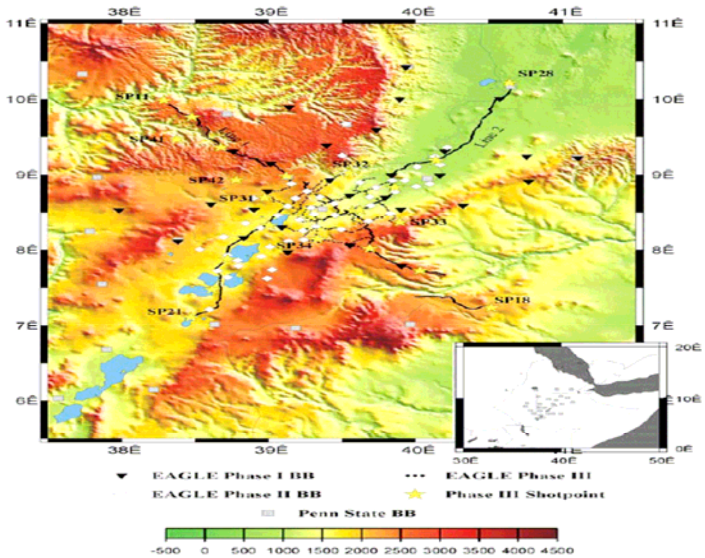


Fig 3.1 Eagle location map

The work involved recording of seismic 'echoes' from the controlled source and natural earth Quakes started since 2001 and finished by 2003, to provide an image of the top 100km or so of the earth over a region in compassing the northeastern part of the main Ethiopian rift where it enters Afar. The project was divided in to three phases where phase I involved 30 long term continuously recording seismic instruments being deployed over an approximately 250x250km^2 centered on the volcanic segment in the rift valley immediately to the north-east of Nazreth.Phase II began in September 2002 and involved the deployment of a further 50 recorders within the phase I array, concentrating the density of recording the center of this region, in particular to study local earth quakes to refine our understanding of the patterns and origin of local seismic activity.

3.2 The controlled source phase of the project (phase III) and Data acquisition

Phase three (III) was the controlled source seismic study that was under taken in January 2003.This project involved the deployment of nearly 1000 recorders over an 8-day period, which recorded the seismic waves generated from the detonation of 19 charges on both boreholes and lakes. Six scientific wells were drilled by Saba engineering Plc.in 2002.The holes were drilled to be 10’’ in diameter and 50 mt deep, though this configuration was not kept at some dill sites by the use of rotary rings through the use of DTH (Robinson, 1992) methods. At sites where more than one hole was drilled, the distance was kept at a minimum of 45mt so that the explosion from one would not detonate the charge in the other hole. Two holes were drilled at sites where they were needed to produce a higher energy from the detonation. Drilling of the wells was finished by 2003.To ensure the safety of the communities and their infrastructure, the holes were drilled far away from buildings, bridges, underground pipelines, electric mains, etc first by optimizing the safety and higher s/n ratio and the shot points are shown below in table 3.3 for this particular study. Except for Sp14, where there was blockage that disallowed the full unloading of the explosive, the drilling of these holes was successfully finished.

Two different types of seismic recorders were used in the project; Gualp 6 TD broadband system and Texan recorders. The Gualp 6 TD came as bundle with a 12v dry cell battery, solar panel, to energize these seismometers and a GPS clock receiver to synchronize their time round the clock. They were buried level and oriented north in hole approximately 50-75cm depths. They were deployed at a nominal spacing of 4-5km.Both the Guralp 6Td$_s$ and the Reftec Texans were deployed using a deployment sheet. This sheet not only helped in the easy recovering of the seismometers but also to identify the location of each seismometer for the understanding of the data it has recorded. The data recording of both kinds of recorders are transferred to a computer system before the active lifetime of the power unit is over and that the data will be permanently preserved. The transferred data stacked by the computer is then used for each specific seismic study and processed in accordance with the appropriate methodology.

The following shot points represent the main Ethiopian rift and adjacent plateaus; Sp11, Sp14, and SP16 shot points are used for this study as tabled below.

Shot points	Name	Easting	Northin	Zon	Elevation dx from shot
Sp 11	Goha Tsion	421571.4	110357	37	2508
Sp 14	Chefe Donsa	513316.3	992624.	37	2369.5
Sp 16	Kula	575890.5	886436.	37	2497.6

Table 3.1 Shot points with their position and elevation

Chapter 4 Theories of seismic wave propagations and types of seismic waves and their characteristics.

The seismic method utilizes the propagation of seismic waves through the earth. If we are to understand how seismic waves work and evaluate the information we get from it in geological terms, we must be familiar with the basic physical principles governing its propagation characteristics. These include its, transmission, absorption (anelastic Attenuation) in the earth materials and its reflection, refraction and diffraction characteristics at discontinuities and this study particularly focus on absorption (anelastic Attenuation). All these characteristics are because of propagation that depends on Elastic properties of earth materials (Rocks). Then we next discuss some of the basic concept of elasticity.

4.1 Theory of elasticity

The theory of elasticity relates the forces that are applied to the external surface of a body to the resulting changes in size and shape. The property of resisting change in size or shape and returning to the undeformed condition when external force is removed is called Elasticity. The relation between the applied force and the deformations are expressed in terms of the concepts of stress and strain. The relation between the stress and strain can describe the characteristics for a particular material such as velocity, attenuation, etc of wave propagation.

4.1.1 Stress and strain

Stress-is defined as force per unit area. If the force is perpendicular to the area, the stress is said to be *normal stress* (or pressure) and when it is parallel to the area element, it is called shear stress. For the purpose of theoretical modeling let's consider infinitesimal element of volume inside a stressed body, the stresses acting up on each of the six faces of the element can be resolved in to components as shown in fig 4.1.1.for two faces perpendicular to the x-axis. Subscripts denote the x-y and z- axes, respectively and δ_{yx} denotes the stress parallel to the y- axis acting up on the surface perpendicular to the x-axis and denotes *the shear stress*. When the two subscripts are the same (as with δ_{xx}) and denote *the normal stress.* When the medium is in static equilibrium, the stresses must be balanced i.e. the three stresses $\delta_{xx,}$ δ_{yx} and δ_{zx} acting on face OABC must be equal and opposite to the corresponding stresses shown on opposite face DEFG, with similar relation for the remaining four faces.i.e.

$\delta_{xy}= \delta_{yx}$, in general $\delta_{ij}= \delta_{ji}$ --(4.1.1)

The total stresses are $\delta_{xx,}$ δ_{yy} and δ_{zz} which are the normal stresses and δ_{xy}, δ_{xz}, δ_{yx}, δ_{yz}, δ_{zx} and δ_{zy} which are the shear stresses. Therefore, mathematically the total stress can be described by the 2nd rank Tensor as δ_{xx} δ_{xy} δ_{xz}

$$\sigma = \begin{vmatrix} \sigma_{xx} & \sigma_{xy} & \sigma_{xz} \\ \sigma_{yx} & \sigma_{yy} & \sigma_{yz} \\ \sigma_{zx} & \sigma_{zy} & \sigma_{zz} \end{vmatrix} \quad \text{---(4.1.2)}$$

For the stress tensor described, as above the diagonal elements δ_{xx}, δ_{yy} and δ_{zz} are *normal stresses* and the off-diagonal elements δ_{xy}, δ_{xz}, δ_{yx}, δ_{yz}, δ_{zx} and δ_{zy} are *shear stresses.*

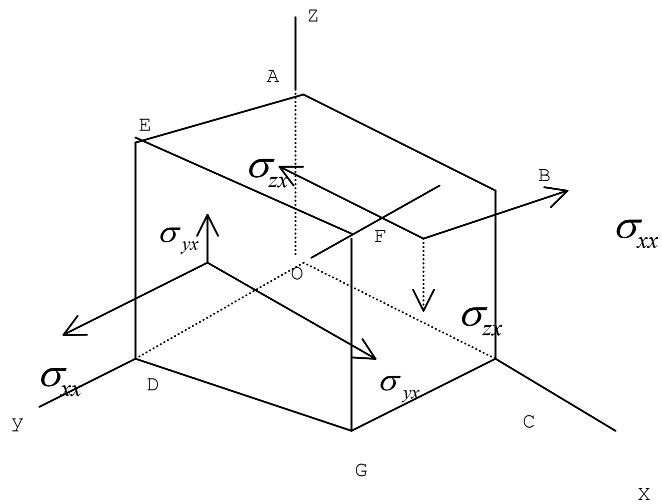


Fig 4.1 Components of stresses on faces Perpendicular to the x-axis.

Strain

When an elastic body is subjected to stresses, changes in shape and dimensions occur. Strain is defined as the relative change (the fractional change) in a dimension or shape of a body. By similar argument just like the derivation of stress above the relation between the strain and the fractional change in dimension can be done (Acki. Richard, 1980). For three displacements (u, v, w), which represent dimensions of a material element that change as a result of applied stress, the strains are thus;

$$\left.\begin{aligned} \varepsilon_{xx} &= \frac{\partial u}{\partial x}, \\ \varepsilon_{yy} &= \frac{\partial v}{\partial y}, \\ \varepsilon_{zz} &= \frac{\partial w}{\partial z} \end{aligned}\right\} \qquad \text{---(4.1.3)}$$

Normal strains

$$\left.\begin{aligned} \varepsilon_{xy} = \varepsilon_{yx} &= \frac{\partial v}{\partial x} + \frac{\partial u}{\partial y}, \\ \varepsilon_{yz} = \varepsilon_{zy} &= \frac{\partial w}{\partial y} + \frac{\partial v}{\partial z}, \\ \varepsilon_{zx} = \varepsilon_{xz} &= \frac{\partial u}{\partial z} + \frac{\partial w}{\partial x} \end{aligned}\right\}. \text{ -- (4.1.4)}$$

Shear strains

Thus the full description of the strains will have components and described as a 2nd rank tensor.

$$\varepsilon = \begin{matrix} \varepsilon_{xx} & \varepsilon_{xy} & \varepsilon_{xz} \\ \varepsilon_{yz} & \varepsilon_{yy} & \varepsilon_{yz} \\ \varepsilon_{zx} & \varepsilon_{zy} & \varepsilon_{zz} \end{matrix} \text{ --(4.1.5)}$$

In addition to these strains, the body is subjected to simple rotation about the three axes given by ,

$$\left.\begin{aligned} \theta_x &= \frac{\partial u/\partial y - \partial v/\partial z}{2}, \\ \theta_y &= \frac{\partial u/\partial z - \partial w/\partial x}{2}, \\ \theta_z &= \frac{\partial v/\partial x - \partial u/\partial y}{2} \end{aligned}\right\} \text{---4.1.6)}$$

The change in strain per unit volume Δ is $\Delta = \varepsilon_{xx} + \varepsilon_{yy} + \varepsilon_{zz} = \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = \nabla.\xi$ -(4.1.7)

4.1.2 Hook's Law and Elastic constants

Strain is always associated to acting stresses since dislocation (deformation) is always come to happen as a net force is acting on a given area. Such change in size & shape due to acting force is described by Hooke s law. Hook's Law states that a strain resulting on a body is *proportional to the* stress that produces it. The stress and strain can both be regard as second order (3x3) matrices so that the Hooke's law proportionality relating them is a fourth order tensor (6x6) matrices whose elements are elastic constants (Landau and

$(i = x, y, z)$

Lifshitz, 1986:32-51); in rock mechanics it is called *stiffiness.* This relation is mathematically described as $\sigma_{kl} = c_{ijkl}\varepsilon_{kl}$, where c_{ijkl} is elastic constant or (stiffness). When the medium is *isotropic*, i.e. when properties do not depend up on direction, it can be expressed in the following simple form (Love, 19944:102)

$\sigma = \lambda\Delta + 2\mu\varepsilon_{ii}$ --(4.1.7)

$\sigma = 2\mu\varepsilon_{ij} \quad (i, j = x, y, z; i \neq j)$ ---(4.1.8),

where the quantities λ and μ are known as Lam'e's constants. These equations are often expressed as a matrix equation, $\sigma = C\varepsilon$, as described below.

$$\begin{vmatrix} \sigma_{xx} \\ \sigma_{yy} \\ \sigma_{zz} \\ \sigma_{yx} \\ \sigma_{yz} \\ \sigma_{zx} \end{vmatrix} = \begin{vmatrix} \lambda + 2\mu & \lambda & \lambda & 0 & 0 & 0 \\ \lambda & \lambda + 2\mu & \lambda & 0 & 0 & 0 \\ \lambda & \lambda & \lambda + 2\mu & 0 & 0 & 0 \\ 0 & 0 & 0 & \mu & 0 & 0 \\ 0 & 0 & 0 & 0 & \mu & 0 \\ 0 & 0 & 0 & 0 & 0 & \mu \end{vmatrix} \begin{vmatrix} \varepsilon_{xx} \\ \varepsilon_{yy} \\ \varepsilon_{z} \\ \varepsilon_{xy} \\ \varepsilon_{yz} \\ \varepsilon_{zx} \end{vmatrix}$$

Stress Stiffness Strain------(4.1.9)

If we write $\varepsilon_{ij} = \sigma_{ij}/\mu$,hence μ is the measure of resistance to shearing strain and referred to *as the modulus of rigidity, incompressibility or shear modulus.*

According to Hooke's law stress and strain are linearly dependent to Each other and the body behaves elastically until the elastic limit is achieved. Below the elastic limit, on relaxation of stress, the body reverts to its pre-stress shape and size. Beyond the elastic limit when stress is large, the body behaves in a plastic or in a ductile manner and permanent deformation results to the extent of fracturing.

Hooke's law also confirms that stress & strain components at a particular point (and time) are dependent on the material's nature and each strain is a linear function of all independent component of the stress and vise versa.

Elastic constants

In addition to Lame's constants as described in equation 4.1.7 & 4.1.8, other elastic constants are also used for perfect description of the nature of seismic wave propagation. The most common are *young's modulus (E), Poisson's ratio* (σ)*, bulk modulus (k), and rigidity or shear modulus (μ)*. Let's define these elastic constants one by one. To define the frist two i.e. young's modulus (E) and Poisson's ratio (σ),we consider a medium in which all stress are zero except (σ_{xx}).Assuming (σ_{xx}) is positive ,dimensions parallel to $(\sigma_{xx})$is increase and dimensions normal to (σ_{xx}) decrease; this means that ε_{xx} (elongation in x-direction) where as ε_{yy} and ε_{zz} are negative and it can be shown that $\varepsilon_{yy} = \varepsilon_{zz}$ .We now define E and σ by the relations

1 Young's modulus (E)=$\frac{\sigma_{xx}}{\varepsilon_{xx}}$, ---(4.1.10)

2 Poisson's ratio$(\ \sigma\ ) = -\frac{\varepsilon_{yy}}{\varepsilon_{xx}} = -\frac{\sigma_{zz}}{\varepsilon_{xx}}$, --(4.1.11),

with the minus sign inserted to make σ positive. We define bulk modulus by considering a medium acted up on only by a pressure $\mathscr{P}$; this is equivalent to the stresses

$\sigma_{xx} = \sigma_{yy} = \sigma_{zz} = -\mathscr{P},\ \sigma_{xy} = \sigma_{yz} = \sigma_{zx} = 0$, the pressure causes the decrease in volume $\Delta \mathscr{V}$ and dilation $\Delta = \Delta \mathscr{V} / \mathscr{V}$; then k is defined as the ratio of Pressure to the dilation that it causes, that is

3 Bulksmodulus (K) = $-\mathscr{P}/\Delta$ -- (4.1.12),

with minus sign to make k positive. And 1/k is compressibility. By substituting the preceding values in Hook'es law, we get the following relation between E,σ and k and Lam'e's constants,λ and μ ;

E=$\mu\frac{(3\lambda + 2\lambda)}{\lambda + \mu}$, --(4.1.13)

$$\sigma = \frac{\lambda}{2(\lambda+\mu)}, \text{---}(4.1.14)$$

$$K = \frac{1}{3}(3\lambda + 2\mu) \text{---}(4.1.15)$$

4.2 Wave equations

Up to this point we have been discussing a medium in static equilibrium. We shall now remove this restriction and consider what happens when the stresses are not in equilibrium. When impulsive seismic wave acting on a medium is not in Equilibrium, the Elastic body will start to go under deformation. Because the net force acting the medium will make particles of the medium change there position as the body is experiencing a change in shape and size in a definite infinitesimal time interval. The action of the stress created motion of particles with in the body and in the process energy is transported from one point to the other, which is described by equation of motion. To derive a wave equation that describe the propagation of seismic wave in earth medium, we will consider fig 4.1 above and assume that the volume element will not be in the state of static equilibrium by the applied stresses. . The Stresses in the front and rare faces are: -

$\sigma_{xx,}$ σ_{yx}, and σ_{zx} & $\sigma_{xx} + \frac{\partial \sigma_{xx}}{\partial x} dx$, $\sigma_{yx} + \frac{\partial \sigma_{yx}}{\partial x} dx$, and $\sigma_{zx} + \frac{\partial \sigma_{zx}}{\partial x} dx$,because the stresses are acting opposite to each other that ,the net (unbalanced)stresses are ;

$[\ \sigma_{xx,} + \sigma_{yx}, + \sigma_{zx}\ ] - [\ \sigma_{xx} + \frac{\partial \sigma_{xx}}{\partial x} dx + \sigma_{yx} + \frac{\partial \sigma_{yx}}{\partial x} dx + \sigma_{zx} + \frac{\partial \sigma_{zx}}{\partial x} dx\ ] = \frac{\partial \sigma_{xx}}{\partial x} dx + \frac{\partial \sigma_{yx}}{\partial x} dx + \frac{\partial \sigma_{zx}}{\partial x} dx$. These stresses act on a face having an area ($dydz$) and affect the volume ($dxdydz$); hence we get for the net forces per unit volume in the directions of the x-, y-, and z- axes the respective values

$$\frac{\partial \sigma_{xx}}{\partial x}, \frac{\partial \sigma_{yx}}{\partial x}, \frac{\partial_{zx}}{\partial x}.$$

Similar expressions hold for the other faces; hence we find for total force in the direction of the x-axis the expression

$$\frac{\partial \sigma_{xx}}{\partial x}, \frac{\partial \sigma_{xy}}{\partial x}, \frac{\partial_{xz}}{\partial x}.$$

Newton's second law of motion states that the un-balanced force equals the mass times the acceleration; thus, we obtain the equation of motion along the x-axis:

$\rho \frac{\partial^2 u}{\partial t^2} =$ Unbalanced force in the x-direction on a unit volume.

$$= \frac{\partial \sigma_{xx}}{\partial x} + \frac{\partial \sigma_{xy}}{\partial x} + \frac{\partial_{xz}}{\partial x} \text{---(4.2.1),}$$

Where ρ is density (assumed to be constant). Similar equations can be written for the motion along the y- and z-axes. Equation (4.2.1) relates the displacement to the stresses. We can obtain an equation involving only displacements by using Hook's law to replace the stresses with strains and then expressing the strains in terms of the displacements, using eqs.(4.1.3),(4.1.4),(7,8) &(9) thus

$$\rho \frac{\partial^2 u}{\partial t^2} = \frac{\partial \sigma_{xx}}{\partial x} + \frac{\partial \sigma_{xy}}{\partial x} + \frac{\partial_{xz}}{\partial x}$$

$$= \lambda \frac{\partial \Delta}{\partial x} + 2\mu \frac{\partial \sigma_{xy}}{\partial y} + \mu \frac{\partial \varepsilon_{xy}}{\partial y} + \mu \frac{\partial \varepsilon_{xz}}{\partial z}$$

$$= \lambda \tfrac{\partial \Delta}{\partial x} + \mu \left[2\frac{\partial^2 u}{\partial x^2} + \left(\frac{\partial^2 v}{\partial x \partial y} + \frac{\partial^2 u}{\partial y^2} \right) + \left(\frac{\partial^2 w}{\partial x \partial z} + \frac{\partial^2 u}{\partial z^2} \right) \right]$$

$$= \lambda \frac{\partial \nabla}{\partial x} + \mu \nabla^2 u + \mu \frac{\partial}{\partial x} \left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} \ \right)$$

$$= (\ \lambda + \mu \ ) \frac{\partial \Delta}{\partial x} + \mu \nabla^2 u, \text{---(4.2.2)}$$

Where $\nabla^2 u$ is Laplacian of $u = \frac{\partial^2 u}{\partial x^2} + \frac{\partial^2 u}{\partial y^2} + \frac{\partial^2 u}{\partial z^2}$ (see eqs.15.14). By analog, we can write the equations for v and w:

$$\rho \frac{\partial^2 v}{\partial t^2} = (\lambda + \mu)\frac{\partial \Delta}{\partial y} + \mu \nabla^2 v, \text{---(4.2.3)}$$

$$\rho \frac{\partial^2 w}{\partial t^2} = (\lambda + \mu)\frac{\partial \Delta}{\partial z} + \mu \nabla^2 w. \text{---(4.2.4)}$$

To obtain the wave equation we differentiate these three equations with respect to x, y and z respectively, and add the result together. This gives

$$\rho \frac{\partial^2}{\partial t^2}\left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z}\right) = (\lambda + \mu)\left(\frac{\partial^2 \Delta}{\partial x^2} + \frac{\partial^2 \Delta}{\partial y^2} + \frac{\partial^2 \Delta}{\partial z^2}\right) + \mu \nabla^2 \left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z}\right), \text{ That is}$$

$$\rho \frac{\partial^{2\Delta}}{\partial t^2} = (\lambda + 2\mu)\nabla^2 \Delta \qquad \text{Or}$$

$$\frac{1}{{v_p}^2}\frac{\partial^2 \Delta}{\partial t^2} = \nabla^2 \Delta$$

---(4.2.5).

$${v_p}^2 = \left(\frac{\lambda + 2\mu}{\rho}\right) \Rightarrow v_p = \sqrt{\frac{\lambda + 2\mu}{\rho}} \text{ ---(4.2.6)}$$

By subtracting the derivatives of eqs. (4.2.3) With respect to z from derivatives of esq. (4.2.4) with respect to y, we get

$$\rho \frac{\partial^2}{\partial t^2}\left(\frac{\partial w}{\partial y} - \frac{\partial v}{\partial z}\right) = \mu \nabla_2 \left(\frac{\partial w}{\partial y} - \frac{\partial v}{\partial z}\right) \text{ , That is}$$

$$\frac{1}{{v_s}^2}\frac{\partial^2\theta_x}{\partial t^2} = \nabla^2\theta_x, \text{---(4.2.7)}$$

Where

$${v_s}^2 = \frac{\mu}{\rho}. \Rightarrow v_s = \sqrt{\frac{\mu}{\rho}} \quad \text{--(4.2.8)}$$

4.3 Seismic wave propagation

The propagation of waves is an important phenomenon that affects a medium when it is affected by a local excitation. From the observation of the propagation of the wave (mechanical wave) in the earth medium applied geophysics are used for Seismic Explorations in which the propagated wave will be picked and recorded by seismic instruments, so that the observed arrival of waves at other places will be analyzed and interpreted to understand geological information according to laws and principles of seismic theories.

The local excitation of a medium is not instantaneously detected at positions that are at a distance from the region of the excitation. It takes time for a disturbance to propagate from its source to other positions. This phenomenon of propagation of disturbances is well known from physical experience, and is usually referred as *wave propagation*. For example, a controlled source explosion, an earthquake or an underground nuclear explosion is recorded in another continent well after it has occurred. These examples illustrate mechanical wave motions or mechanical wave propagation.

For simplicity of the analysis we will consider one dimensional wave propagation as an introduction to basic concepts of wave propagation and its possible extension to layered media. Mechanical waves originate in the forced motion of a portion of a deformable medium due to some stresses. As elements of the medium are deformed the disturbance, or wave, progresses through the medium. In this process the resistance offered to deformation by the consistency of the medium, as well as the resistance to motion offered by inertia, must be overcome. As the disturbance propagates through the medium it carries a large amounts of energy in the forms of kinetic and potential energies (kinetic energy and strain energy will be discussed in 4.4). Energy can be transmitted over considerable distances by wave motions. The transmission of energy is effected because motion is passed on from one particle to the next and not by any sustained bulk motion of the entire medium. Mechanical waves are characterized by the transport of energy through motions of particles about an equilibrium position.

Deformability and inertia (theory of elasticity 4.1) are essential properties of a medium for the transmission of mechanical wave motions. If the medium were not deformable any part of the medium would immediately experience a disturbance in the form of an inertial force, or acceleration, upon application of a localized excitation. Similarly, if a hypothetical medium were without inertia there would be no delay in the displacement of particles and the transmission of the disturbance from particle to particle would be effected instantaneously to the most distant particle. All real materials are, of course, deformable and possess mass and thus all real materials transmit mechanical waves.

The inertia of the medium offers resistance to motion, but once the medium is in motion inertia in conjunction with resilience of the medium tends to sustain the motion. If, after a certain interval the externally applied excitation (stress) becomes stationary, the motion of the medium will eventually subside due to frictional losses (anelastc attenuation) and a state of static deformation will be reached. The importance of dynamic effects depends on the relative magnitudes of two characteristic times: the time characterizing the external application of the disturbance and the characteristic time of transmission of the disturbances across the body. As it has been derived above under the sub topic (4.2), the general one-dimensional wave propagation equation it can be generalize in three dimentions is, $\rho \frac{\partial^2 u}{\partial t^2} = \lambda \frac{\partial \theta}{\partial x} + 2\mu \frac{\partial^2 u}{\partial x^2}$.

4.4 Some Effects of the medium on wave propagation

- **Energy density and energy intensity I**

Probably the most important feature of any wave is the energy associated with the motion of the medium as the wave passes through it. Usually we are not concerned with the total energy of the wave but rather with the energy in the vicinity of the point where we observe; the energy density is the energy per unit volume.

It is clear that the wave field at a particular moment is a continuity of the field in the last moment. By using the concept of wave front, which is defied as the surfaces on which the wave motion is in the same phase, this observation is described as the Huygens' principle. Huygens' principle states that every point on a wave front can be regarded as a new source of waves. In Figure 4.3.1, the points P_1, P_2, P_3, P_4, and P_i, can be regarded as the news sources.

After an infinitesimal time of dt, the waves from each of the new source propagated to a distance of vdt, and forms a new wave front A'B'. Be aware of that between A and B there are infinite number of points so the newly formed wave front is a continuous phenomenon.

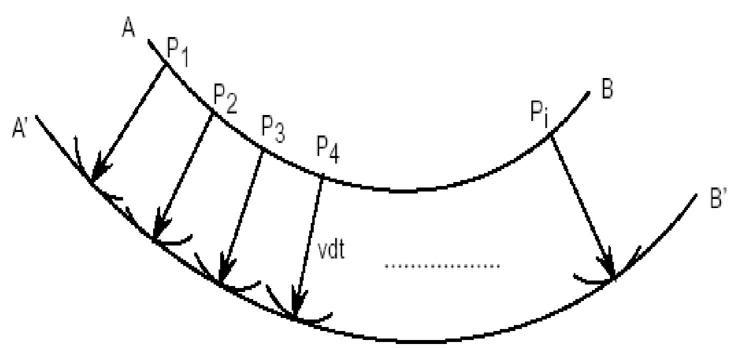


Figure 4 .2 Illustration of the Huygens' principle.

From Figure 4.3.1, it is easy to see that the shape of the wave front depends on medium's velocity. For the wave generated by a point source propagates in a medium with constant velocity, the wave front is in a shape of a spherical shell.

At one particular place r, the simplest wave displacement can be written as

$u = A\cos wt$, the displacement u ranges from $-A$ to $+A$ .Because the displacement varies with time, each element of the medium has the particle velocity at this point is $v = \frac{\partial u}{\partial x} = Aw\sin wt$ and an associated kinetic energy (K.E)= $\frac{1}{2}mv^2$ and kinetic energy per unit volume is *Energy density* given by

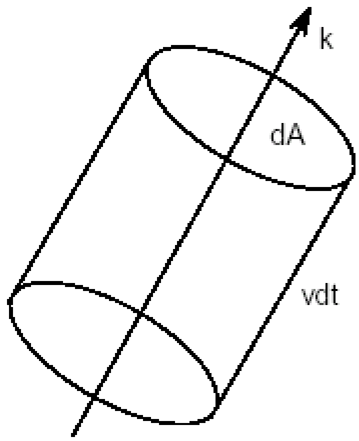
$$E_k = \frac{K.E}{V} = \frac{\frac{1}{2}mv^2}{V}, m = \rho V$$

$\Rightarrow E_k = \frac{1}{2}\rho v^2 = \frac{1}{2}\rho A^2 w^2 \sin^2 wt$**,** the expression varies from zero to a maximum of

$E_k = \frac{1}{2}\rho A^2 w^2$. The wave also involves potential energy resulting from the elastic strains created during the passage of the wave as the wave oscillates back and forth, the energy is converted back and forth from kinetic to potential forms, and the total energy remains fixed. Then the energy density for the wave is $E = \frac{1}{2}\rho w^2 A^2 = 2\pi^2 \rho v^2 A^2$ ------------(4.4.1), thus we see that the energy density is proportional to the first power of the density of the medium and to the second power of the frequency and amplitude of the wave.

- **Energy intensity I**: - Energy intensity is the total energy flow through a unit area in a unit time, so that it is the energy density we have learned above times the volume, then divided by the area and time.i.e.I=EV

Imagine a cylinder below, it happened have the wave energy propagation direction coincides with the axis of the cylinder



sothat $I = \frac{E_{total}}{dAdt} = \frac{EvdtdA}{dAdt} = EV = \frac{1}{2}\rho Vw^2 A^2 = 2\pi^2 \rho Vv^2 A^2$ ---------------------------(4.4.2)

where v is the propagation velocity of the waves. It is clear that energy intensity can also be called as *energy flow density*. The strain energy is a kind of the potential energy that stored in the elastic medium when it is deformed. From physical principle, the summation of the potential energy and the kinetic energy get to be a constant at any given moment. Thus for the sake of energy conservation, we should expect the total energy at any moment should be a constant depending on the amount of energy the source has radiated.

4.5 Types of seismic waves and their characteristics

Depending on the material medium through which they move there are two types of waves: linear and non-linear waves.

Linear waves: A wave propagating in a medium is governed by a three-dimensional wave equation, which is a second order linear partial differential equation has two major assumptions:

(i) There is no frictional resistance to particle motion

(ii) The displacement of the particle is small and does not exceed the elastic limit of the material. Thus the medium in which the wave propagates is described as frictionless and offers no resistance to wave motion. One example of this type of waves is light waves traveling in vacuum with a constant speed of light, c.

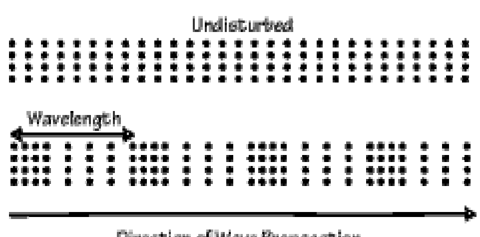
Non-Linear Waves: However, the above assumptions result in a drastic simplification of waves in nature. Non-linearity is rather the more common situation in wave travel (Elmore and Heald, 1985). The non-linearity is as a result of opposing forces such as the frictional forces that oppose particle displacement, causing loss of energy, spreading of energy and diffraction. The velocity of the waves is no more constant but depends on the wave frequency or equivalently on wavelength. Such a medium is then said to possess the property of dispersion. Dispersion of Non Linear Waves is a phenomenon where waves either traveling to different directions or having different frequencies propagate in different speeds. Dispersion is known to fall into two basic classes: intrinsic, which is based on anelasticity; and scattering, which is based on local wavelength-scale variations in the rock formation. Dispersion can occur both naturally (in a depressive medium) and or artificially by providing an artificial resistance to wave motion (Howard, 1993) such as shallow-water waves in restricted canals that result in turbulence, scattering, and dispersion at the exit point. The foundation of this thesis is then based on nonlinear waves in which there is anelasticity behavior of the earth material which in turn produce anelatic attenuation (5.3.2-A).

After saying this much about waves in general terms, lets come to the waves for the earth and their types. They are seismic waves. Seismic waves are: Any mechanical vibration sensed by personal perception is initiated from a source and travels to the location where the vibration is noted. The vibration is merely a change in the stress state due to some input disturbance. The vibration emanates in all directions that support displacement. The vibration readily passes from one medium to another, and from solids to liquids or gasses and in reverse. A vacuum cannot support mechanical vibratory waves, while electromagnetic waves transit through a vacuum. The direction of travel is called the ray, ray vector, or ray path. A source produces motion in all directions and the locus of first disturbances will form a spherical shell or wave front in a uniform material.

Seismic waves are also propagating vibrations that carry energy from the source of the shaking outward in all directions.i.e. They are tiny packets of Elastic Strain Energy that migrate out – ward from the source. Seismic waves can be distinguished by a number of properties including the speed the waves travel, the direction that the waves move particles as they pass by, where, and where they don't propagate. Based on these properties and difference in energy package, seismic waves are classified as follows: body waves and surface waves.

Body waves: These waves propagate throughout the entire body of the solid Earth. Body waves can be further subdivided into two different categories.

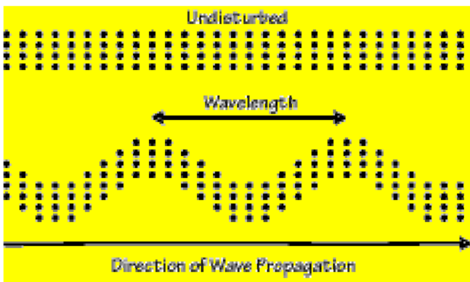
P-wave: It is a form of longitudinal wave and it is the fastest moving wave in any seismic event. Applied geophysics relies heavily on this class of waves. They are the first waves to arrive on a complete record of ground shaking because they travel the fastest (their name derives from this fact - P is an abbreviation for primary, first wave to arrive). They typically travel at speeds between ~1 and ~14 km/sec. The slower value corresponds to a P-wave traveling in water, the higher number represents the P-wave speed near the base of Earth's mantle. The vibration caused by P waves is a volume change, alternating from compression to expansion in the direction that the wave is traveling. P-waves travel through all types of media - solid, liquid, or gas. Elastic body waves passing through homogeneous, isotropic media have well-defined equations of motion. Most geophysical texts, including Grant and West (1965), include displacement potential and wave Equations. Utilizing these equations, computations for the wave speed may be uniquely determined. The velocity of a wave depends on the elastic properties and density of a material. If we let k represent the bulk modulus of a material, μ the shear-modulus, and ρ the density, then the P-wave velocity, which we represent by V_p, is given from equs. (4.2.6) by: $Vp = \sqrt{\dfrac{\kappa + \frac{4}{3}\mu}{\rho}}$



As a P-wave passes, the ground is vibrated in the direction that the wave is propagating.

Figure 4.3 P-wave propagation

S-wave: It is a form of transverse wave. Until the recent decade, it has not received much attention in applied geophysics, although it has been studied intensively in earthquake seismology since the beginning of the20th century. Secondary, or S waves, travel slower than P waves and are also called "shear" waves because they don't change the volume of the material through which they propagate, they shear it. They are transverse waves because they vibrate the ground in a direction "transverse", or perpendicular, to the direction that the wave is traveling.



y.

Figure 4.4 S- wave propagation

The S-wave speed, call it V_s , depends on the shear modulus (μ) and the density of the medium (ρ) and given as from equs. 4.2.8; $V_s = \sqrt{\frac{\mu}{\rho}}$. Even though they are slower than P-waves, the S-waves move quickly. Typical S-wave propagation speeds are on the order of 1 to 8 km/sec. The lower value corresponds to the wave speed in loose, unconsolidated sediment, the higher value is near the base of Earth's mantle.

An important distinguishing characteristic of an S-wave is its inability to propagate through a fluid or a gas because a fluids and gasses cannot transmit a shear stress and S-waves are waves that shear the material.

In general, earthquakes generate larger shear waves than compress ional waves and much of the damage close to an earthquake is the result of strong shaking caused by shear waves. Therefore we can see from the above discussions about body waves is that the distinguishing property of the propagation of body waves is the fact that the propagation velocity waves is independent of the wave frequency but determined by Elastic modulus and density of the rock medium through which they migrate.

Surface waves: surface impacts, explosions and waveform changes at boundaries produce surface waves. Besides body waves, elastic waves within the Earth can also propagate along surfaces (free surface) very much like water ripples. The surface waves may carry greater energy content than body waves. These wave types arrive last, following the body waves, but can produce larger horizontal displacements in surface structures. Therefore surface waves may cause more damage from earthquake vibrations. Two recognized disturbances, which exist only at "surfaces" or interfaces, are Love and Raleigh waves. Traveling only at the boundary, these waves *attenuate rapidly with distance* from the surface. Surface waves travel slower than body waves. Love waves travel along the surfaces of layered media, and are most often faster than Raleigh waves. Love waves have particle displacement similar to SH-waves. Raleigh waves exhibit vertical and horizontal displacement in the vertical plane of ray path. A point in the path of a Raleigh wave moves back, down, forward, and up repetitively in an ellipse like ocean waves.

The surface waves of interest to geophysics are those propagating at the surface of the Earth. Surface waves always arrive with much larger amplitudes than body waves. This thesis is in actual sense depends on this types of waves. The two basic types of Surface waves are; Raleigh and love waves:

Raleigh waves: Are developed by harmonic oscillators, as steady-state motion is achieved around the oscillator's block foundation. It travels at a velocity of about 90% of the S-wave velocity (e.g., VR = 0.9194 VS). It travels with a retrograde elliptical motion (i.e., vertical displacement is about 1.5 times the horizontal displacement.). It is usually responsible for most of the earthquake damages because of its vertical motion and its large amplitude. Raleigh waves are the slowest of all the seismic wave types and in some ways the most complicated. Like Love waves they are depressive so the particular speed at which they travel depends on the wave period and the near-surface geologic structure, and they decrease in amplitude with depth and with offset along the surface. Typical speeds for Raleigh waves are on the order of 1 to 5 km/s. Raleigh waves are similar to water waves in the ocean (before they "break" at the surf line). As a Raleigh wave passes, a particle moves in an elliptical trajectory that is counterclockwise (if the wave is traveling to your right). The amplitude of Raleigh-wave shaking decreases with depth.

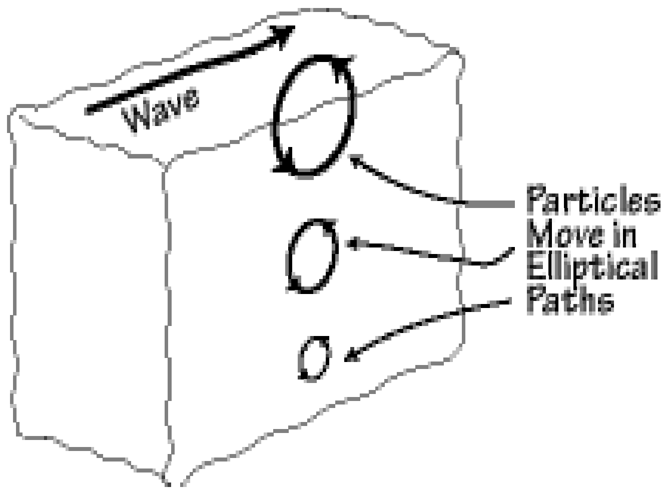


Figure 4.5- Raleigh wave motion (propagation)

LOVE WAVES: Love waves are horizontally polarized surface waves. They exist only on a layered Earth with a shear wave velocity of the top layer smaller than that of the substratum. It travels somewhat slower than the Raleigh waves and its velocity is frequency dependent (i.e., depressive). Because of its complexities, Love waves are not as heavily used for geophysical analyses, unlike P-, S- and Raleigh waves. Love waves are transverse waves that vibrate the ground in the horizontal direction perpendicular to the direction that the waves are traveling. They are formed by the interaction of S waves with Earth's surface and shallow structure and are depressive waves. The speed at which a depressive wave travels depends on the wave's period. In general, earthquakes generate Love waves over a range of periods from 1000 to a fraction of a second, and each period travels at a different velocity but the typical range of velocities is between 2 and 6 km/second. Love waves are transverse and restricted to horizontal movement - they are recorded only on seismometers that measure the horizontal ground motion.

Figure 4.6 - Love wave motion (propagation)

The following table summarizes types of seismic waves with their properties

Wave Type (and names)	Particle Motion	Typical Velocity	Other Characteristics
P, Compressional, Primary, Longitudinal	Alternating compressions ("pushes") and dilations ("pulls"), which are directed in the same direction as the wave is propagating (along the ray path); and therefore, perpendicular to the wave front.	VP ~ 5 – 7 km/s in t ypical Earth's crust; >~ 8 km/s in Earth's mantle and core; ~1.5 km/s in water; ~0.3 km/s in air.	P motion travels fastest in materials, so the P-wave is the first-arriving energy on a seismogram. Generally smaller and higher frequency than the S and Surface-waves. P waves in a liquid or gas are pressure waves, including sound waves.
S, Shear, Secondary, Transverse	Alternating transverse motions (perpendicular to the direction of propagation, and the ray path); commonly approximately polarized such that particle motion is in vertical or horizontal planes.	VS ~ 3 – 4 km/s in typical Earth's crust; >~ 4.5 km/s in Earth's mantle; ~ 2.5-3.0 km/s in (solid) inner core.	S-waves do not travel through fluids; so do not exist in Earth's outer core (inferred to be primarily liquid iron) or in air or water or molten rock (magma). S waves travel slower than P waves in a solid and, therefore, arrives after the P wave.
L, Love, Surface waves, Long waves	Transverse horizontal motion, perpendicular to the direction of propagation and generally parallel to the Earth's surface.	VL ~ 2.0 - 4.4 km/s in the Earth depending on frequency of the propagating wave, and therefore the depth of penetration of the waves. In general, the Love waves travel slightly faster than the Raleigh waves.	Love waves exist because of the Earth's surface. They are largest at the surface and decrease in amplitude with depth. Love waves are depressive, that is, the wave velocity is dependent on frequency, generally with low frequencies propagating at higher velocity. Depth of penetration of the Love waves is also dependent on frequency, with lower frequencies penetrating to greater depth.
R, Raleigh, Surface waves, Long waves, Ground roll	Motion is both in the direction of propagation and perpendicular (in a vertical plane), and "phased" so that the motion is generally elliptical – either protrude or retrograde.	VR ~ 2.0 - 4.2 km/s in the Earth depending on frequency of the propagating wave, and therefore the depth of penetration of the waves.	Raleigh waves are also depressive and the amplitudes generally decrease with depth in the Earth. Appearance and particle motion are similar to water waves. Depth of penetration of the Raleigh waves is also dependent on frequency, with lower frequencies penetrating to greater depth.

Table 4.1 Seismic Waves

Chapter 5

Seismic Wave Attenuation

5.1 Theoretical background

A fundamental feature associated with the propagation of stress wave in all real materials (earth) is the absorption of energy due to some mechanisms and the resulting change in the shape of transient waveforms. Laboratory work on the absorption in rocks showed the loss per cycle or wavelength to be essentially independent of frequency. Kolsky (1956) and Lomenitz (1957) gave linear description of the absorption that could account for the absorbed frequency independence. Despite this fact non –linear friction is commonly assumed to be the dominant attenuation mechanism, specially in crustal rocks (McDonald etal, 1958;Knoff, 1964;white, 1966;Gordon and Davis, 1968;Locker etal.,1977;Johnston and Toksoz,1977).. Different type of theory for attenuation has been advocated by Ricker (1953,19977) .He modeled the absorption is described by adding a single term to the wave equation of the type $A = A_o e^{iwt-ikx}$ gets an added damping term $e^{-\alpha x}$ which is the contribution for the intrinsic attenuation, so the solution becomes $A = A_o e^{iwt-ikx} e^{-\alpha x}$ this actually shows the frequency dependence of attenuation.. The effects of anelasticity on the wave propagation in rocks were modeled by Li etal (1976). He found that the change in elastic moduli implied by the attenuation over the frequency range covered by the seismic body waves and free oscillations are nearly independent of frequency. The linear description of attenuation in which Q is independent of frequency we call constant Q model (CQ) specified by two parameters i.e., phase velocity at an arbitrary reference frequency, and Q and the nearly constant Q model (NCQ) in which there is cutt off parameter i.e. when there is frequency range restriction. Fundamental discussion of individual anelastic damping mechanism indicate appreciable frequency dependent of Q of any particular mechanism (Jackson& Anderson 1970,Rev. Geophys.space phys.8.1), but measurements on rocks have nevertheless yielded virtually frequency- independent values of Q over wide frequency ranges (Knopoff 1964, rev.Geophys.2, 625). This frequency independence is the consequence of the superposition of very many anelastic responses (due to atomic displacements) in rocks, having a very wide range of individual relaxation times, and so smearing out the frequency dependence which would appear if they could be observed individually. Whatever the underlying reason, the simple fact the Q of a rock is not a significant function of frequency makes it a very useful parameter. We supposed up to now that the elasticity of material in which a wave propagating, with attenuation is linear, in the sense that strain is proportional to stress. But this is a self-contradiction. Attenuation by what mechanism is a stress- srain hysteresis. Indeed the energy lose per cycle is represented by the area of the loop traced out in its stress-strain diagram by a material which is cyclically strained. The fact of hysteresis therefore immediately calls the linearity assumption in to question and couples a reconsideration of what we mean by linearity in the present context. Firstly we may note that a material may have non- linear elasticity with out attenuation Such a material would be represented by a stress- strain graph which is curved but re-traces it self on an increasing and decreasing strains with out forming a loop. such a material would dissipate none of the energy of a propagating wave , but would deform the wave form of an initially sinusoidal harmonic wave, producing a wave form with a higher harmonic content and in appropriate conditions generating a shock-front wave form. Such a non-linearity most readily become apparent in waves of extremely large amplitude and so has little relevance to seismology, in which we are primarily concerned with elastic waves of very small strain amplitudes. So we are interested in linearity in which there is no harmonic generation by a wave but nevertheless attenuation occurs; waves may be supper imposed with out modifying one another and in particular the attenuation of several superimposed waves are independent. It is not clear what this means in terms of microscopic behavior of an elastic medium, but it serve as a definition of what we mean by *linearity of the attenuation mechanism*. As a summary and conclusion a very wide range of theories and empirical approaches suffices the attenuation of elastic waves, but the attenuation of elastic pulses are still the subject of conflicting ideas and it is beyond the scope of this studies.

The Quality factor Q is a dimensionless parameter that usually measures the internal friction (or anelstic attenuation) of waves and has the following different definitions. Before we come to different definitions of Q, let see the next concepts. As a wave is propagated through real materials, wave amplitude attenuates as a result of a variety of processes, which we can summarizes macroscopically ‘‘as internal friction.’’ as the following two examples show; i) the strains and stresses occurring within a propagating wave can lead to irreversible changes in the crystal defect structures of the medium. ii) work may also be done on a grain boundaries within the medium if adjacent material grains are not elastically bonded. Such media are said to be *anelastic*, since the configuration of material particle is to some extent dependent on the history of applied stress. The gross effect of internal friction is summarized by the dimension less quantity Q, defined in various ways for slightly an elastic media in the following ways.

If a volume of material is cycled in stress at a frequency w, then a dimensionless measure of the internal friction (or the anelasticity) is given by $\frac{1}{Q(\omega)} = -\frac{\Delta E}{2\pi E}$, where E is the peak strain energy stored in the volume and $-\Delta E$ is the energy lost in each cycle because of imperfections in the elasticity of the material. This definition is can be used only in special experiment under which we can find a material element with stress waves of unchanging amplitude and period. More commonly one observes either a) the temporal decay of amplitude in standing wave at fixed wave number or b) the special decay in propagating wave at a fixed frequency .In the case of either a) or b), for a medium with linear stress-strain relation, (assuming attenuation is a linear phenomena), wave amplitude A is proportional to $E^{\frac{1}{2}}$ (A –represent a maximum particle velocity or stress component in wave and assuming also that $Q >> 1$,so that successive peaks have almost the same strain energy.)

Hence $\frac{1}{Q(\omega)} = -\frac{1}{\pi}\frac{\Delta A}{a}$, from which we obtain the amplitude fluctuations due to attenuation in both (a) and (b). Thus in the case of (a) we can have $A(t) = A_o e^{-\frac{\omega t}{2Q}}$, at successive times A decreases a fraction $\frac{\pi}{Q}$, given that initially A=A_o.from this exponential decay values of A(t), we define the value of a *temporal Q* in which the attenuation of the earth’s free oscillations. For case (b) the derivation of the form $A = A(x)$ for distance x can be obtained just like it is derived in 5.4 and we can have $A(x) = A_o e^{-\frac{\omega x}{2CQ}}$, assuming that the direction of maximum attenuation is along x-axis which is also the direction of propagation and hence $\Delta A = (\ \frac{dA}{dx}\ )\lambda, where\ \ \lambda$ is the wave length given in terms of w and phase velocity by $\lambda = \frac{2\pi c}{\omega}$. From observations of exponentially decaying values of A(x), we use this equation to define the value of *spatial Q*. These values of spatial Q and temporal Q is explained in details (Acki Richard under attenuation of surface wave, p.294).

Attenuation and Scattering of seismic waves are important parameters to quantify and to physically characterize the earth medium & we ignore scattering here by considering homogenous earth. We first discuss physical mechanisms of attenuation and compile measurements of attenuation of seismic waves, and in the earth material: intrinsic absorption. As a model of attenuation, we consider the dominant attenuation mechanism as intrinsic attenuation by correcting the attenuation by geometrical spreading and derive a mathematical model for attenuation coefficient as a function of offset and frequency.

Let us see an overview of the attenuation of body waves and surface waves. The observed seismic-wave amplitudes usually decay exponentially with increasing travel distance after the correction for geometrical spreading, and decay rates are proportional to Q^{-1}, which characterizes the attenuation. For spherically outgoing S-waves of frequency f *in* a uniform velocity structure, the spectral amplitude at a travel distance r goes roughly as

$$A^{sDirect}\left(r, f\right) \approx \frac{e^{\frac{rfQ_S}{V_S}}}{r}$$

, where V_s is the S-wave Velocity.

Anderson and Hart (1978) proposed QS^{-1}~ 0.002, QP^{-1} ~0.0009, and ratio QP^{-1} /QS^{-1} ~0.5 for frequencies < 0.05 Hz for depths < 45 km. Yoshimoto *et al.* (1998) found rather strong attenuation in the shallow crust, QS^{-1} ~ 0.0034 $f^{0.12}$ and QP^{-1}~ 0.052 $f^{-0.66}$ for 25-102 Hz. Carpenter and Sanford (1985) reported *QP*-1/ QS^{-1} ~ 1.5 for 3-30 Hz in the upper crust of the central Rio Grande Rift. Observed characteristics may be summarized as follows: QS^{-1}

is of the order of 10^{-2} at 1 Hz and decreases to the order of 10^{-3} at 20 Hz. It seems reasonable to write the frequency dependence of attenuation in the form of a power law as $QS^{-1} \sim f^{-n}$ for frequencies higher than 1 Hz, where the power n ranges from 0.5 to 1.

5.2 Seismic Waves Attenuation in Rocks

- ### Experimental Observations-1

Attenuation of seismic waves in rocks has been measured on many rock types and under various conditions. Generally, attenuation in rocks is much higher than in minerals. For example, in calcite (mineral), $Q \approx 1900(Q^{-1} \approx 5x10^{-5})$, while in limestone, $Q \approx 200(Q^{-1} \approx 5x10^{-3})$). This means that attenuation is to a great deal controlled by defects, in homogeneities, structure and bonding properties of rocks. Attenuation coefficient α is frequency-dependent, it grows with increasing frequency. Attenuation decreases with increasing cementation and depth. Thus, velocity and attenuation show opposite behavior with respect to many actors, e.g. porosity, pressure, consolidation.

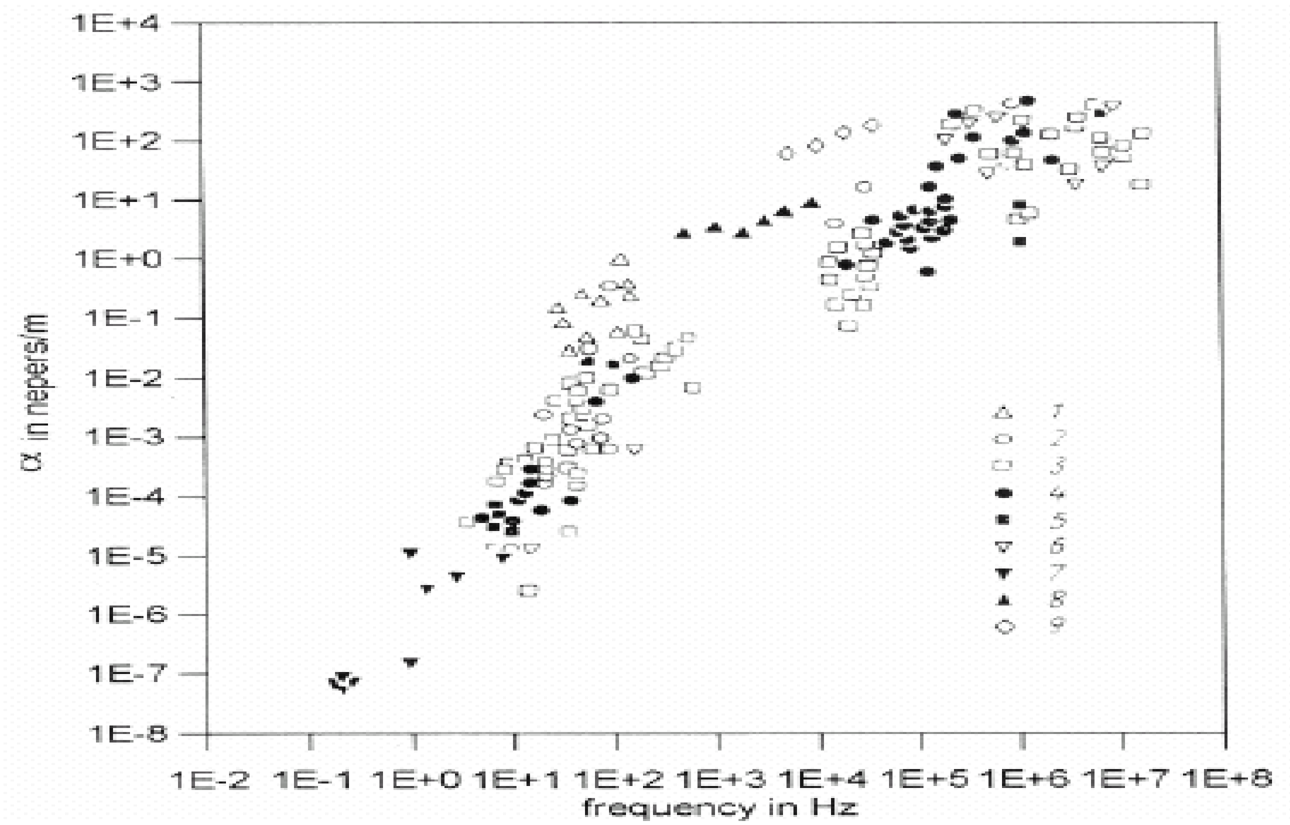


Fig 5.2-(a) Attenuation coefficient of various rock types as a function of frequency (drawn on the basis of a figure after Attewell and ramana,1966 and Militzer etal,1978)

1-sediments, unconsolidated 2-Sediments low consolidated

3-sediments consolidated 4-,5-eruptive rocks

6-metamorphic rocks 7- from deep seismic reflections

7- limestone (Germany) 8-sand, dry

- **Experimental Observations-2**

Rock types

Magma tic and metamorphic rocks

Consolidated sedimentary rocks

Shaly rocks

Unconsolidated rocks

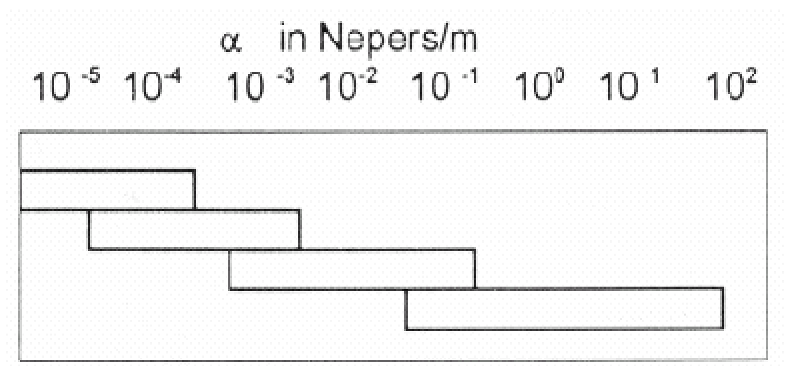


Fig 5.2-(b) Mean value range of attenuation coefficient for some rock types at a frequency of about 50…100HZ.

5.3Attenuation mechanisms

5.3.1 Overview

One of the least understood aspects of seismic wave propagation in the earth is the absorption /dissipation/ of wave energy in to heat. Increasing interest in seismic attenuation has created a need for better understanding of the physical processes involved in energy absorption. In particular we are interested in this study in the mechanisms that convert seismic energy into heat and not in any of the other processes, which affect seismic amplitudes (O'Doherty and Anstey, 1971;sheriff, 1975).

A seismic wave loses energy as it propagates through the earth. The physical processes which are responsible for seismic energy absorption (the primary mechanisms) are spherical divergence, scattering, and intrinsic attenuation. Spherical divergence (5.3.2B) is independent of frequency and a correction is commonly applied to seismic data. Scattering results from multiples that lag the Primary seismic energy and tend to cause some cancellation of primary energy. This attenuation is frequency dependent, which is dominated by intrinsic attenuation and is difficult to separate from it. Intrinsic attenuation is energy loss due to friction. Grain to grain friction in sedimentary rock causes some attenuation.

5.3.2 Types of attenuation mechanisms

A.Anelastic attenuation (Intrinsic attenuation).

Absorption is the energy loss caused by the imperfection or defect of the material, in the form of energy conversion from mechanic to thermal. This loss can be accounted for by using the absorption coefficient α in the form as

$A = A_o e^{-\alpha x}$, Now we introduce the concept of the quality factor Q, which is defined as the ratio of the total elastic energy and the energy lost in one

Cycle, i.e., $Q = \frac{2\pi E}{\Delta E}$, Q can be thought as after how many cycles of vibration the elastic energy can be dissipated, apparently larger Q means many cycles to dissipate the energy so that the material tends to be more pure elastic. In contrast, if only after very few cycles the energy is gone, the material is far more from elastic.

Energy is proportional to the square of the amplitude, i.e., we have $E = \frac{1}{2}\rho\omega^2 A^2$; taken the reference point at point 1, and so we have

$$\frac{2\pi}{Q} = \frac{\Delta E}{E} \approx \frac{A_1^{\ 2} - A_2^{\ 2}}{A_1^{\ 2}} = \frac{(A_1 + A_2)(A_1 - A_2)}{A_1^2} \approx \frac{2A_1(A_1 - A_2)}{A_1^{\ 2}}$$

$\approx \frac{2(A_1 - A_2)}{A_1} = 2\left(1 - \frac{A_2}{A_1}\right) \approx 2\ln\left(\frac{A_1}{A_2}\right) = 2\Delta$, Taylor expansion has been applied in the last step, since $\ln x = 1 - \frac{1}{x} + ...(x > 0)$, On the other hand, from the original definition of the absorption coefficient we have the amplitudes at 2 points with only one cycle apart (one wavelength in space) can be expressed as

$$A_1 = A_o e^{-\alpha x_1}$$

$$A_2 = A_o e^{-\alpha(x_1 + \lambda)}$$

then

$$\frac{A_1}{A_2} = \frac{A_0 e^{-\alpha x_1}}{A_0 e^{-\alpha(x_1+\lambda)}} = e^{-\alpha x_1}.e^{\alpha(x_1+\lambda)} = e^{-\alpha x_1 + \alpha x_1 + \alpha\lambda = e^{\alpha\lambda}}$$

And $\ln\left(\frac{A_1}{A_2}\right) = \ln\left(e^{\alpha\lambda}\right)$, so we got that $\Delta = \alpha\lambda$ and $Q = \frac{2\pi E}{\Delta E} = \frac{2\pi}{2\Delta} = \frac{\pi}{\alpha\lambda}$

(detailed derivation for this –5.4). This is the relation between the quality factor Q and the absorption coefficient α. The absorption is the mechanism responsible for complete dead-of of seismic vibrations.

The general subject of **wave** attenuation by internal friction (called intrinsic attenuation) is very large, as can be seen from identifying three deferent aspects:

1) Studies are made up of the fundamental (microscopic) processes that cause attenuation. The effect of variety of crystal defects, grain-boundary processes, and some thermo elastic processes have been reviewed by Mason (1958), Jackson and Anderson (1970), and Nowick and berry(1972).
2) The frequency dependence of Q is studied as a macroscopic phenomenon in a variety of materials. From observations of the frequency dependence of Q in seismic waves (Archambeau etal.,1969, Solomon,1972,1973), one may constrain the possibilities for earth composition.
3) Many authors have developed macroscopic stress-strain relations to replace Hookes law, and hence (with $\rho \ddot{u}_i = \tau_{ij,j}$) have obtained equations of motion for materials with some particular $Q = Q(\omega)$.

The anelastic attenuation of seismic waves in the top few kms of the earth's crust is believed to be partly due to fluid within or between cracks. Measurement of 1/Q, which is proportional to the fractional loss of energy per cycle of sinusoidal wave, may there fore provide information on both the concentration of fractures within a rock body and the manner in which these fractures are interconnected. In this case the study of attenuation in fracture is very important i.e. now days naturally fractured reservoirs have attracted interests of exploration and production geophysicist. Because in many instants, natural fractures control the permeability of the reservoir. The formation of natural fracture arises from geological process of physical deformation or digenesis, which induces strains exceeding the maximum strength of the rock. Main processes responsible for the formation of fractures are listed by Landes (1959) as; the structural deformation arises for folding, and faulting, deep and rapid erosions of overburden that allows uplifts and expansion of depth pressure rocks, volume reduction caused by dewatering of shales, cooling of igneous rocks. It is these all fracture process that affects the amplitude and pulse shape of the incident seismic waves, which of course give us an indication, that certain earth materials containing hydrocarbon reservoirs. Q measurements within the uppermost layers of the earth's crust are spares, and usually accomplished by the *spectral ratio methods*, however this method suffers form several deficiencies, including frequency dependence losses due to energy partition at the boundaries and multiplicity of arrivals within a wave train of sufficient length for spectral analysis.

Ricker (1953) pioneered an extenuative and simpler technique of *wave let broadening* measurements to determine attenuation. Gladwinand Stacy(1974,Phys.earth and planetary introduction,p-332-336) have suggested that the use of seismic pulse rise times that are equivalent to Ricker's measuerements,may have advantages over spectral ratio measurements in which this study based on, because of very short wave length of signal required; the problems caused by multiple arrivals and by reflection and transmission losses can thus be partly over come.

There are several review papers that discuss proposed mechanisms for intrinsic absorption that lead to frequency-independent QP^{-1} and QS^{-1} (Knopoff, 1964; Jackson and Anderson, 1970; Mavko et al., 1979; Dziewonski, 1979). For seismic waves to remain causal in the presence of attenuation, the relationship between frequency-dependent Q^{-1} and velocity dispersion was discussed by Liu et al. (1976). Many proposed mechanisms are based on the observation that crustal rocks have microscopic cracks (fractures) and pores, which may contain fluids. These features have dimensions much smaller than the wavelengths of regional seismic phases. Walsh (1966) proposed frictional sliding on dry surfaces of thin cracks as an attenuation mechanism. Nur (1971) proposed viscous dissipation in a zone of partially molten rock to explain the low velocity/high attenuation zone beneath the lithosphere. Even though the addition of water reduces the melting temperature of rocks, it is unlikely that melted rock exists in most regions of the lithosphere. Mavko and Nur (1979) examined the effect of partial saturation of cracks on absorption: fluid movement within cracks is enhanced by the presence of gas bubbles. O'Connell and Budiansky (1977) proposed a model in which fluid moves between closely spaced adjacent cracks. Tittmann et al. (1980) measured an increase of QS-1 with increasing content of volatile in dry rocks. They found that the rapid increase was due to an interaction between absorbed water films on the solid surface by thermally activated motions. Thermally activated processes at grain boundaries have been proposed as an absorption mechanism for the upper mantle

(Anderson and Hart, 1978; Lundquist and Cormier, 1980). Spatial temperature differences induced by adiabatic compression during wave propagation will be reduced by thermal diffusion (Zener, 1948; Savage, 1966a), which removes vibrational energy from the wave field. Grain-sized heterogeneities in a rock increase the amount of predicted absorption by this mechanism, which is called thermoplastic effect. Savage (1966b) investigated thermoelasticity caused by stress concentrations induced by the presence of cracks. Most of the mechanisms discussed above can predict QS^{-1} having values in the range of 10-3; however, the importance of various mechanisms varies with depth, temperature, fracture content, fracture aspect ratios, pressure, and the presence of fluids. Aki (1980) discussed a relation between physical dimensions and the observed and partially conjectured frequency-dependence of QS -1 having a peak on the order of 0.01 around 0.5 Hz. He preferred thermo elasticity as the most viable model at lithospheric temperatures since the required scales for rock grains and cracks along with the amount of attenuation are in closest agreement with observations.

For the intrinsic attenuation (absorption), as the wave motion passes through real earth medium (an elasticity), the wave energy is transformed in to other forms. Or the elastic energy associated with the wave motion is gradually absorbed by the medium, reappearing ultimately in the form of heat. This process is called *attenuation (absorption).* Toksoz and Johnston (1981) summarize much of literature regarding absorption. The measurement of absorption is very difficult, mainly because it is not easy to isolate absorption from other effect making up attenuation.

B.Geometrical spreading, Scattering and energy partitioning at an interface

Geometrical spreading:-It is observed that seismic waves decrease in amplitude due to spherical spreading and due to mechanical or other loss mechanisms in the rock units that the wave passes through. From the concept of energy intensity we have discussed under the sub- topic 4.4 and using equs. (4.4.2), we can discuss the concept of geometrical spreading which is taken us correction for the energy distribution. The Seismic energy induced at a particular point will propagate in every direction with spherical wave front. The total energy at the instant of seismic energy release will be distributed over all spherical wave front, which is getting larger and large as the wave propagates out ward.

Let's imagine two wave fronts, which make 2 spherical shells whose centers is the coincided location of the source as Fig 5.3 below shows.

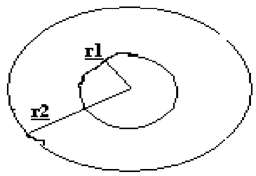


Fig 5.3 spherical wave front having two shells of radii r_1 & r_2

The radius to the outer shell is r_2, which is greater than that of the radius of the inner shell r_1. Thus, the surface areas of the outer and inner shells are $4\pi r_2^{\,2}$ and $4\pi r_1^{\,2}$, respectively. By energy conservation, the total energy flowing through the outer shell and the inner shell at a given time should be keep the same so that we have

$I_2 S_2 = I_1 S_1$, Where E and S are energy density and Area respectively.

$$I_2 4\pi r_2^{\,2} = I_1 4\pi r_1^{\,2}$$

$$\Rightarrow I_2 = \left(\frac{r_1}{r_2} \right) I_1$$

$$\Rightarrow \frac{1}{2}\rho\omega^2 u_2^{\,2} = \left(\frac{r_1}{r_2} \right) \rho\omega_2 u_1^{\,2}$$

$$\Rightarrow u_2 = \left(\frac{r_1}{r_2} \right) u_1$$

This state that the amplitude is decaying against 1/r for the waves generated by a point source, since the shape of the wave front is spherical, this is generally referred as the ***geometric spreading*** for spherical waves and hence we make correction for decrease in amplitude due to this spherical spreading to extract intrinsic attenuation only.

In a similar argument for an infinitely long line source, the shape of the wave front is a cylinder; this is called the cylindrical wave, we can have $u_2 = \sqrt{\frac{r_o}{r}} u_1$

This state that the amplitude is decaying against $\frac{1}{\sqrt{r}}$ for waves generated by a line source, since the shape of the wave front is cylindrical, this is generally referred as the geometric spreading for cylindrical waves.

If the shape of the wave front is planar, this wave is called the plane wave; there is no amplitude decay for plane wave. In summary we can view the geometric spreading of wave energy as:

Point source: spherical wave $\frac{1}{r}$ decay;

Line source: cylindrical wave $\frac{1}{\sqrt{r}}$ decay;

Plane source: plane wave no decay;

The geometric spreading alone cannot lead to the complete dead off of seismic wave energy. The ultimate dead off of the kinetic energy of seismic waves is due to the energy absorption caused by the imperfection of the earth materials, i.e., the elastic energy has been completely transferred to earth mantle (discussed above under A part).

Scattering: - Another important attenuation mechanism is the reduction in amplitude of a wave by the scattering of its energy by diffraction by objects whose dimensions are on the order of the wavelength. If a is an average linear dimension of velocity in homogeneities then the attenuation coefficient is given approximately by: $\alpha \approx \frac{a^3}{\lambda^4}$. So attenuation increases rapidly with decreasing wavelength. Consider attenuation is an unconsolidated medium with a velocity of 250 m/sec and a frequency of 1000 Hz. Then, λ = 0.25 m, and α = a3 .256. The wave would fall to 1/e of its initial amplitude when a = 157 m.

It might be reasonable to expect in homogeneities with a characteristic dimension on the order of 15 cm in the overburden so it is likely that the very high attenuation observed in near surface unconsolidated sediments is due to scattering. Scattering due to heterogeneities distributed in the earth also causes a decrease in amplitude with travel distance (Aki, 1980), where the characteristic frequency is determined by a characteristic spatial scale, such as the correlation length of random media or the crack length. Eventhogh the study of scattering using the scalar wave equation in inhomogeneous media is beyond the scope of this study, it is possible to study in details mathematically.

- **Energy partitioning at an interface:** After we discuss geometric spreading and absorption, which occurs for even uniform medium, we need discuss energy partitioning at interfaces caused by heterogeneous medium.

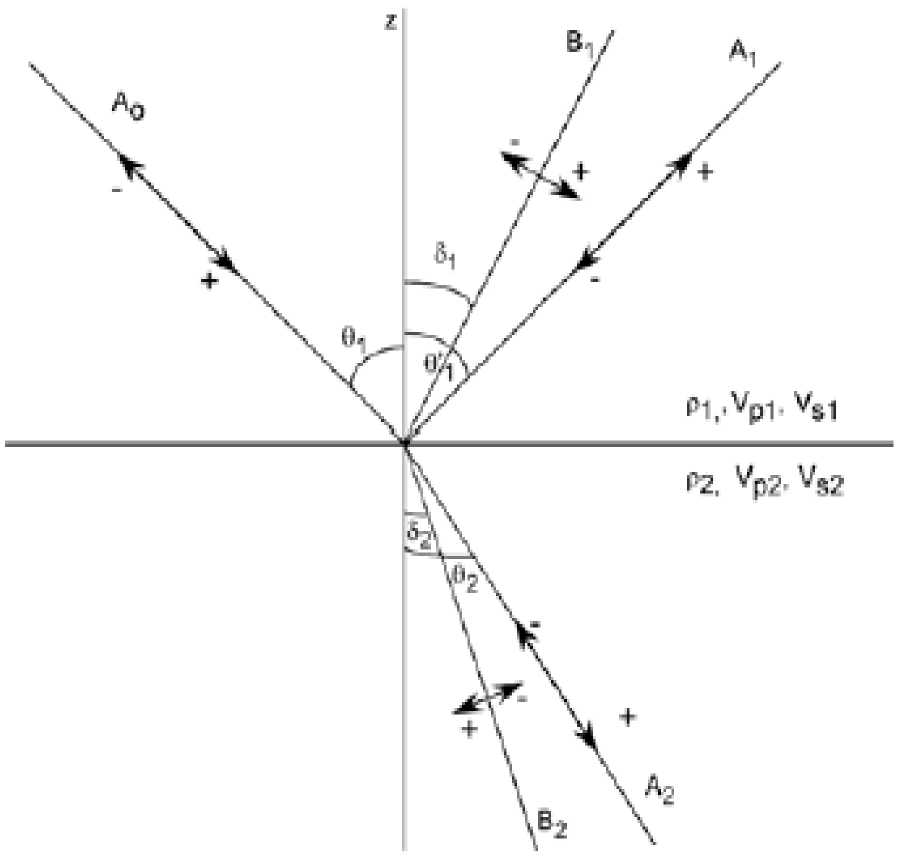


Figure 5.4 a plane p-wave impinging at the interface with $\rho_2 v_2 > \rho_1 v_1$.

When the medium becomes inhomogeneous, there will be interfaces between media with different properties. At an interface the incident seismic wave will be reflected, and refracted, so that the incident seismic kinetic energy will be chucked into fractions. The physics controls the seismic reflection and refraction is the Snell's law expressed as

$\frac{\sin\theta_1}{v_1} = \frac{\sin\theta_2}{v_2}$, the meanings of the symbols are referred to Figure 5.2. Only consider geometry, and kinetics – not consider the causes of the deformation or motion.

The premise of this approximation is that the frequency of the waveforms is assumed to be infinitely high, or the wavelength is very, very short compared with the features it studies. No diffraction phenomenon is considered in the treatment, only reflections and refractions.

Reflection Coefficient: - The process of wave reflection may be defined as the return of all or part of a sound beam when it encounters the boundary between two media. The most important rule of reflection is that the angle of incidence is equal to the angle of reflection. Where both these angles are measured relative to an imaginary line, which is normal to the boundary. Reflection is often quantified in term of the reflection coefficient 'R'. R is defined simply as the ratio of the reflected and incident wave amplitudes.

$$R = \frac{A_r}{A_{i1}}$$

Where 'A$_i$' and 'A$_r$' are the incident and reflected wave amplitudes respectively. The value of the reflection coefficient relates to the magnitude of reflection from the interface between two media with different physical properties. The acoustic impedance (Z) is simply the product of the density (ρ) of the medium and the speed (v) of the wave i.e. $Z = \rho v$.

Acoustic impedance is measured in Rayles (1 Rayle = 1 m/s.kg/m3 = 1 kg/m2/s).

The full expression for sound reflection coefficient is

$$R = \frac{\left(Z_2/Z_1 \right) - \sqrt{1-(n-1)\tan^2\alpha_i}}{\left(Z_2/Z_1 \right) + \sqrt{1-(n-1)\tan^2\alpha_i}}$$

Where $n = (\frac{C_2}{C_1})^2$ -is refractive index of the medium and α_i is the angle of incidence of the wave ray.

Notice that since energy is always conserved the remaining energy that is not absorbed must be either dissipated (e.g. in the form of heat) or transmitted into the second medium whereby it will undergo refraction if the velocities in the two layers differ. If dissipation is negligible (a loss-less medium) the amplitude of the transmitted wave will be (1-R).

NORMAL INCIDENCE: - If the waves are normally incident to the boundary, the reflection equation can be simplified to: $R = \frac{Z_2 - Z_1}{Z_2 + Z_1}$, for $\alpha_i = 0$

Thus, acoustic reflection is a simple function of the impedance of the two media. If the two media have the same impedance there will be no reflection. Since the impedance is the product of velocity and density. It is possible for example to have two media with different densities or sound speed but the same acoustic impedance. Acoustic reflection coefficients have values that range between -1 and +1. From this range we can identify 4 different types of reflection:

1) $Z_2 >> Z_1, R \Rightarrow 1$ (Rigid boundary), i.e., most of the acoustic energy will be reflected with out a change in phase.

2) $Z_2 << Z_1, R \Rightarrow -1$ (Soft or pressure lease boundary), i.e., most of the acoustic energy is reflected with a 180 degree phase change.

3) $Z_2 = Z_1, R = 0$, (No Reflection)

4) Similar acoustic impedance, -1 << R << 1, some phase change.

5.4 Surface wave attenuation estimation for the region

The ability to estimate correct attenuation coefficients impacts the ability to reliably get a damping (amplitude) ratio profile. i.e. attenuation estimates at individual frequencies (offsets) affect the damping ratio profile at depths proportional to that frequency-specific wavelength and attenuation coefficients stem from a convolution of the material damping

ratios from the surface to a depth proportional to the wavelength, weighted by an eigenfunction(Complex – valued scaling function), and therefore, attenuation estimates at a specific depth depend on the material damping properties from the surface to that depth. Consequently, misestimating the attenuation will impact the damping ratio profile. Most attenuation estimates have suffered from the following major
limitations:

1.) Incorrect physical and geometric spreading model,

2.) Inability to remove noise optimally,

The model incompatibility represents the greatest impediment to improving attenuation estimates and has produced an unnecessarily complicated attenuation model. The inability to remove noise optimally introduces varying noise power into the attenuation estimate. Noise removal or minimization is very important in any geophysical methods as far as it increases the resolution of the required result .For example the following discusses the effects of noise inclusion. If the background seismic noise field exhibits stationary statistics at each frequency of interest, inclusion of the noise in the estimation process leads to conservative (i.e. too low) material (intrinsic) attenuation estimates. Consider a simple plane wave field example, consisting of a single wave with amplitude = 10 at position x_1 = 0 m, and amplitude = 8 at position x_2= 10 m. Assume a stationary background noise field is also present in the measurements, where the noise measured at the two locations is independent, identically distributed white Gaussian noise, and the spectral amplitude of the noise = 3 at the frequency of interest. Viewing attenuation as the relative decline of energy between two points, referenced by the original amplitude, the attenuation coefficients for the *include noise* and *remove noise* cases are the following:

Including stationary Noise

$$\frac{A_1(Signal + Noise) - A_2(Signal + Noise)}{A_1(signal + Noisae)} = \frac{(10+3)-(8+3)}{10+3} = \frac{2}{13} = 0.15$$

$$Attenuation_{(Signal+Noise)} = \frac{0.15}{10} = 0.015\left(1/m\right)$$

Remove Stationary Noise:

$$\frac{A_1(Signal) - A_2(Signal)}{A_1(Signal)} = \frac{10-8}{10} = \frac{2}{10} = 0.2$$

$$Attenuation_{(Signal)} = 0.2/10 = 0.02\left(\frac{1}{m} \right)$$

In this case, a higher attenuation estimate results from removing noise. The nonstationary noise power case introduces the possibility of increasing energy as a function of distance. The simple example frames the discussion regarding the effects of trying to correct noisy measurements. The traditional methods of noise correction increase the estimate uncertainty under certain circumstances, which has undesirable consequences on the inverted damping ratio soil profile. Advanced noise removal, estimation, and cancellation techniques exist in advanced digital signal processing. So in this study noise coexisting associated with the required trace is filtered by band pass filter using appropriate time window Removal.

Except for nonlinear behavior near the explosion (earthquake) foci, seismic strains are small and seismic oscillations take place in the linear domain of elasticity. Attenuation of harmonic signal is therefore exponential, and the magnitude of the attenuation is describable by the exponential rates of decay.

a) The physics of seismic wave attenuation

The same general equation governs wave propagation in cylindrical, spherical and plane wave fields, which means all motions can be modeled by identical parameters. The parameters include the amplitude, wave numbers, and a complex-scaling model. The general equation will be introduced as: A single surface wave propagating with a single frequency and single wavelength in a dissipative medium can be described with the following equation:

$$u(K, x, \omega, t) = A_o(K, \omega) e^{-\alpha(K,\omega)x} e^{j\omega t} R(K, x) \text{----------------------------(5.5.1)}$$

where $u(K, x, \omega, t)$ = the measured displacement (amplitude) at spectral components **k** and ω and at vector position **x** and time t, A_o = initial amplitude of the propagating wave, α = the material (intrinsic) attenuation coefficient, $e^{j\omega t}$ = harmonic time dependence, and), $R(K,x)$ = a complex-valued scaling function, which includes phase and geometric spreading information. The important features of the model include the following:

1.) Material (intrinsic) attenuation α - The attenuation of wave energy in a horizontally (vertically) heterogeneous soil profile is a material parameter that is a function only of frequency and wave number,

2.) Governing geometric model R (R for Raleigh) - Depending on the geometry of the problem, i.e. either plane wave, spherical or cylindrical, R equals a complex exponential or Hankel function solution to the plane and spherical or cylindrical wave equation, respectively, evaluated at the argument (**k**,**x**),

3.) Amplitude A_0 - The wave amplitude at some reference position x_0 and reference time t_0.

If multiple waves are present in the wave field, superposition sums yield the model of the motion. The dependence of the geometric model R and material attenuation on frequency and wave number displays an intimate relationship with the dispersion relation. To see the intrinsic attenuation with general model, use the general model and two reference points along a linear axis, consider a single wave propagating with single amplitude A0 at a single frequency w_0, single wave number k_0, and single attenuation coefficient $\alpha 0$. Since the motion is at a single temporal frequency, and using a fixed reference time, the dependency on the temporal motion characteristics can be suppressed, allowing focused attention on the spatial properties of the model. If material attenuation equals zero, the displacements u measured at the two positions x_1 and x_2 along the linear axis, where x_1 is closer to the signal source, equal

$$u(k_o, x_1) = A_o(K_o)R(K_o, x_1),$$
$$u(K_o, x_2) = A_o(K_o)R(K_o, x_2), \qquad \text{------(5.5.2)}$$

where the motion u is complex-valued, the real-valued part corresponding to the actual motion, and the function R yields a complex-valued scaling of the amplitude A0, i.e. a change in the magnitude and phase of the motion. The complex-valued scaling factor between the two positions x_1 and x_2 is given by

$$\frac{u(K_o,x_2)}{u(K_o,x_1)}=\frac{A_o(K_o)R(K_o,x_2)}{A_o(K_o)R(K_o)R(K_o,x_2)}=\frac{R(K_o,x_2)}{R(K_o,x_1)}$$.---------------------------(5.5.3)

For a plane wave, the amplitude A_0 would not decay, but the phase of the motion would change as the wave propagated from point 1 to 2. For a cylindrical wave and spherical wave, the function R determines the phase change and the geometric spreading of the energy, and therefore, the amplitude A_0 would decrease as the wave propagates from position 1 to 2. Introducing material attenuation into the model, the displacements at x_1 and x_2 equal

$$u(k_o,x_1)=A_{o(k_o)}e^{-\alpha(k_o)x_1}R(k_o,x_1)$$

--(5.5.4)

$$u(k_o,x_2)=A_{o(k_o)}e^{-\alpha(k_o)x_2}R(k_o,x_2)$$

. The complex-valued scaling of motion between position x_1 and x_2 now equals

$$\frac{u(k_o,x_2)}{u(k_o,x_1)}=\frac{A_{o(k_o)}e^{-\alpha(k_o)x_2}R(k_o,x_2)}{A(k_o)e^{-\alpha(k_o)x_1}R(k_o,x_1)}=e^{-[\alpha(k_o)(x_2-x_1)]}\frac{R(k_o,x_2)}{R(k_o,x_1)}$$ ---(5.5.5)

Notice the original wave amplitude A_0 always cancels out, indicating the relative nature of material attenuation. Considering Equation 5.5.5 as a filter or linear system places the attenuation problem into a more common perspective. The output u (x_2) equals the input u(x1) scaled by a complex-valued filter coefficient defined entirely by the wave number and attenuation coefficient, and therefore, the following equation represents the site transfer function $$\frac{u(x_2)}{u(x_1)}=e^{-\alpha(x_2-x_1)}\frac{R(k_o,x_2)}{R(k_o,x_1)}$$ ---------------------------------(5.5.6).

The function $\frac{R(k_o,x_1)}{R(k_o,x_2)}$ equals the phase change and geometric spreading due to the site-specific wavenumbers, and therefore, is a function of the same parameters as the dispersion curve, i.e. wavenumber and frequency. In plane wave analysis, geometric spreading will not be a factor, yielding a $\frac{R(k_o,x_1)}{R(k_o,x_2)}$=1 for all frequencies and wave numbers. And in this study we considered spherical wave fields and far-field so that

$\frac{R(k_o, x_1)}{R(k_o, x_2)} \Rightarrow \frac{x_1}{x}$, where x_1 is the offset distance from the source for the first geophone is taken for the correction of geometrical spreading and equation 5.5.6 becomes $\frac{u(x_2)}{u(x_1)} = e^{-\alpha(x_2 - x_1)} \frac{x_o}{x}$ as it can be further seen from mathematical model of the attenuation for the region below using phenomenological model.

b) Mathematical modeling of seismic wave attenuation for the region under study:
For purely elastic earth, geometric spreading, and the reflection and transmission of energy at boundaries control the amplitude of seismic pulse. But the real earth is not perfectly elastic, and propagating waves attenuate with time due to various energy lose mechanisms as discussed above. The successive conversion of potential energy (particle position) to kinetic energy (particle velocity) as a wave propagates is not perfectly reversible and other work is done, such as movements along mineral dislocation or shear heating at grain boundaries, that taps the wave energy. This process usually described as internal friction, and we "model" the internal friction effects (intrinsic attenuation) with phenomenological descriptions because the microscopic process are complex.

The simplest description of attenuation can be developed for an oscillating mass on a spring. As figure 5.4 below where mass M attached to a spring with spring constant K.

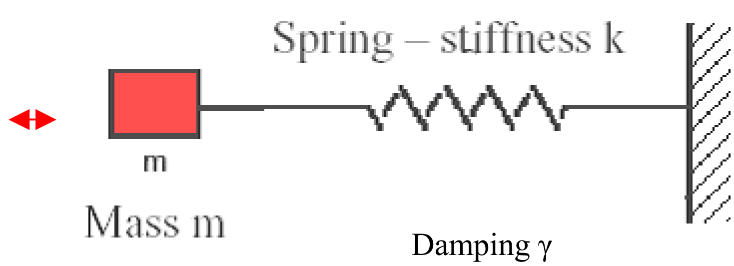


Fig 5.5 An oscillating mass attached to a spring

Let us consider first frictionless (just like perfectly elastic earth). The equation of motion for this system is as follows:

$m\frac{d^2x}{dt^2} + kx = 0$ --(5.5.7),

the general solution to this is, harmonic oscillation:

X=A $e^{iw_o t} + Be^{-iw_o t}, w_o = \sqrt{\frac{k}{m}}$, once the motion starts, it will continue forever, oscillating at the natural frequency of the system w_o . But this is not the case in the real earth, and then we introduce intrinsic attenuation by adding a damping force, such as friction between the moving mass and underlying surface. In this case there is an added force, proportional to the velocity of the mass as: $m\frac{d^2x}{dt^2} + \gamma\frac{dx}{dt} + kx = 0$ ---------- -----(5.5.8)

$$\frac{d^2x}{dt^2} + \frac{\gamma}{m}\frac{dx}{dt} + \frac{k}{m}x = 0$$

$$\omega_o = \left(\frac{k}{m}\right)^{\frac{1}{2}} \cdot \gamma$$

$$\frac{d^2x}{dt^2} + \varepsilon\omega_o\frac{dx}{dt} + \omega^2{}_o x = 0$$ ---(5.5.9),

Where $\varepsilon = \frac{\gamma}{m\omega_o}$ =called Coefficient of friction and

The solution of this equation has the form:

$x(t) = A_o e^{-\varepsilon\omega_o t}\sin\left(\omega_o t\sqrt{1-\omega^2}\right)$---(5.5.10) ,

where $A_o e^{-\varepsilon\omega_o t} = A(\varepsilon)$ is harmonic oscillator that decays exponentially with time. Here in our case we can express ε in the form of quality factor, Q: $\varepsilon = \frac{1}{2Q}$, then

$A(t) = A_o e^{-\frac{\omega_o t}{2Q}}$ ---(5.5.11),

where Q - is parameter that defines intrinsic attenuation, which is called quality factor, it is the fractional loss of energy per cycle of oscillations, in another words

$\frac{1}{Q} = -\frac{\Delta E}{2\pi E}$ (Knoff,1964)---(5.5.12).

For our purpose i.e. measurement of attenuation from amplitude, observation in terms of amplitude than energy is preferred, then we have $E \propto A^2$, using this and equs (5.5.12) and differentiating, we have $dE \propto 2AdA$, hence $\frac{dE}{E} = \frac{2dA}{A}$, integrating there fore; $A = A_o e^{\frac{-\pi}{Q}\cdot\frac{t}{\tau}}$, in terms of travel x, this can be modified using the relationship: $x = ct$, $c = \frac{\lambda}{\tau}$, where c, τ and λ are the velocity, period and wave length of the wave respectively. Substituting this, we have $A = A_o e^{-\frac{\pi}{Q}\cdot\frac{x}{\lambda}}$ ---------------------------------(5.5.13)

$A = A_o e^{-\alpha x}$ ---(5.5.14),

Where $\alpha = \frac{\pi}{Q\lambda} = \frac{\pi f}{Qc}$, ---(5.5.15)

is attenuation coefficient. For two different positions x_0 and x from the source, we have amplitudes $A_o(x_o) \Rightarrow A_1$ from reference geophone (the first geophone) because here we considered this geophone records the initial elastic wave amplitude A_o (as it is verified below) for the corresponding shot points and at any position x from the source is A. Therefore equs. (5.5.14) can be written for different positions from the source as;

$A_1 = A_o e^{-\alpha x_1}$, $A_2 = A_o e^{-\alpha x_2}$, $A_3 = A_o e^{-\alpha x_3}$ and so on. Taking the amplitude ratio of any amplitude at different position to the first geophone A_1, example for the second geophone, we have

$$\frac{A_2}{A_1} = \frac{A_o e^{-\alpha x_2}}{A_o e^{-\alpha x_1}} = e^{-\alpha(x_2 - x_1)}$$

$\Rightarrow A_2 = A_1 e^{-\alpha(x_2 - x_1)}$, Here A_1 is now considered A_o and rewriting this equs. In general for ' n' number of geophone from the source, we have $\Rightarrow A_n = A_1 e^{-\alpha(x_2 - x_1)}$ and Combining the spherical divergence (spherical spreading) term $\frac{x_o}{x}$ (to see the intrinsic attenuation effect only. Assuming other attenuation mechanism like scattering and diffraction are negligible), to this equation, we have

$$\Rightarrow A_n = A_1 \frac{x_o}{x} e^{-\alpha(x_2 - x_1)}$$

$\Rightarrow \alpha = \frac{1}{x_2 - x_1} \ln\left(\frac{A_1 x_o}{A_n x}\right)$, The unit of attenuation coefficient α in this case is $\frac{Nepers}{meter}$ 0r when we take $20\log\left(\frac{A_1}{A_n}\right)$ in terms of natural logarism, α is measured in $\frac{Decible}{meter}$. If amplitudes of the two consecutive cycles are $A(x)$ and $A(x+\lambda)$, where λ is wavelength, and using frequency f and velocity of the wave propagation v, one can derive the logarithmic decrement

$\delta = \ln\left[\frac{A(x)}{A(x+\lambda)}\right] = \alpha\lambda = \alpha\left(\frac{v}{f}\right)$ --(5.5.16).

In many cases, attenuation is expressed by dimensionless Knopoff quality factor Q (as discussed above 5.5.12) and dissipation factor Q^{-1}, $Q^{-1} = \left(\frac{\alpha v}{\pi f}\right)$ ----------------------- (5.5.17)

Chapter 6 Data processing, analysis, and attenuation measurement

6.1 Data processing and analysis.

In this study seismic wave attenuation measurement is undertaken with a particular concern on surface wave recordings by GURALP 6TD. The waves are then utilized for this particular task and processed according to the needs of data processing procedure of the study in the following way.

1) Surface wave recordings from the shots: - Data from (GohaTsion shot - representing the Western Ethiopian plateau, Chefe Donsa shot - representing the main Ethiopian Rift and Kula shot- representing Eastern Ethiopian plateau) are independently stacked (Table 6.1,6.2, and 6.3) and traces are examined one by one.
2) A single surface wave trace is truncated and filtered and then the trace is further processed so that accurate amplitude of the wave is read for each geophone positions from the corresponding shots.
3) Amplitude spectrum and Group velocity are read for some frequency ranges at specific offset for each shot points.
4) The reference amplitude A_1 for the first geophone is taken as the initial elastic wave\` amplitude A_o based on the mathematical derivation illustrated under 5.5b.
5) The trace, which passed through all the above process, is further be changed from qualitative form (traces or graphical form) to quantitative (number) describing parameters to be used for the tasks required. The following tables show from which the above data processing is undertaken (Table 6.1,6.2 and 6.3).

No	Geophone No	Location		Elevation	Offset (Km)
		Easting	Nor thing		
1			1103578.8	2530	0.58
2			1100870.1	2562	2.93
3	1011	427754.0	1095391.9	2592	10.21
4	1014	429875.5	1093009.0	2559	13.4
5	1018	432129.0	1089261.9	2588	17.74
6	1023	434605.0	1084830.6	2560	22.79
7	1026	435914.0	1083230.1	2572	24.88
8	1030	440547.5	1080632.6	2593	29.74
9	1037	446617.9	1076276.4	2673	37.01
10	1042	448338.6	10709337.3	2565	42.18
11	1046	44994.9	1067397.3	2486	44.18
12	1054	4544o8.2	1062671.4	2669	52.42
13	1062	454080.7	1055340.9	2450	58.14
14	1081	462918.6	138778.5	2426	76.8
15	1085	4633794.0	1035198.7	2484	80.3
16	1089	466194.8	1032144.3	2602	84.2
17	1094	470172.1	1028505.4	2578	89.4
18	1097	472625.6	1028101.0	2614	91.1
19	1101	475965.2	1025951.0	2717	94.8
20	1110	483573.0	1022862.9	3262	101.8
21	1120	491944.1	1014703.4	2645	113.4

Table 6.1 Goha Tsion shot shot-western Ethiopian plateau (Geophone No, Elevation, Location and Offset).

No	Location				Off set (Km)
	Geophone No	Easting	Northing	Elevation	
1	1157	514208.8	991094.8	2348	1.79
2	1163	516065.3	986922.4	2309	6.35
3	1371	622605.1	851696.6	2490	14.9
4	1179	526760.4	977479.6	2415	20.3
5	1182	527905.9	974571.6	1400	23.2
6	1204	543958.1	961313.9	1536	43.9
7	1219	549953.6	948247.8	1870	57.6
8	1226	552660.2	942597.5	1817	63.7
9	1231	550843.5	938088.5	1479	66.2
10	1235	552982.9	936423.1	1453	68.8
11	1242	559319.7	938980.3	1316	70.7
12	1246	562679.2	939835.0	1266	72.3
13	1258	569545.0	930325.2	1803	84
14	1262	572261.1	927201.8	2087	88.1
15	1266	573883.6	924316.9	2170	91.4
16	1274	573801.4	920288.8	2591	97.1
17	1278	575870.8	914139.1	2646	100.4
18	1281	576406.4	911657.7	2708	102.7
19	1289	576782.6	907614.1	2686	110.6
20	1315	575824.4	881534.0	2478	127.5
21	1324	5827885.4	876868.2	2476	135
22	1329	586371.7	873281.7	2477	140
23	1337	592024.6	867683.8	2489	147.7

Table 6.2 Chaffe Donsa shot -the Main Ethiopian Rift (Geophone No, Elevation, Location and Offset).

No	Geophone no	Location		Elevation	Offset (Km)
		Easting	Nor thing		
1	1315	575824.4	881534.0	3478	4.9
2	1320	580780.7	880525.8	2472	7.7
3	1324	582785.4	876868.2	2476	11.8
4	1329	586371.7	873281.7	2477	16.9
5	1333	588907.1	870164.0	2465	20.9
6	1337	592024.6	867683.8	2489	24.8
7	1340	594382.8	865813.7	2479	27.8
8	1344	597394.2	862730.6	2474	32.1
9	1346	598900.7	861529.1	2472	35.1
10	1351	60321.6	85997.1	2462	38.1
11	1356	607340.3	857006.3	2450	43.1
12	1360	611185.9	853924.2	2458	48.1
13	1366	617634.9	853858.0	2483	53
14	1370	621384.0	851750.5	2485	57.3
15	1373	625113.5	851838.1	2490	60.2
16	1377	628549.2	849876.2	2498	64.2

Table 6.3 Kula shot -Eastern Ethiopian plateau (Geophone No, Elevation, Location and Offset).

- The representative examined wave traces of the data for the regions are given in fig 6.1, 6.2 and 6.3 for the western, the main rift and eastern Ethiopian plateau respectively.

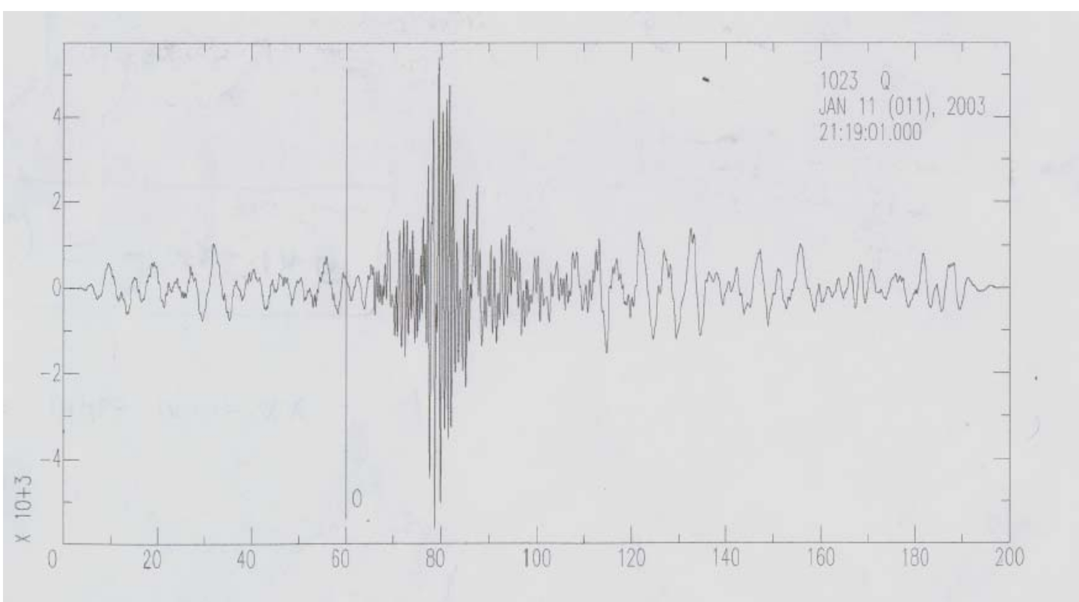


Fig 6.1 trace from Goha Tsion Shot (Geophone No 1023

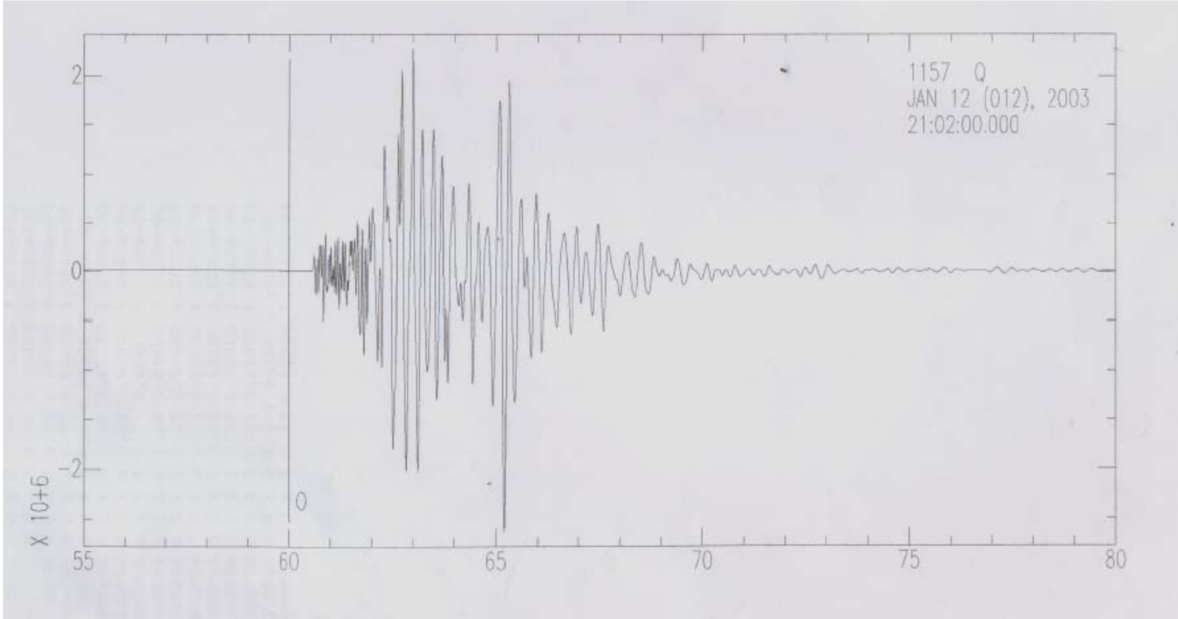


Fig 6.2 trace from chaffe Donsa Shot (Geophone No1157)

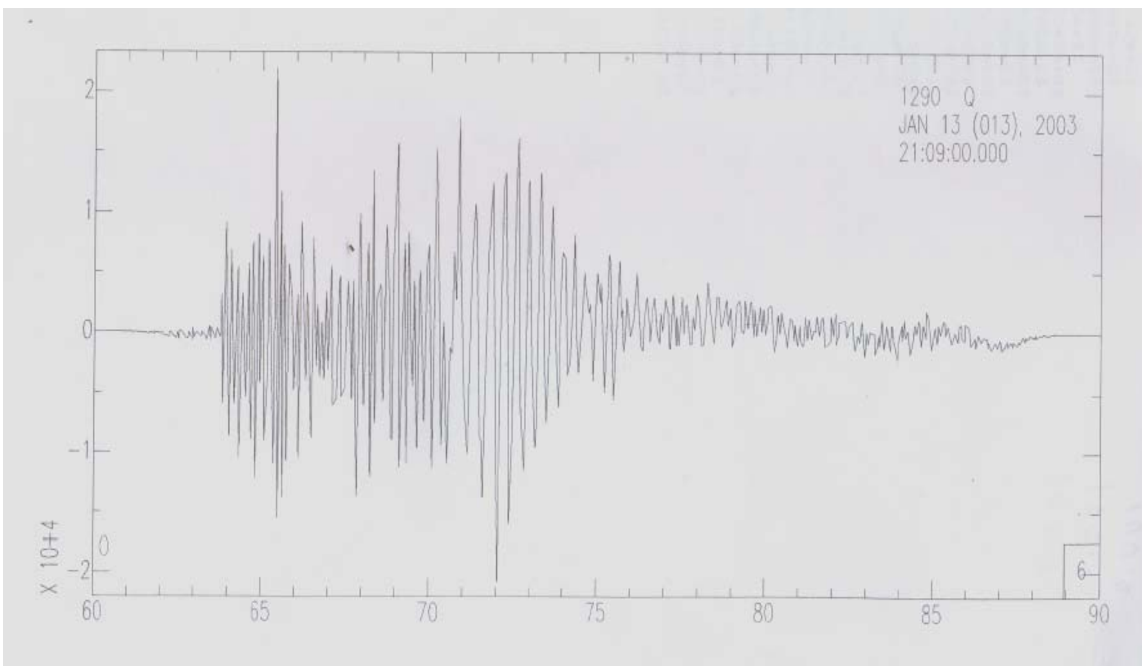


Fig 6.3 trace from Kula shot (Geophone No 1290)

6.2 Measurement of seismic attenuation coefficient α and Q^{-1} structure of the study area.

6.2.1 Measurement of attenuation coefficient α as a function of distance and variation of amplitude as a function of distance.

The measurement considers attenuation coefficient α measurements between the first geophone (the reference geophone) and any other geophone for the corresponding shot points.

The attenuation coefficient α is computed from the mathematically modeled attenuation in 5.5(b) given as $\alpha = \frac{1}{x_n - x_1} \ln\left(\frac{A_o x_o}{A_n x_n}\right)$ or in terms of inverse quality factor $Q^{-1} = \frac{\alpha v}{\pi f}$ the geometrical spreading is corrected by multiplying the whole geophone positions to the offset from the source and hence we analyze the seismic recordings that may be affected only due to intrinsic absorption provided the other attenuation mechanisms are negligible.

Amplitude and attenuation coefficients versus offset distance are plotted on a logarithmic scale and contrasting of seismic attenuation coefficient (with offset) for the main Ethiopian rift and adjacent plateaus are under taken under the discussion part of the study in chapter7.

- The results are presented in figures 6.4 for the western, 6.5 for the main Ethiopian rift and 6.6 for the eastern Ethiopian plateau.

Offsets (m)	Amplitude (cm/sec)	Attenuation coefficient α (Nep/m)
580	0.1587	0.0024903
0	0.09147	0.0007597
10210	0.07372	0.000593
13400	0.07266	0.0004594
17740	0.07266	0.0005653
22790	0.000873	0.0006291
24880	0.00006198	0.000481
29740	0.000262	0.0004002
37010	0.0001875	0.0003566
42180	0.0001654	0.0003373
44180	0.0001969	0.0002638
52420	0.0006556	0.0002657
58140	0.0001441	0.0001838
76800	0.000688	0.0001997
80300	0.000106	0.0001562
84200	0.00195	0.000173
89400	0.000206	0.0001742
91100	0.0001401	0.000167
94800	0.000151	0.0001538
101800	0.000192	0.0001634
113400	0.0000121	

Table6.4Attenuation Coefficient as a function of distance for the Western Ethiopian plateau (Goha Tsion shot)

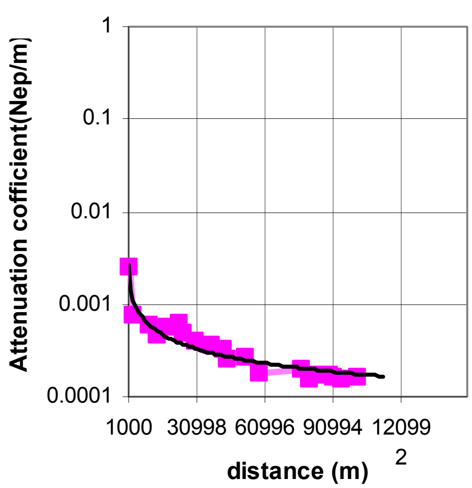


Fig 6.4 Attenuation coefficient as a function of distance curve for Western plateau

Offsets (m)	Amplitude (cm/sec)	Attenuation coefficient α (Nep/m)
1790	0.023	0.0022885
6350	0.00453	0.0010762
14900	0.00027	0.0007831
20300	0.00025	0.0006986
23200	0.00018	0.0003731
43900	0.00016	0.0002828
57600	0.000196	0.0002588
63700	0.00017	0.000248
66200	0.000186	0.000238
68800	0.000198	0.0002422
70700	0.000097	0.0002315
72300	0.000143	0.0002015
84000	0.00013	0.0001913
88100	0.000144	0.0001849
91400	0.000142	0.0001732
97100	0.00016	0.0001706
100400	0.00012	0.0001648
102700	0.00015	0.0001531
110600	0.000157	0.0001345
127500	0.00014	0.0001278
135000	0.000133	0.0001226
140000	0.000148	0.0001173
[illegible]	[illegible]	[illegible]

Table6.5Attenuation Coefficient as function of distance for the Main Ethiopian Rift (Chaffe Donsa shot)

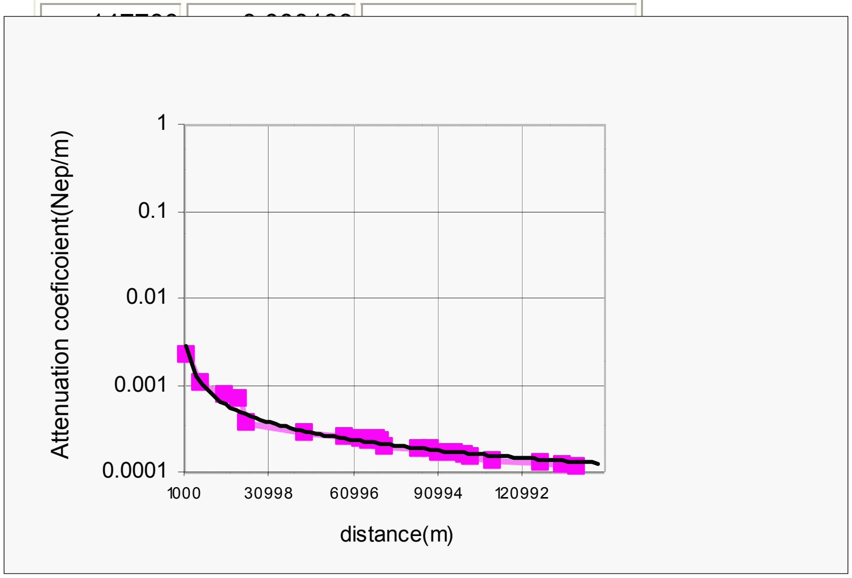


Fig 6.5 Attenuation coefficient as a function of distance curve for the Main Ethiopian Rift

Offset (m)	Amplitude (cm/sec)	Attenuation coefficient α (Nep/m)
4900	0.00267	0.004444232
7700	0.00232	0.001983005
11800	0.00103	0.00122712
16900	0.00052	0.000955738
20900	0.000365	0.000779139
24800	0.00035	0.000706058
27800	0.000202	0.000611915
32100	0.000145	0.000538239
35100	0.000234	0.000500825
	0.000175	0.000449719
43100	0.000114	0.000398263
48100	0.000124	0.00035231
53000	0.000177	0.000324567
57300	0.00018	0.000320187
60200	0.000094	0.000292956
64200	0.00014	

Table6.6Attenuation Coefficient as function of distance for Eastern Ethiopian plateau (Kula shot)

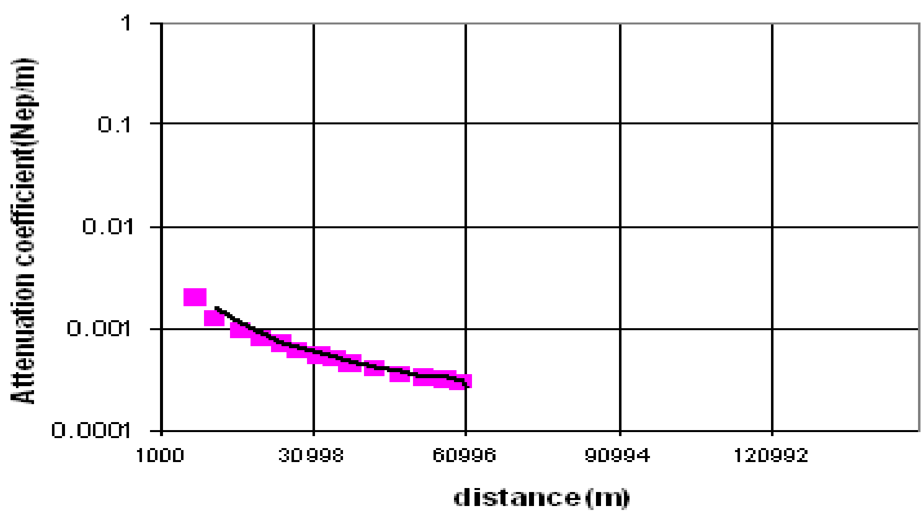


Fig 6.6 Attenuation coefficient as a function of distance curve for the Eastern Ethiopian plateau

♦ The Amplitude variations as a function of distance

The amplitude variation as a function of distance is seen from amplitude reading at different offset distances from the corresponding shot points.
Using the amplitude readings with offsets for the corresponding shot points, the following amplitude decay curve is obtained (Fig 6.7,6.8 &6.9) for Western, the main rift and Eastern Ethiopian plateau respectively.

Offset (m)	Amplitude (cm/sec)
4770	0.1094
9760	0.00447
12240	0.00119
20010	0.000393
23860	0.000131
26100	0.000131
32300	0.00012
40040	0.000109
44900	5.76E-05
48500	1.98E-05
52300	1.26E-05
57400	7.33E-05
68900	6.36E-06
62500	1.06E-06
63600	8.98E-06
69400	8.98E-06
80900	3.14E-05

Table6.7 Amplitude variations as a function of distance for western Plateau (GohaTsion shot)

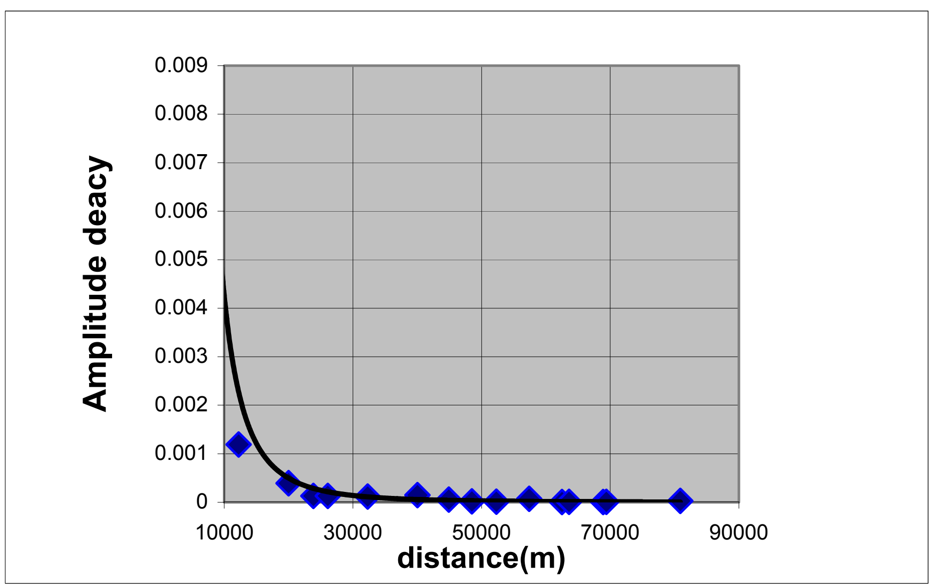


Fig 6.7 Amplitude decay as a function of distance curve for western Ethiopian plateau.

Offsets (m)	Amplitude (cm/sec)
1790	0.023
6350	0.00453
14900	0.00027
20300	0.00025
23200	0.00018
43900	0.00016
57600	0.000196
63700	0.00017
66200	0.000186
68800	0.000198
70700	0.000097
72300	0.000143
84000	0.00013
88100	0.000144
91400	0.000142
97100	0.00016
100400	0.00012
102700	0.00015
110600	0.000157
127500	0.00014
135000	0.000133
140000	0.000148
147700	0.000133

Table 6.8 Amplitude variations as a function of distance for Main Ethiopian Rift (Chaffe Donsa shot)

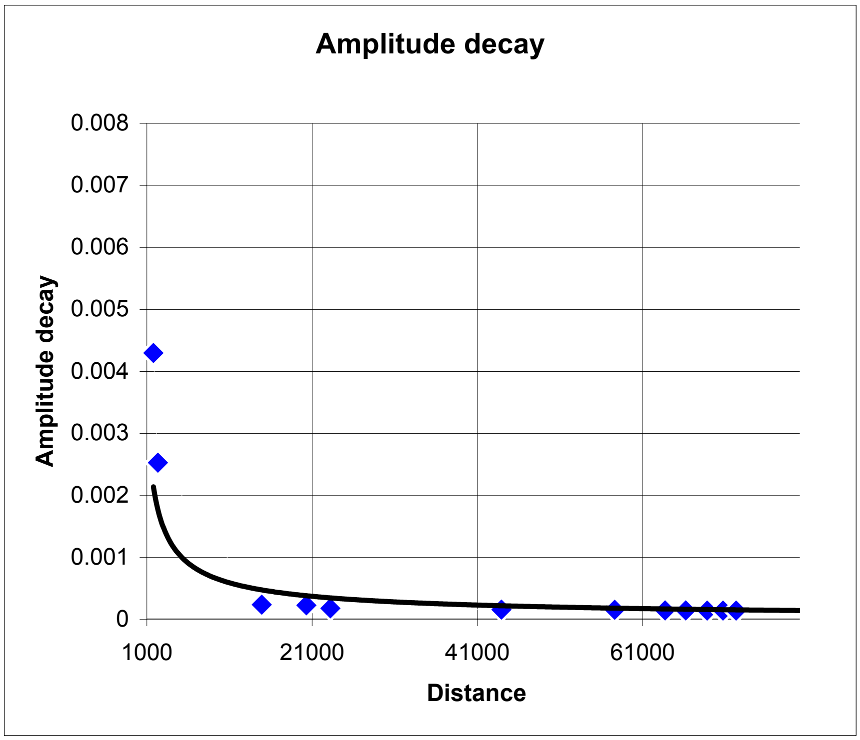


Fig 6.8 Amplitude decay as a function of distance for curve the Main Ethiopian Rift.

Offsets (m)	Amplitude (cm/sec)
4900	0.00267
7700	0.00232
11800	0.00103
16900	0.00052
20900	0.000365
24800	0.00035
27800	0.000202
32100	0.000145
35100	0.000234
38100	0.000175
43100	0.000114
48100	0.000124
53000	0.000177
57300	0.00018
60200	0.000094
64200	0.00014

Table6.9 Amplitude variations as a function of Distance for the Eastern Ethiopian (Kula shot)

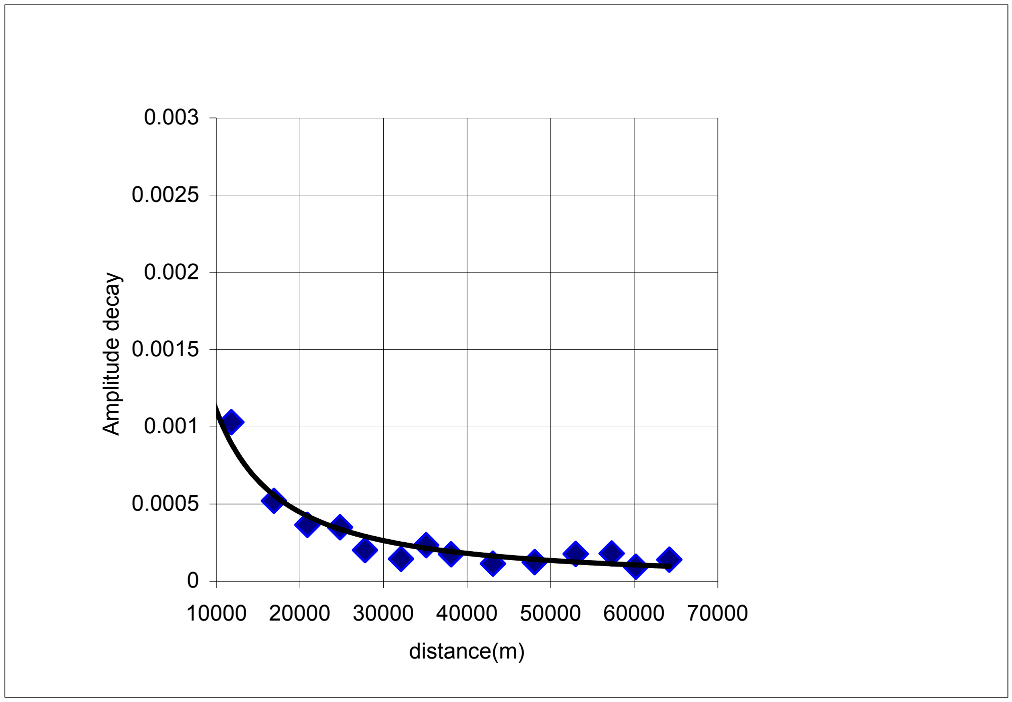


Fig 6.9 Amplitude decay as a function of distance curve for Eastern Ethiopian plateau.

6.2.2 Measurement of seismic attenuation coefficient α as function of frequency.

The attenuation of seismic energy occurs primarily as the result of three differing phenomena discussed under 5.4;

(i) Geometric spreading,

(ii) Scattering, and

(iii) Absorption of energy by the medium in which it is traveling. Geometric spreading is independent of the material in which the elastic wave is traveling. Scattering tends to be significant only for wavelengths close to the size of any heterogeneity in the medium. Absorption, or intrinsic attenuation, is a function of the characteristics of the material (grain size, mineralogy, saturation, etc.) and the frequency of the elastic wave. The intrinsic attenuation, Q^{-1}, can be defined as the energy loss per cycle as the stress wave propagates through the material (O'Connell & Budanisky 1978). The term Q^{-1} is known as the specific dissipation function, and Q as the quality factor of the material. In this measurement it is seen the interaction between the attenuation coefficient $\alpha(\omega)$ and the frequency of the wave for the three region under study. From the measured $\alpha(\omega)$, the inverse quality factor Q-1 of the region is computed .The inverse quality factor Q^{-1} is the physical parameter, which usually describes the intrinsic properties of the medium. From the equation (5.4.8) above

$\alpha\left(\frac{v}{f}\right) = \ln\left(\frac{A_o x_o}{A_n x_n}\right)$, and instead of taking A(x) in this equation, take the amplitude spectrum A(f) from a single geophone at a fixed offset for all regions. Then. $\alpha = \frac{f}{v}\ln\left(\frac{A_0(f)}{A(f)}\right)$. From equation (5.5.17) we see that the relation between $\alpha(\omega)$ and the specific dissipation factorQ^{-1} are given as $Q^{-1} = \frac{\alpha v}{\pi f}$. Then the computed attenuation coefficient with frequency $\alpha(\omega)$ and Q^{-1} are presented in Table 6.10,6.11 and 6.12 and the resulting attenuation coefficient with frequency curve and the Q^{-1} structure of the region are obtained.

Frequency (Hz)	Velocity (km/s)	Velocity (m/s)	Amplitude spectrum (cm-s)	Attenuation coefficient $\alpha(\omega)$ (Nep/m)	Q-1
0.454545	1.607	1607	1097	5.86907E-06	0.006608
0.47619	1.361	1361	1120	6.68608E-05	0.060858
0.5	1.282	1282	1328	0.000134819	0.110088
0.526316	1.245	1245	1550	0.000198161	0.149283
0.555556	1.225	1225	1753	0.00027282	0.191582
0.588235	1.216	1216	2002	0.000373858	0.246127
0.625	1.218	1218	2376	0.0004935	0.306284
0.666667	1.221	1221	2870	0.000608076	0.354679
0.714286	1.217	1217	3341	0.000723152	0.39239
0.769231	1.212	1212	3761	0.000864183	0.433633
0.833333	1.211	1211	4281	0.000995346	0.460648
0.909091	1.214	1214	4660	0.001009056	0.429138
1	1.208	1208	4221	0.000982587	0.378014
1.052632	1.178	1178	3595	0.00099677	0.35525
1.111111	1.132	1132	3347	0.001280337	0.415416
1.176471	1.103	1103	4043	0.001669123	0.498372
1.25	1.085	1085	5246	0.001935285	0.534977
1.333333	1.079	1079	5885	0.001599586	0.41225
1.428571	1.077	1077	4003	0.000811782	0.194905
1.538462	1.051	1051	2023	0.000896577	0.195063
1.5625	1.05	1050	2024	0.001499056	0.19498
1.587302	1.048	1048	3004	0.001528779	0.19345
1.666667	1.045	1045	3010	0.001677689	

Table 6.10 Attenuation coefficient as a function of frequency and dissipation factor Q-1 for western Ethiopian plateau.

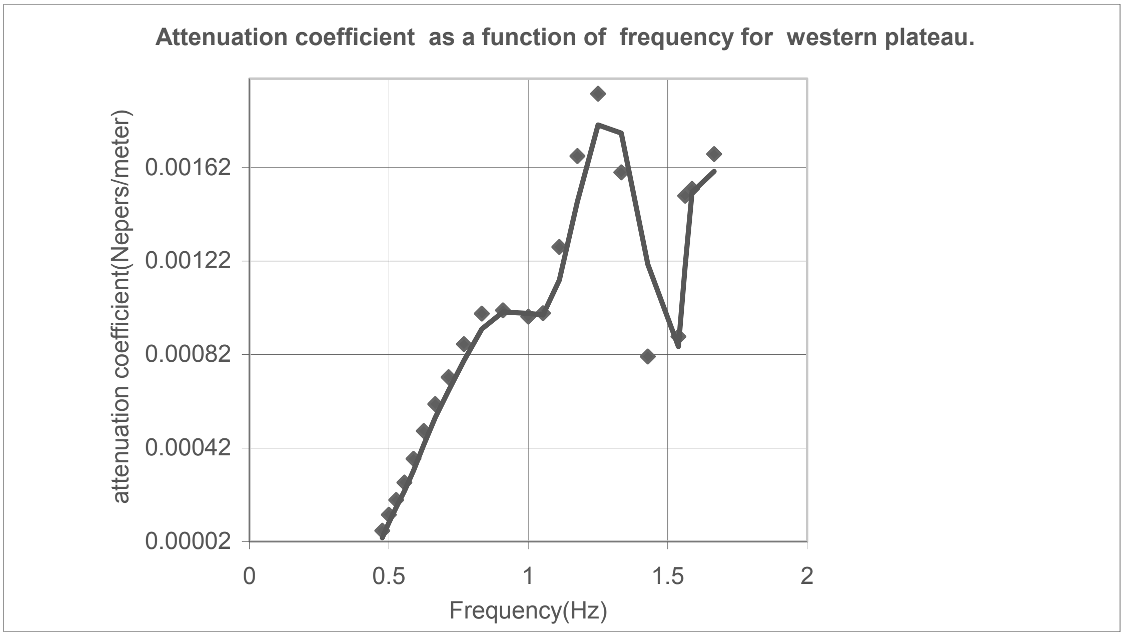


Fig 6.10 Attenuation coefficient as a function Of frequency curve for western Ethiopian plateau.

The Q^{-1} structure of western plateau is given in fig 6.11 just below.

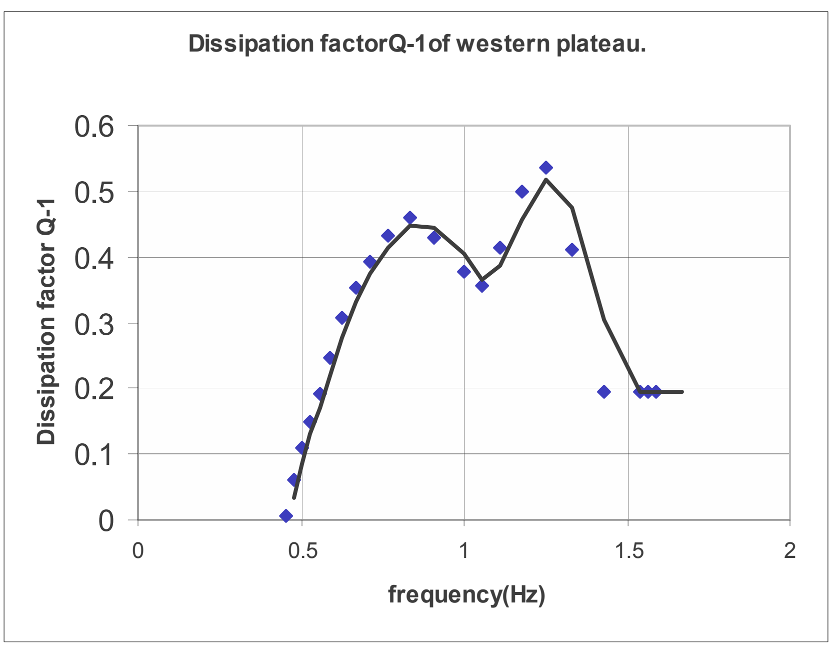


Fig6.11Dissipation factor Q^{-1} for western plateau.

- The measurement is taken at the offset distance of 22.789Km from Goha Tsion shot.

Period (sec)	Frequency (Hz)	Velocity (km/s)	Velocity (m/s)	Amplitude spectrum (cm-s)	Attenuation coefficient $\alpha(\omega)$ (Nep/m)	Q-1
0.5	2	0.939	939	79510	0.000604	0.090374386
0.48	2.083333	0.901	901	105600	0.001278	0.176057042
0.46	2.173913	0.887	887	138200	0.002032	0.264083498
0.44	2.272727	0.846	846	182200	0.002829	0.335358005
0.42	2.380952	0.811	811	227900	0.003752	0.407009205
0.4	2.5	0.776	776	285400	0.004695	0.46416079
0.38	2.631579	0.741	741	341500	0.005771	0.517527677
0.36	2.777778	0.714	714	403800	0.006799	0.556594236
0.34	2.941176	0.687	687	456500	0.007745	0.576143469
0.32	3.125	0.666	666	485400	0.008842	0.600147513
0.3	3.333333	0.64	640	523400	0.009873	0.603717382
0.29	3.448276	0.631	631	529300	0.010352	0.603295924
0.28	3.571429	0.618	618	528600	0.010852	0.598010728
0.27	3.703704	0.606	606	519900	0.011285	0.588055744
0.26	3.846154	0.592	592	503900	0.011681	0.57258067
0.25	4	0.576	576	480000	0.011969	0.548897735
0.24	4.166667	0.56	560	445600	0.013569	0.580768045
0.23	4.347826	0.548	548	492500	0.011681	0.468881713
0.22	4.545455	0.537	537	346600	0.010701	0.402627276
0.21	4.761905	0.529	529	281500	0.008925	0.315762453
0.2	5	0.511	511	214300	0.007595	0.247214031
0.19	5.263158	0.502	502	172800	0.009039	0.274554308
0.189	5.291005	0.5	500	188290	0.009642	0.290181966
0.18	5.555556	0.49	490	197760	0.010653	0.299231409
0.179	5.586592	0.488	488	203460		

Table 6.11 attenuation coefficient as a function of frequency and quality factor Q^{-1} for the Main Ethiopian Rift.

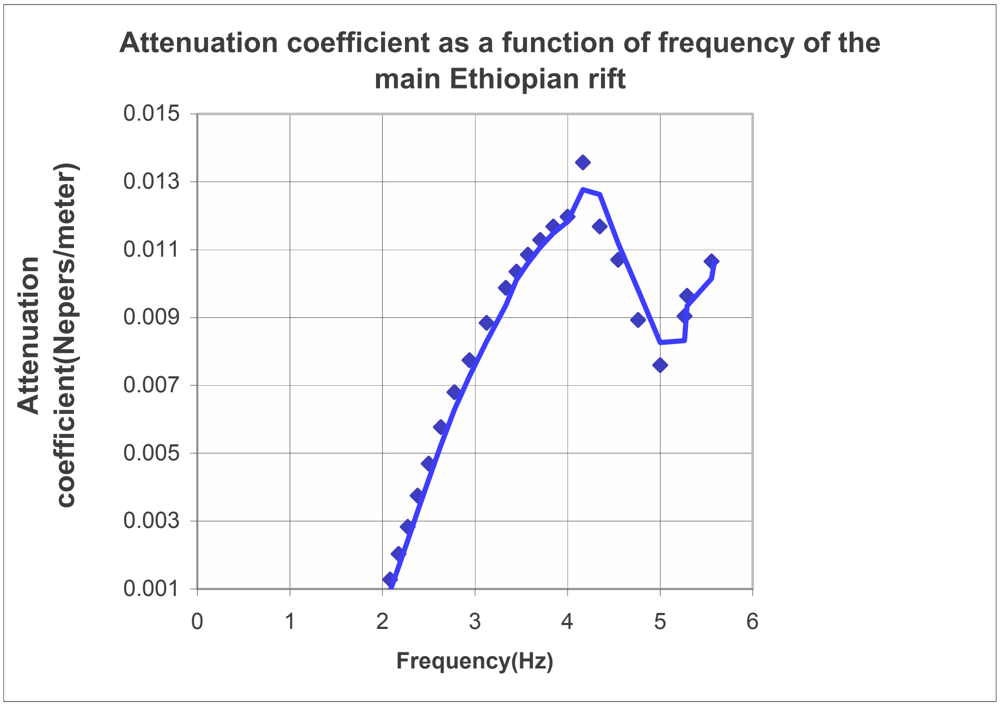


Fig 6.12 Attenuation coefficient as a function of frequency $\alpha(\omega)$ of the Main Ethiopian Rift.

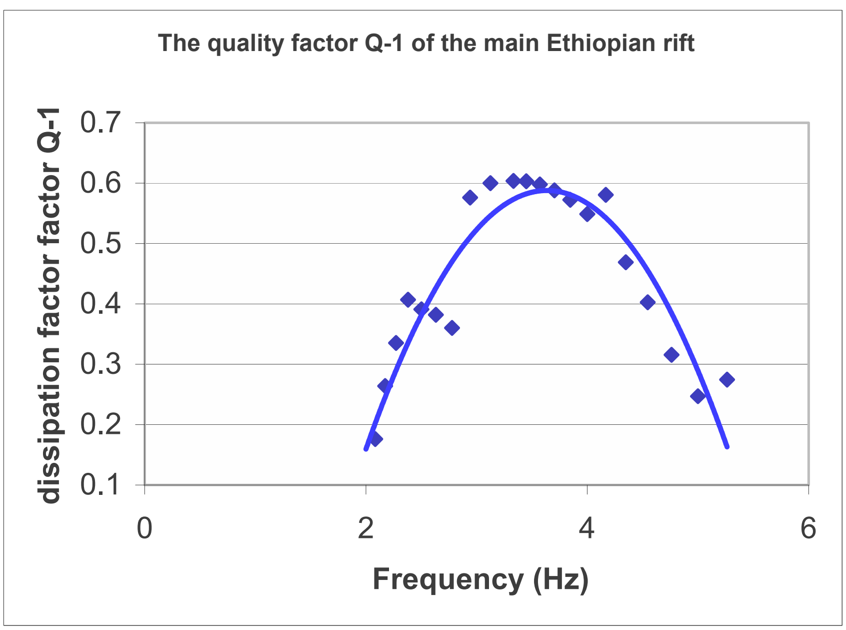


Fig 6.13 the inverse Quality factor Q^{-1} for the main Ethiopian rift.

- The measurement is taken at the offset distance of 1.79km from chaffe Donsa shot- 1157-sac.

Period (sec)	Frequency (Hz)	Velocity (km/s)	Velocity (m/s)	Amplitude spectrum (cm-s)	Attenuation coefficient (Nepers/meter)	Q-1
1.7	0.588235	1.988	1988	440.8	4.45361E-05	0.047934
1.6	0.625	1.936	1936	512.4	9.16986E-05	0.09046
1.5	0.666667	1.907	1907	585.6	0.000178547	0.162654
1.4	0.714286	1.843	1843	734.6	0.000304195	0.249963
1.3	0.769231	1.753	1753	966.3	0.000416952	0.302609
1.2	0.833333	1.699	1699	1140	0.000514755	0.33423
1.1	0.909091	1.723	1723	1259	0.000701519	0.423436
1	1	1.731	1731	1666	0.000729386	0.402091
0.95	1.052632	1.724	1724	1558	0.000764985	0.39901
0.9	1.111111	1.714	1714	1543	0.00117539	0.577438
0.85	1.176471	1.714	1714	2702	0.001309211	0.607449
0.8	1.25	1.718	1718	2969	0.001449778	0.634578
0.75	1.333333	1.713	1713	3233	0.001639807	0.670937
0.7	1.428571	1.698	1698	3624	0.001721249	0.651553
0.65	1.538462	1.675	1675	3410	0.002233385	0.774394
0.6	1.666667	1.641	1641	5015	0.002673531	0.838331
0.55	1.818182	1.599	1599	6130	0.003178966	0.890364
0.5	2	1.563	1563	7218	0.003617371	0.900311
0.48	2.083333	1.541	1541	7447	0.003851694	0.907332
0.46	2.173913	1.517	1517	7613	0.00412009	0.915631
0.44	2.272727	1.493	1493	7814	0.004424359	0.925621
0.42	2.380952	1.465	1465	8063	0.004774624	0.935613
0.4	2.5	1.435	1435	8320	0.005081557	0.928922
0.38	2.631579	1.409	1409	8147	0.005414405	0.923242
0.36	2.777778	1.387	1387	8003	0.00538309	0.856014
0.34	2.941176	1.367	1367	6480	0.004909756	0.726738
0.32	3.125	1.329	1329	4318	0.004149325	0.561983
0.3	3.333333	1.291	1291	2574	0.003736739	0.460904
0.29	3.448276	1.277	1277	1874	0.004053818	0.478105
0.28	3.571429	1.248	1248	1978	0.004915574	0.547038
0.27	3.703704	1.225	1225	2456	0.00532699	0.561115
0.25	4	1.222	1222	2567		

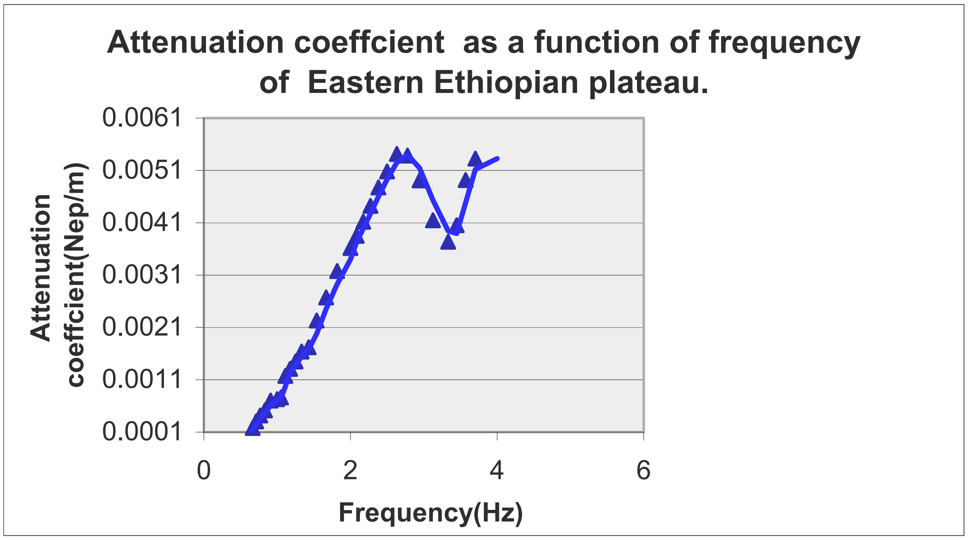


Fig 6.14 Attenuation coefficient as a function of frequency $\alpha(\omega)$ of the Eastern Ethiopian plateau.

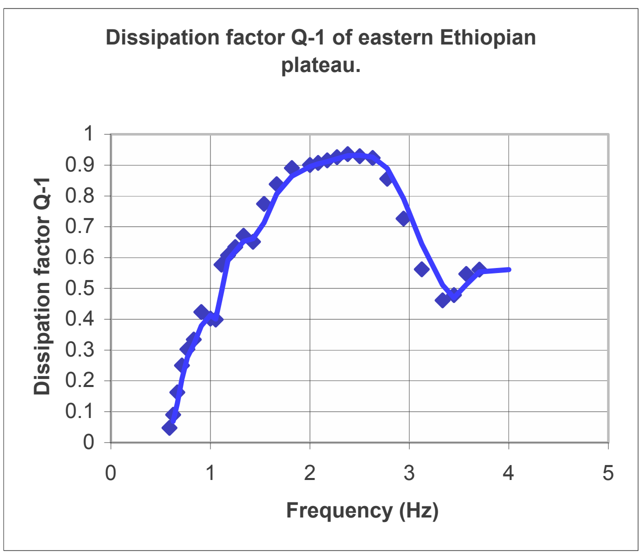


Fig 6.15 the inverse Quality factor Q^{-1} for Eastern plateau.

- The measurement is taken at the offset distance of 17.66km from Kula shot

- Comparison of computed attenuation coefficient as a function of Frequency curves for the regions under study.

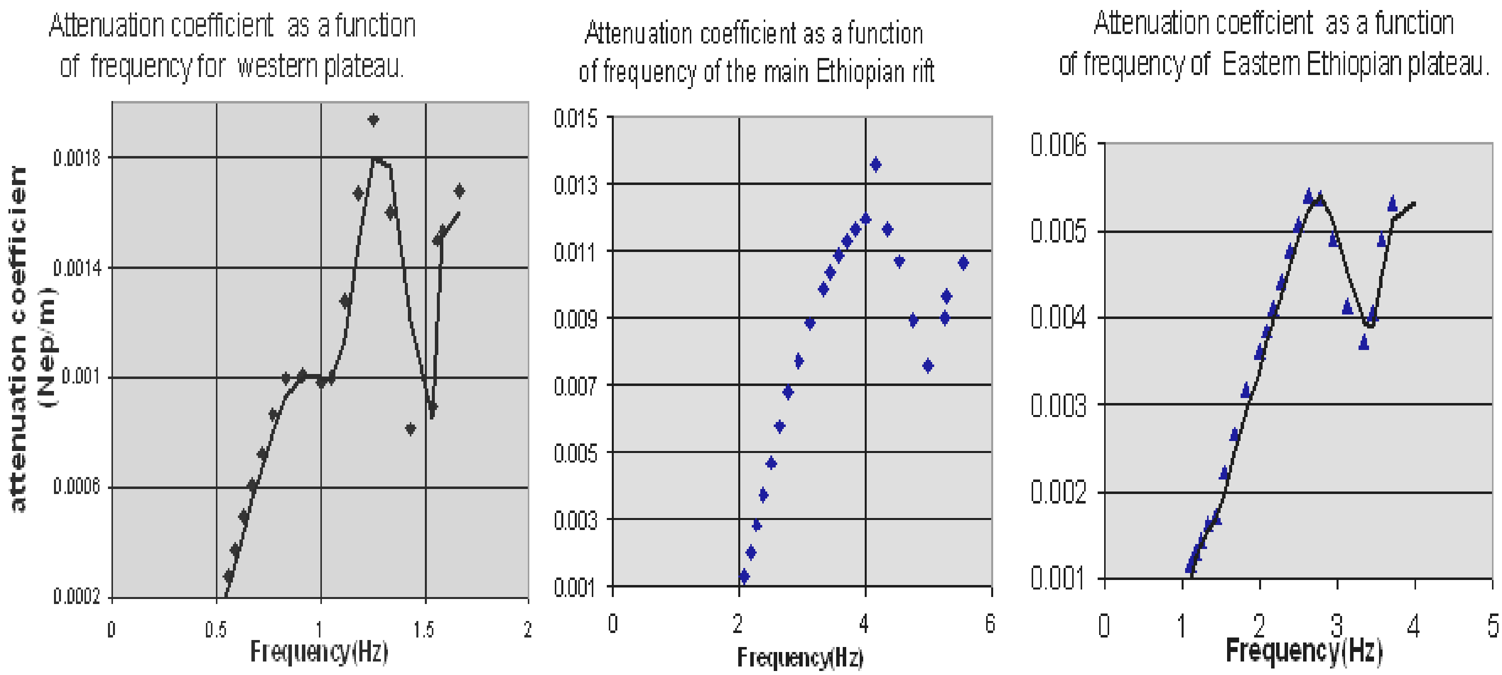


Fig 6.16-A when attenuation coefficient as a function of frequency curves for the region under study are compared.

❖ Comparison of dissipation factors Q^{-1} curves computed for the regions under study.

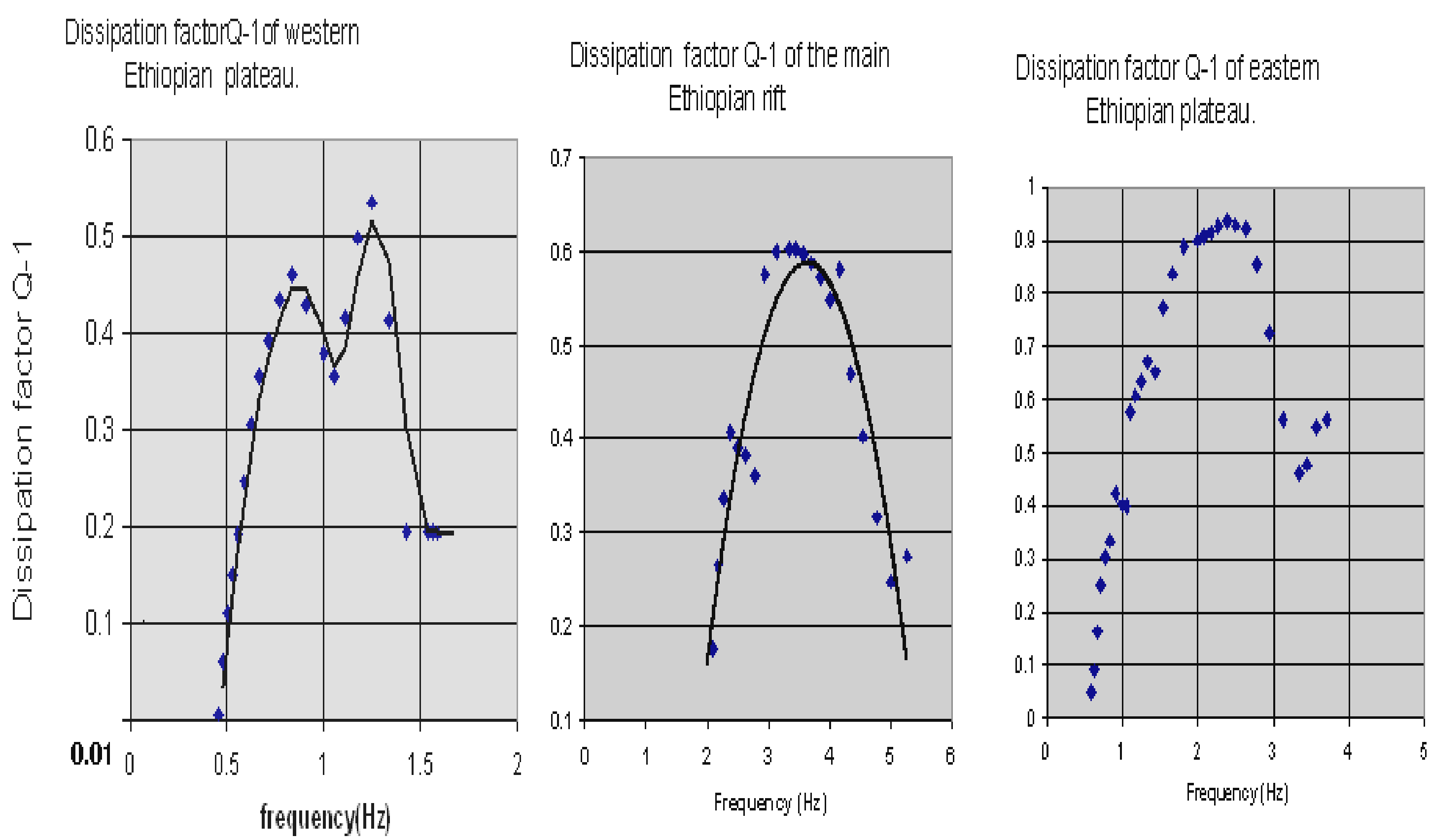


Fig 6.16-B when Dissipation factors Q^{-1} curves for the region under study are compared.

6.2.3 Measurement of seismic attenuation coefficient α as a function of Depth.

For the further discussions and interpretations from the results of attenuation coefficient as a function of frequency $\alpha(\omega)$ obtained above, we need to know depths of the corresponding attenuation coefficients for the frequency ranges we considered. Using period and phase velocity of the waves, one can calculate the depths of the frequency ranges for each attenuation coefficient curves for the regions. The depths are then calculated for the given frequency ranges from wavelength equation $\lambda = CT$, where C and T are velocity and period respectively. Then Depth $\approx$ wavelength, i.e. Depth =velocity times period $\Rightarrow$ $Z \approx VxT$. Then from this equation, the depths for each velocity and period for the regions under study are given in table 7.1,7.2 and 7.3. The attenuation coefficients as a function of frequency curves presented above are then associated with the corresponding computed depths. The possible geologic conditions that affect the shape of the attenuation curves and the attenuating behaviors of the layers of the earth materials under each region can then be identified.

Frequency (Hz)	Period (s)	Velocity (m/s)	$\alpha(\omega)$	Depth (m)
0.454545	2.2	1607	5.86907E-06	3535.4
0.47619	2.1	1361	6.68608E-05	2858.1
0.5	2	1282	0.000134819	2564
0.526316	1.9	1245	0.000198161	2365.5
0.555556	1.8	1225	0.00027282	2205
0.588235	1.7	1216	0.000373858	2067.2
0.625	1.6	1218	0.0004935	1948.8
0.666667	1.5	1221	0.000608076	1831.5
0.714286	1.4	1217	0.000723152	1703.8
0.769231	1.3	1212	0.000864183	1575.6
0.833333	1.2	1211	0.000995346	1453.2
0.909091	1.1	1214	0.001009056	1335.4
1	1	1208	0.000982587	1208
1.052632	0.95	1178	0.00099677	1119.1
1.111111	0.9	1132	0.001280337	1018.8
1.176471	0.85	1103	0.001669123	937.55
1.25	0.8	1085	0.001935285	868
1.333333	0.75	1079	0.001599586	809.25
1.428571	0.7	1077	0.000811782	753.9
1.538462	0.65	1051	0.000896577	683.15
1.5625	0.64	1050	0.001499056	672
1.587302	0.63	1048	0.001528779	660.24
1.5762	0.58	1046	0.00151234	358.8

Table 7.1 computed depths of western Ethiopian plateau.

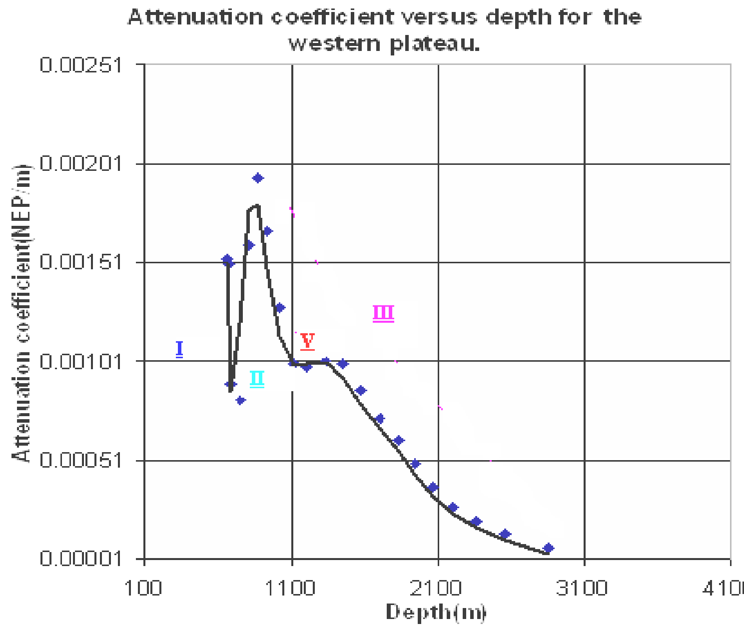


Fig 7.1 Attenuation coefficient versus depth for Western plateau.

❖ Similarly depths versus the corresponding attenuation coefficient for the main Ethiopian rift are presented in table 7.2.

Frequency (Hz)	Period (S)	Velocity (m/s)	Depth (m)	$\alpha(\omega)$ (Nep/m)
2	0.5	939	469.5	0.000604
2.083333	0.48	901	432.48	0.001278
2.173913	0.46	887	408.02	0.002032
2.272727	0.44	846	372.24	0.002829
2.380952	0.42	811	340.62	0.003752
2.5	0.4	776	310.4	0.004695
2.631579	0.38	741	281.58	0.005771
2.777778	0.36	714	257.04	0.006799
2.941176	0.34	687	233.58	0.007745
3.125	0.32	666	213.12	0.008842
3.333333	0.3	640	192	0.009873
3.448276	0.29	631	182.99	0.010352
3.571429	0.28	618	173.04	0.010852
3.703704	0.27	606	163.62	0.011285
3.846154	0.26	592	153.92	0.011681
4	0.25	576	144	0.011969
4.166667	0.24	560	134.4	0.013569
4.347826	0.23	548	126.04	0.011681
4.545455	0.22	537	118.14	0.010701
4.761905	0.21	529	111.09	0.008925
5	0.2	511	102.2	0.007595
5.263158	0.19	502	95.38	0.009039
5.291005	0.189	500	94.5	0.009642
5.555556	0.18	490	88.2	0.010653
5.586592	0.179	488	87.352	

Table 7.2 computed depths for the Main Ethiopian Rift

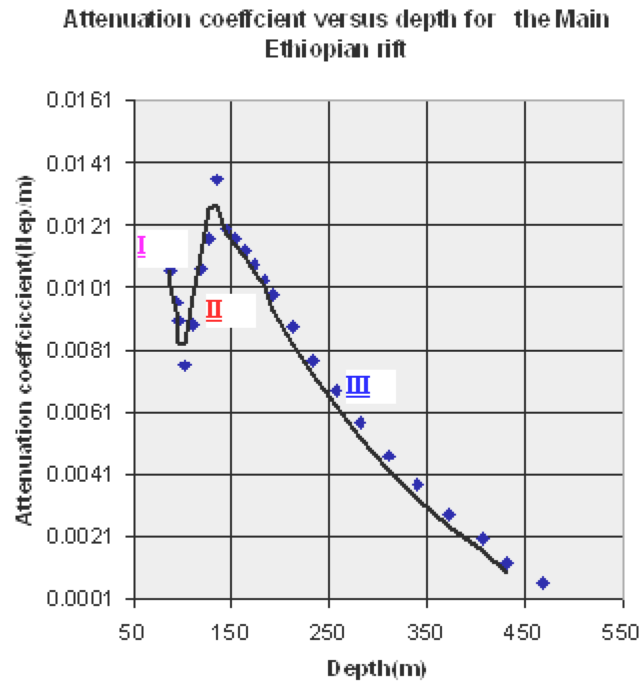


Fig 7.2 Attenuation coefficient versus depth for the Main Ethiopian Rift

❖ In a similar argument, depths versus the corresponding attenuation coefficient for the Eastern Ethiopian plateau is computed in table 7.3.

Frequency (Hz)	Period (s)	Velocity (m)	Attenuation coefficient (Nep/m)	Depth (m)
0.588235	1.7	1988	4.45361E-05	3379.6
0.625	1.6	1936	9.16986E-05	3097.6
0.666667	1.5	1907	0.000178547	2860.5
0.714286	1.4	1843	0.000304195	2580.2
0.769231	1.3	1753	0.000416952	2278.9
0.833333	1.2	1699	0.000514755	2038.8
0.909091	1.1	1723	0.000701519	1895.3
1	1	1731	0.000729386	1731
1.052632	0.95	1724	0.000764985	1637.8
1.111111	0.9	1714	0.00117539	1542.6
1.176471	0.85	1714	0.001309211	1456.9
1.25	0.8	1718	0.001449778	1374.4
1.333333	0.75	1713	0.001639807	1284.75
1.428571	0.7	1698	0.001721249	1188.6
1.538462	0.65	1675	0.002233385	1088.75
1.666667	0.6	1641	0.002673531	984.6
1.818182	0.55	1599	0.003178966	879.45
2	0.5	1563	0.003617371	781.5
2.083333	0.48	1541	0.003851694	739.68
2.173913	0.46	1517	0.00412009	697.82
2.272727	0.44	1493	0.004424359	656.92
2.380952	0.42	1465	0.004774624	615.3
2.5	0.4	1435	0.005081557	574
2.631579	0.38	1409	0.005414405	499.32
2.777778	0.36	1387	0.00538309	464.78
2.941176	0.34	1367	0.004909756	425.28
3.125	0.32	1329	0.004149325	387.3
3.333333	0.3	1291	0.003736739	370.33
3.448276	0.29	1277	0.004053818	349.44
3.571429	0.28	1248		330.75
3.703704	0.27	1225		305.5

Table 7.3 computed depths for the Eastern Ethiopian plateau.

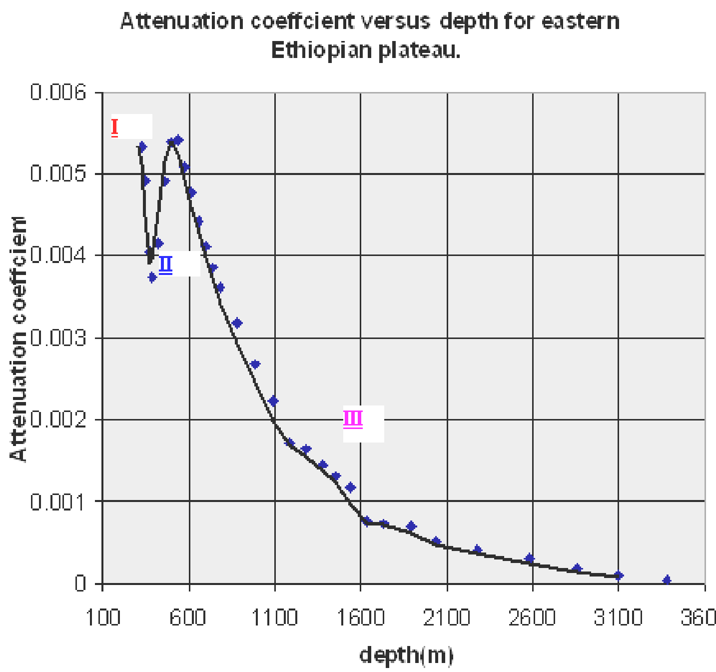


Fig 7.3 Attenuation coefficient versus depth for the Eastern Ethiopian plateau.

Chapter 7 Discussions, conclusions and recommendations

The measurements of attenuation coefficient (as a function of offset, frequency, and depth) and the dissipation factor Q-1 for the Western, the Main Ethiopian Rift and Eastern Ethiopian plateau could be analyzed to:

a) See the variation of attenuation coefficient and amplitudes of the waves with distance from the source. Variation of attenuation coefficient and amplitude of the waves as a function of distance for the aforementioned plateaus are presented in fig 6.4-6.9.

- *The attenuation coefficient as a function of distance for western Ethiopian plateau in fig 6.4 varies roughly from $3x10^{-3}$_$2x10^{-4}$ (Nep/m) in the distance range of 0.58km_101.8km(Goha Tsion shot).*
- *In the main Ethiopian rift the coefficient in fig 6.5 varies from $2x10^{-3}$_ $1x10^{-4}$(Nep/m) in the distance range of 1.79km_140km(chaffe Donsa shot)*
- *In Eastern Ethiopian plateau, it varies from 2x 10^{-3}-$3x10^{-4}$ (Nep/m) fig 6.6(Kula shot). The results from fig 6.7-6.9 for the regions under study verifies that the amplitude of the waves decay as the distance from source increases for all regions. Therefore, from the computed values of attenuation coefficient as a function of distance, we say that it decreases as the distance from the source increases for the whole regions.*

b) Investigate the relationship between the attenuation coefficient $\alpha(\omega)$ and the frequency content of the waves of the regions as it can be seen from the fig 6.10,6.12 and 6.14 of western, the main rift and eastern Ethiopian plateau respectively.

- *The rough values of attenuation coefficients as a function of frequency $\alpha(\omega)$ for western plateau in fig 6.10 increases from $6.8x10^{-5}$_$1x10^{-3}$Nep/m with frequency range of0.48Hz-0.91Hz and varies from $1x10^{-3}$_$9.8x10^{-4}$ in the frequency range of 0.91Hz-1Hz and then increases from $9.8x10^{-4}$_$1.9x10^{-3}$ with frequency range of 1.43Hz_1.7Hz.But it decreases from $1.9x10^{-3}$_$8x10^{-4}$ in the frequency range of 1.25HZ-1.43Hz and finally it increases from $8x10^{-4}$_$1.7x10^{-3}$ in frequency range of 1.43HZ-1.66Hz.*
- For *the main Ethiopian Rift on the other hand, it increases from $1.3x10^{-3}$_$1x10^{-2}$ (Nep/m) in the frequency range of 2.1Hz_4.2Hz.But it decreases from $1x10^{-2}$_$7.6x10^{-3}$(Nep/m) in the frequency range of 4.Hz-5Hz and finally it increases from $7.6x10^{-3}$_$1x10^{-2}$ in the frequency range of 5Hz_5.6Hz.*
- *In the Eastern Ethiopian plateau, it increases from $9.2x10^{-5}$_$5.4x10^{-3}$ (Nep/m) in the frequency range of 0.63Hz_2.6Hz and decreases from $5.4x10^{-3}$_$3.7x10^{-3}$(Nep/m) in the frequency range of 2.6Hz_3.3Hz and finally, it increases from $3.3x10^{-3}$_$5.3x10^{-3}$(Nep/m) in the frequency range of 3.3Hz_3.7Hz. Then from these values we*

generalize that the attenuation coefficient increases for small and high frequencies.

c) Get the dissipation factor Q-1 structure of the regions under study as it is represented in Fig 6.11,6.13 and 6.15 for the western, the main rift and eastern Ethiopian plateau respectively.

- *The dissipation factor Q-1 for Western Ethiopian plateau fig 6.11, increases from 6.6x10-3_0.5 in the frequency range of 0.45Hz_1.25Hz but, it decreases from 0.5-0.2 in the frequency range of 1.25Hz_1.43HZ and its peak value is then 0.5.*
- *For the Main Ethiopian Rift fig6.13, it increases from 0.17_0.7 in the frequency range of 2.1Hz_3.3Hz and decreases 0.7_0.25 in frequency range of 3.3Hz_5Hz and its peak value is 0.7*
- *And finally for Eastern Ethiopian plateau fig6.15, it increases from $4.7x10^{-2}$_0.9 in the frequency range of 0.6Hz_2.4Hz.But it decreases from 0.9_0.5 in frequency range of 2.4Hz_3.3Hz and its peak value is 0.9,which is the maximum dissipation factor for the whole regions. We can infer from these values that the dissipation factor Q^{-1} increases at low frequencies and decreases at high frequencies for all regions.*

d) Investigate the relationship between attenuation coefficient as a function of frequency and depth.

The analysis of the attenuation coefficient as a function of frequency associated with depth give us the geologic descriptions based on the attenuating behavior of the geology of the regions under study at various depths. Constructive ideas about the geology of the region have been obtained from a published article entitled as 'Sub-basaltic imaging of Mesozoic sediments using surface waves' (T.Mammo, etal: 2003)

- *From the attenuation coefficient as a function of frequency* $\alpha(\omega)$ *versus depth curve for Western Ethiopian plateau in fig 7.1, the Roman number I shows relatively less attenuating layers of the earth material in the depth intervals 380m_650m –upper part of the geology of the region. This depression of curve (less attenuating region) within this depth interval could be the signature of the Volcanics. The Roman number III shows relatively less attenuating layers of the earth material just like in I, in the depth interval 868m-3535m- lower parts. The geology of the region, in this depth interval could be basement. On the other hand, the Roman number II shows a high attenuating (peak curve) layers of the medium in depth interval* ***754m-*** *868m –the middle layer of the earth material, this layer in this depth interval could be the Mesozoic sediments. This means that the attenuating behavior of the upper part* *(volcanoes) and the lower part (basements) is high, and less in the middle part (sediments) in this study area.*

- *From fig 7.2 attenuation coefficient as a function of frequency* $\alpha(\omega)$ *versus depth curve of the Main Ethiopian Rift, it is seen that the Roman numbers I and III show relatively high attenuating layers of the earth material in the depth intervals 88m_100m - upper parts of the earth material and 144m_ 469m-lower parts of it respectively. The upper part and the lower part of this region in the given depth intervals respectively are volcanoes and basements. On the other hand, the roman number II shows less attenuating layer in the depth interval 102m-144m -the middle layer of the earth material (Mesozoic sediments). And there fore this again shows that the upper part of the geology of the main rift is less attenuating medium, which means, it is volcanic rocks. The middle part is the high attenuating medium, that is Mesozoic sediment and the lower part having less attenuating medium, which is the basement. Finally from fig7.3 of attenuation coeffcient versus depth for the Eastern Ethiopian plateau, we see that the Roman I-shows a less attenuating medium in the depth interval 330m-400m-upper part of the layer that is volcanoes in this depth interval. II-shows a high attenuating medium in the depth interval 400m-*

600m middle part of the layer that is Mesozoic sediment and III- shows a less attenuating medium just like in I, in the depth interval 600m-3380m-in the lower part layer of the region that is the basements.

➢

From the discussions and results obtained above, we can conclude the following.

❖ The attenuation coefficients and distances computed for all regions are inversely related i.e. as distance increases attenuation coeffcient decreases. Attenuation coeffcient as a function of distance is affected by distance from the source and the parameter of the source. The attenuation coeffcient versus distance curve obtained for all regions have more or less identical behaviors, even though the distances and the parameter of the sources taken are different. The attenuation coefficient as a function of frequency curves computed for all regions revels that attenuation coefficient increases with increasing frequency in some frequency ranges and decreases as frequency increases in another frequency ranges.i.e.the curves show peaks and valleys in the corresponding frequency ranges.

➢ The reasons for the high and low attenuating behaviors of any medium are described below.

♦ The high attenuating medium-characterizes, the medium may be [: - fractured [this makes rocks weaker, more Compliant (softer) which inurn absorbs the wave more.] [See5.4.2A]

- Highly weathered

-Fluid content-in which attenuation is high, especially in the presence of hydrocarbons and saturation of gas. [See5.4.2A]

- Presence of sediments and etc. [See5.4.2A]

The attenuation of seismic waves when traveling through viscous fluid-saturated rocks is generally higher than in dry rocks in most of the frequency bandwidth. Furthermore, changes in the spectral and amplitude characteristics of the seismic signal could be associated with the presence of fluids and fractures.

♦ However, the low attenuating medium shows the presence of consolidated rocks, volcanoes (igneous) rocks and less fractured medium.

In general, the upper part layers of all regions under study may be volcanic rocks -resulting in the low attenuating medium and the middle part layers may probably be sediments in which the wave attenuates more. The lower part layers of the regions could be basement rocks in which attenuation is low just like the upper part layer.

The measurement of seismic attenuation at western, the main rift and Eastern Ethiopian plateau has been analyzed to investigate the interrelationship between the attenuation coefficient and distance from the source, amplitude and offset, attenuation coefficient, amplitude spectrum and frequency content of the wave. The attenuating behaviors of regions under study for various ranges of frequency and depth are also discussed that gives rise to the various information about the medium.

Based on the results obtained from the study of seismic wave attenuation measurements across the main Ethiopian rift and adjacent plateaus, the following conclusions can be forwarded.

1) The attenuation coefficient determined for the studied area varies as a function of both offset and frequency of the wave.

2) The attenuation coefficient and amplitude decrease with distance and the curves have similar behaviors for the whole regions.

3) Attenuation coefficient as a function of frequency curve computed for the regions have similar trend or behaviors and characterized by peaks and valleys.

4) Depending on the attenuating behaviors of the regions attempts have been made to determine the various lithologies.

Based on the results obtained from the investigation and the possible geologic condition of the area, the following recommendations are forwarded.

1) In Ethiopia there is no attenuation model developed for exploration work, engineering study and earth quake hazard analysis. Therefore, the determined attenuation coefficient $\alpha(\omega)$ and dissipation factor Q^{-1} of the area under investigation can be taken as a reference for further study on seismic wave attenuation measurements.

2) A more comprehensive attenuation models that eliminate all factors that affect the attenuation result should be considered.

3) Further studies could be undertaken on the seismic wave attenuation measurements from identical source for the whole regions under study so that we can have attenuation contrasts between the regions.

BIBLIOGRAPHY

1. Aki, K. Attenuation of shear waves in the lithosphere for frequencies from 0.05 to 25 Hz, *Phys. Earth Planet. Inter.* **21,** 50-60. 1980.

2. Aki, K. and Richards, P.G., (1980), Quantitative Seismology, Theory and Methods, Vol. 1 W.H. Freeman & Co., New York.
3. Anderson, D.L., and R.S.Hart, Q of the earth, J.Geophys.Res., 83,5869-5882,1978.
4. Attewel, P.B., and Ramana, Y.V., 1966,wave attenuation and internal friction as a function of frequency in rocks: Geophysics v.31, P.1049.
5. Dobrin, M.b.and Savit C.H., 1988,Introduction to Geophysical prospecting. 3rd edition Mc
6. Eagle phase III; the controlled source seismic project, June 2003,field work report, Addis Ababa university Ethiopia
7. Fisseha Ayele; thesis 2006 A.A.U.
8. Frank D.Stacey, Physics of the earth second edition, university of Queens land Australia.
9. Futterman, 1962; W.I., Depressive body waves, J.Geophy.Res. 67,5279-5291, 1962
10. G.M.Di paola, 1972,The Ethiopian rift valley (Between $7^0$00' and 8^0 40' lat.north): Bulletin Volcanology, p.517-560.
11. G.Wolde Gebreil etal.1999, temporal relations and geotechnical features of felsic volcanism Geophysics, 35, 378-403,197
12. Gladwin and Stacy, Phys.earth and planetary introduction,p-332-336,1974
13. Gordon, R.B., and L.A.Davis, velocity and attenuation of seismic waves in imperfectly elastic rock, J.geophys.Res., 73,3917-3935,1968.
14. Graw Hill.Newyork. In the central sector of the main Ethiopian rift; Acta Vulcanologica-vol.11 (1) p.53-67
15. Johnston, D.H., and N.Toksoz, Attenuation of seismic waves in dry and saturated rocks (abstract), Geophysics, 42 1511,1977.
16. Kennett, B.L.N. (1983) Seismic Wave Propagation In Stratified Media, Cambridge University Press, 342 pp.

17. Kjartansson, E., Thermal relaxation, an attenuation mechanism for porous rocks (abstract), Eos trans. AGU, 59,324,1978.
18. Kjartansson, E., constant Q-wave propagation and attenuation, J.Geophysics.Res., 84,4737-748,1979.
19. Knopoff, L., Q, Rev.Geophys.space phys., 2,625-660,1964.
20. Kolsky, H., The propagation of stress pulses in viscoelastic solids, phys. Mag., 1,693-710,1956.La Jolla, CA 92093-0225,April 4, 2006
21. Liu, H.P., D.L.Anderson, and H.Kanamori, velocity dispersion due to anelasticity :Implications for seismology and mantle composition, Geophys.J, 47,41,1976.
22. Lomenitz, C., Linear dissipation in solids, J.Appl.phys., 28,201-205,1957. .
23. Love, A.E.H., A Treatise on the mathematical Theory of Elasticity, Dover, New York, 1944.
24. Mavko, G.M., and A.Nur, wave attenuation in partially saturated rocks, Geophysics, 44,161-178,1979.
25. McDonal, F.j., F.a.Angona, R.L Mills, R.L. Sengbush, R.G.Van Nostrand, and J.E. White, Attenuation of shear and comp rational waves in Pierre shale, Geophysics, 23,421-439,1958.
26. O'Connell, R.J., and B.J., and B.Budiansky, measures of dissipation in elastic media, Geophys.Res.Lett.5, 5-8,1978.
27. O'Doherty, R.F., and Anstey, N.A., 1971,Reflections on amplitudes: Geophys.prosp., v.19p.430-458. and Anstey, 1971;sheriff, 1975.
28. Paul .S. Havge measurement of attenuation from vertical seismic
29. Paul A.Mohr, 1971,The Geology of Ethiopia.
30. Peter M. Shearer, 227C—Body wave seismology, University of California, San Diego
31. Press, F., Seismic wave attenuation in the crust, J.Geophys.res., 69,4417-4418,196 Geophysics Vol. 46 Nov 1981
32. R.E.sherif exploration seismology, second edition University press
33. Ricker, N., The form and laws of propagation of seismic wavelets, Geophysics, 18,10-40,1953.
34. Ricker, N., Transient waves in visco-elastic media, 278pp., Elsevier, Amsterdam, 1977.
35. Robson, E.S &coruch, C: basic exploration Geophysics, John Wiley &Sons, New York, 1998.

36. Solomon, S.C., On Q and seismic dicrimination, Geophys.J., 31,163,1972. 1972.
37. Strick, E., A predicted pedestal effect for pulse propagation in constant Q-solids,
38. Telford W.M., Gildart L.p., Sherif R.E., Keys D.A., 1976:Cambrige, U.K.
39. Telford.W.B applied geophysics, Cambridge university press.
40. T.Mammo, etal: 2003, A.A.U.Sub-basaltic imaging of Mesozoic sediments using surface waves.
41. Walsh, J.B., Seismic wave attenuation in rock due to friction, J.Geophys.Res., 71,2591-2599,1966.
42. White, J.E., Static friction as a source of seismic wave attenuation, Geophysics, 31,333-339,1966.
43. Zener, C., Elasticity and Anelasticity of metals, University of Chicago Press, Chicago, III, 1948.
44. Zywicki, D.J. (1999) "Advanced Signal Processing Methods Applied to Engineering Analysis of Seismic Surface Waves." Ph.D. Dissertation, Georgia Institute of Technology.

DECLARATION

I declare that this thesis is my original work and has not been prepared for any degree in any university, and that all the sources of materials used for the thesis have been duly acknowledged.

Dr.Tilahun Mammo_________________ (Advisor)

Signature___________________

Geremew Lamessa

Signature_____________________

Date and place of submission, July 2006,Addis Ababa University