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SEISMIC METHODS: REFRACTION AND SEISMIC MONITORING

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ABSTRACT

Seismic methods are applied in the study of the earth's subsurface physical characteristics. Seismic methods involve measurement of impulses generated on or below the earth surface and recorded by a receiver on the surface. The impulses which are the seismic wave are generated either artificially by an energy source such as an explosive or naturally through geological processes. Travel time of the seismic wave sent or received through a geological setting, depends on: depths of studied structures and propagation velocities of seismic waves along paths of their propagation from the source to the receiver.

In seismic refraction method, the results give generalized expressions relating travel time, offset distance and velocity to thickness of subsurface layers. This is used in the determination of the thickness of layers below the surface. While in seismic monitoring, applies passive seismic methods to locate earthquakes, their distribution and their intensities. Modelling the nature and the intrinsic characteristic of the earthquakes is applied in mapping of the geological and or geothermal properties/characteristic suitable for resource exploitation.

1. INTRODUCTION

Geophysical methods generally are applied to delineate the variation of the earths physical properties from inferring on the physical parameter measured on the surface (Hersir and Björnsson, 1991; Roy, 2008; Telford et al., 1990). Seismic methods as one of the geophysical methods, use seismic wave to characterise the internal structure of the solid earth by the distribution of physical properties that affect the seismic wave propagation (Stein and Wysession, 2003).

Seismic methods deals with distribution of elastic properties, or velocities and densities in the subsurface. In their application to geothermal exploration, the physical properties measured are interpreted to fit the different geological conditions that best outlines the properties of a geothermal system as described in various research work (Árnason et al., 2000; Flóvenz et al., 2005; Foulger, 1982; Georgsson, 2014; Hersir and Björnsson, 1991; Mariita, 2009; Onacha et al., 2005, 2007, 2010; Simiyu, 2000; Simiyu and Keller, 2000, 2001; Simiyu and Malin, 2000).

Application of seismic methods in exploration therefore requires one to appreciate:

a) Basic seismology and how seismic waves propagate through a medium;
b) How seismic refraction is used in the determination of earth's velocity structure; and

c) Application of seismic monitoring in determination of various physical characteristics of the subsurface

2. BASIC SEISMOLOGY

2.1 Introduction

Seismology is the study of the generation, propagation, and recording of elastic waves in the Earth and of the sources that produce them. This elastic waves are due to deformational energy or elastic disturbances that expand spherically outward from the source as a result of transient stress imbalances in the elastic media in which they occur (Lay and Wallace, 1995; Shearer, 2009).

The sources of the seismic wave are classified as either natural sources that includes plate tectonic movements, ocean tides, volcanic eruptions, meteoritic impacts and or artificial sources such as induced explosions and heavy mechanical vibrations (Bormann, 2002).

The theory of wave propagation in elastic media has been described by various authors and gives the characteristic properties of an elastic wave as it is travels through a given medium (Aki and Richards, 2002; Dobrin and Savit, 1988; Shearer, 2009; Stein and Wysession, 2003; Telford et al., 1990).

2.2 Wave propagation

Seismic waves are elastic waves and the theory of linear elasticity describes their characteristics. This is because they do not cause a permanent deformation of the material in which they propagate (Dobrin and Savit, 1988). Where, seismic waves propagate as patterns of particles travelling through a medium with velocities that depend upon their elastic properties and densities.

The theory of elasticity provides mathematical relationships between stress and strain in a medium. Seismology is concerned with very small deformations (relative length changes of $\approx 10^{-6}$) over short periods of time (Dobrin and Savit, 1988; Lay and Wallace, 1995; Telford et al., 1990).

Deformations within a medium are composed of components that involve length changes and angular distortions called strain. Strain components depend linearly on derivatives of the displacement components permitting only very small strains and small spatial derivatives in the displacement field. The trace of the strain tensor is called the cubic dilatation, θ (Equation 1) illustrated in Figure 1.

$$\theta = \frac{\partial u_1}{\partial x_1} + \frac{\partial u_2}{\partial x_2} + \frac{\partial u_3}{\partial x_3} = \nabla \cdot u \qquad (1)$$

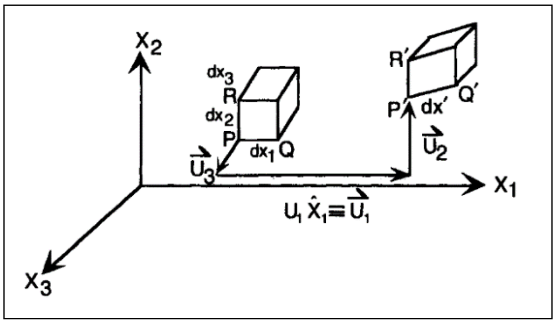

FIGURE 1: Displacements of a small cubic volume with a corner at point P to a new position with corner at p' (Lay and Wallace, 1995)

Stress is defined as a force acting on a unit area. Stress, which is a vector, has both magnitude and direction. For a linear deformation of an elastic medium (Figure 2), stress is directly proportional to strain, and the proportionality constant is defined as Young's Modulus (Equation 2).

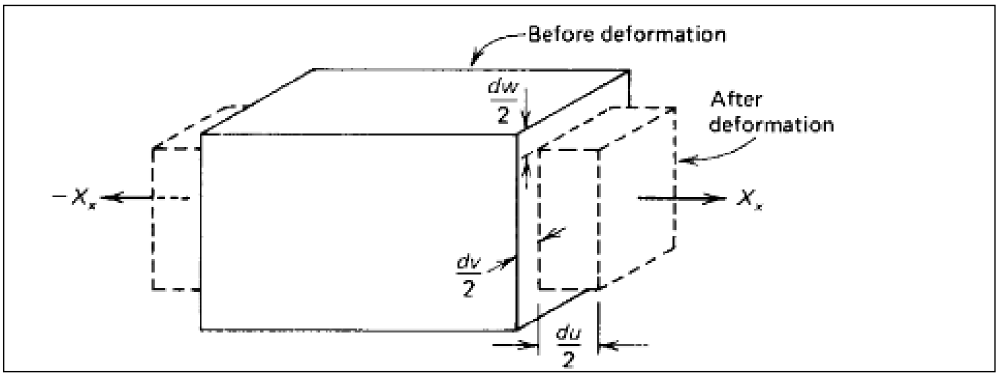

FIGURE 2: Deformation of a rectangular block (Dobrin and Savit, 1988)

$$E\frac{\partial u}{\partial x} = \mathrm{Xx} - \sigma\mathrm{Yy} - \sigma\mathrm{Zz}$$

$$E\frac{\partial v}{\partial y} = -\sigma\mathrm{Xx} + \mathrm{Yy} - \sigma\mathrm{Zz} \qquad (2)$$

$$E\frac{\partial w}{\partial z} = -\sigma\mathrm{Xx} - \sigma\mathrm{Yy} + \mathrm{Zz}$$

where E = Young's modulus;
Xx = Stress;
$\frac{\partial u}{\partial x}$ = Strain; and
σ = Poisson ratio.

If the stress result from excess pressure ΔP above the ambient pressure, all three of the stress components will be the same and each will be equal to ΔP **(**Equations 3 and 4**)**.

$$E\left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z}\right) = (1 - 2\sigma)(\mathrm{Xx} + \mathrm{Yy} + \mathrm{Zz}) \qquad (3)$$

and

$$\frac{\left(\frac{\Delta V}{V}\right)}{\Delta P} = \frac{\theta}{\Delta P} = \beta; therefore, \frac{1}{\beta} = k = \frac{E}{3(1 - 2\sigma)} \qquad (4)$$

Where β = Compressibility; and
k = Bulk modulus.

This formula relates the constant for cubical dilatation resulting from pressure to the constants relating linear strain and liner stress. For small deformations involved in seismic wave propagation, shear stress is proportional to shear strain with the proportionality constant defined as rigidity modulus, μ **(**Equation 5 and 6**)**.

$$Xy = Yx = \mu\alpha = \mu\left(\frac{\partial u}{\partial y} + \frac{\partial v}{\partial x}\right) \qquad (5)$$

where

$$\mu = \frac{E}{2(1+\sigma)}$$

and

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

In addition,

$$\lambda = \frac{E\sigma}{(1+\sigma)(1-2\sigma)}$$

where λ is one of Lame's coefficients.

Therefore, the Hook's relations for all stress components in terms of strains are given as:

$$Xx = 2\mu\frac{\partial u}{\partial x} + \lambda\theta$$

$$Yy = 2\mu\frac{\partial v}{\partial y} + \lambda\theta \qquad (6)$$

$$Zz = 2\mu\frac{\partial w}{\partial z} + \lambda\theta$$

2.3 Equation of motion

The general form of wave equation, which is more applicable to the propagation of seismic waves through the earth, assumes deformation in three directions, each component of stress being associated with strain in more than one direction (Dobrin and Savit, 1988). A simple case in which stress and stain is confined in a single direction is illustrated in Figure 3.

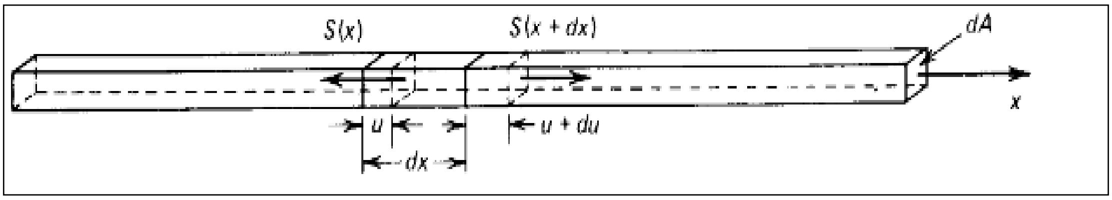

FIGURE 3: Elastic deformation in an element caused by longitudinal stress along axis X and stress is S(x) (Dobrin and Savit, 1988)

According to Newton's second law of motion,

$$\big(S(x+\partial x) - S(x)\big)\partial A = \rho\frac{\partial^2 u}{\partial t^2}\partial x \partial A \qquad (7)$$

where $\big(S(x+\partial x) - S(x)\big)\partial A$ = The net force;
$\frac{\partial^2 u}{\partial t^2}$ = The acceleration; and
$\rho \partial x \partial A$ = Mass.

Therefore,

$$S(x+\partial x)-S(x)=\left(\frac{\partial S}{\partial x}\right)\partial x$$

which is equivalent to

$$S=E\frac{\partial u}{\partial x}$$

Hence with appropriate substitution,

$$E\frac{\partial^2 u}{\partial x^2}\partial x\partial A=\rho\frac{\partial^2 u}{\partial t^2}\partial x\partial A \quad (8)$$

And on simplification,

$$\frac{\partial^2 u}{\partial x^2}=\frac{\rho}{E}\frac{\partial^2 u}{\partial t^2} \quad (9)$$

This is the classical one-dimensional wave equation of motion which is equivalent to**:**

$$\frac{\partial^2 q}{\partial x^2}=\frac{1}{v^2}\frac{\partial^2 u}{\partial t^2} \quad (10)$$

where v is the velocity of propagation and a convenient solution for a harmonic wave is:

$$q=Asin(k(Vt-x)) \quad (11)$$

For a three-dimensional case, the equation of motion for a compressional wave is:

$$\frac{\partial^2 \theta}{\partial x^2}+\frac{\partial^2 \theta}{\partial y^2}+\frac{\partial^2 \theta}{\partial z^2}=\frac{\rho}{\lambda+2\mu}\frac{\partial^2 u}{\partial t^2} \quad (12)$$

and

$$Vp=\sqrt{\frac{\lambda+2\mu}{\rho}}$$

where Vp = The velocity of the compressional (P) wave.

For shear strain, the three-dimensional equation of motion is:

$$\frac{\partial^2 \alpha}{\partial x^2}+\frac{\partial^2 \alpha}{\partial y^2}+\frac{\partial^2 \alpha}{\partial z^2}=\frac{\rho}{\mu}\frac{\partial^2 u}{\partial t^2} \quad (13)$$

where α is the shear strain and the equation of velocity of the shear (S) wave is

$$Vs=\sqrt{\frac{\mu}{\rho}}$$

2.4 Types of seismic waves

A seismic wave is the transfer of energy of by way of particle motion. The particle motion characterises the type of seismic wave. The seismic waves a classified into two basic categories: body waves and surface waves.

2.4.1 Body waves

Body waves propagate outward in all directions from a source (such as an earthquake) and travel through the interior of the Earth.

They include:

a) Compressional/primary/p-wave; and
b) Shear/secondary/s-wave.

Compressional waves

Particle motion associated with compressional waves consists of alternating condensations and rarefactions of adjacent particles in a solid a medium. The motion of particles is always in the direction of wave propagation as illustrated in Figure 4.

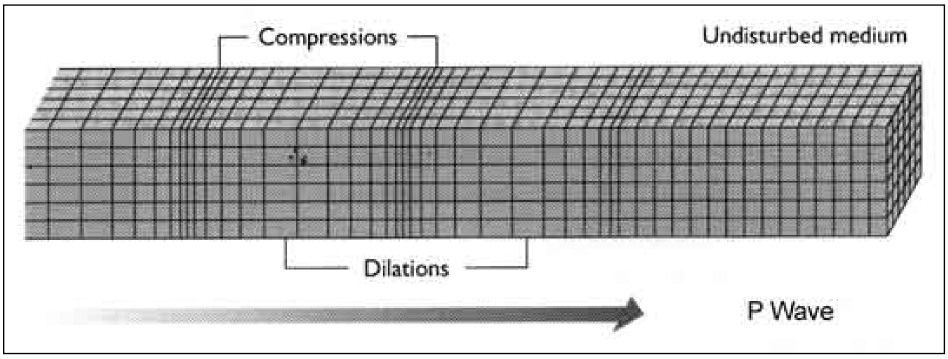

FIGURE 4: Particle separation during the Passage of compressional pulse (Rogers, 2012)

Shear waves

Shear waves also called Secondary or S waves, travel slower than compressional waves and they don't change the volume of the material through which they propagate, they shear it. S-waves are transverse waves because they vibrate the ground in a direction transverse or perpendicular, to the direction that the wave is travelling (Bormann, 2002, Shearer, 2009; Telford et al., 1990). The motion of individual particles is always perpendicular to the direction of wave propagation.

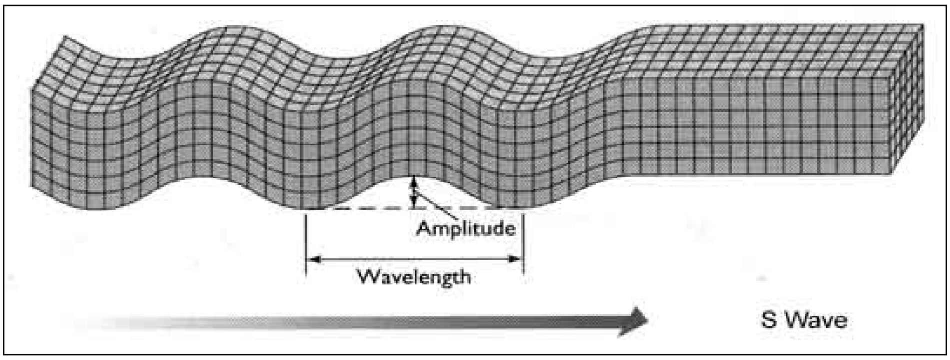

FIGURE 5: Particle deformation along line of wave travel during passage of shear pulse through solid material (Rogers, 2012)

Surface waves

Surface waves propagate approximately parallel to the Earth's surface. There are two types of surface waves: Rayleigh waves and Love waves. For laterally homogeneous models, Rayleigh waves are radially polarized (P/SV) and exist at any free surface, whereas Love waves are transversely polarized and require some velocity increase with depth (or a spherical geometry) Figure 6. Surface waves are generally the strongest arrivals recorded at teleseismic distances and they provide some of the best constraints on Earth's shallow structure and low-frequency source properties (Shearer, 2009).

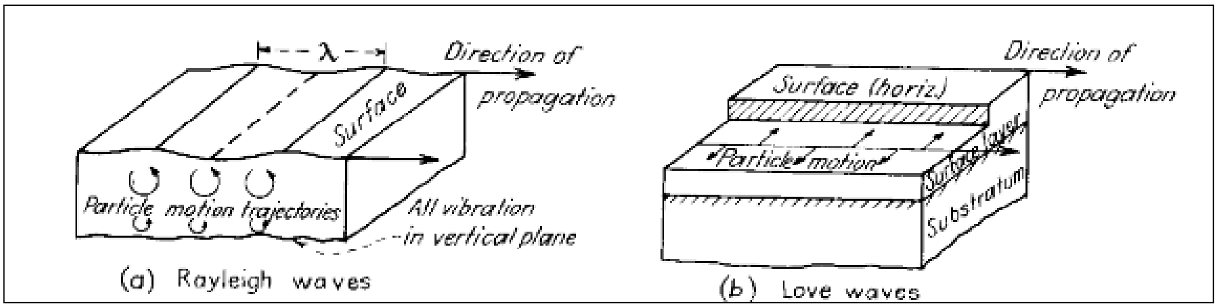

FIGURE 6: Characteristics of Rayleigh waves (a) and Love waves (b) (Dobrin and Savit, 1988)

2.5 Seismic instruments

2.5.1 Seismograph

A seismograph is a system that produces a permanent recording of ground motion (Bormann, 2002). Since the ground motion is a vector, a seismograph is composed of independent components that record the vertical and horizontal ground motions.

The seismograph is composed of:

a) Seismometer: a harmonic oscillator that reacts in a predictable way to ground motion.
b) Sensor: converts the mechanical output from the seismometer into a form of energy that can be recorded.
c) Recorder: produces a permanent time history of the ground motion.

2.6 Other terminologies

Other terms commonly encountered in seismology as described in (Shearer, 2009; Stein and Wysession, 2003; Telford et al., 1990) are:

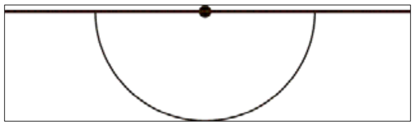

FIGURE 7: Seismic wave-front

Seismic wave-front: A 3D-surface of constant phase (Figure 8).

Seismic velocity: The speed at which a wave-front travels through a material, and depends on the material's elastic properties. In a homogeneous medium, a wave-front is spherical.

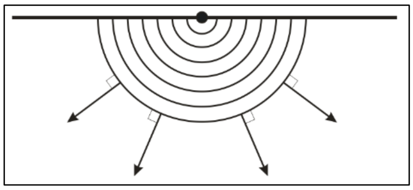

FIGURE 8: Seismic ray

Seismic ray: Is the infinite number of wave-front normal/arrows perpendicular to the wave-front. Indicates the direction of travel at that point on the wave-front.

Huygens' principle: Every point on a wave-front is considered to be a Huygens' source that give rise to another circular wave-front in the direction the wave is traveling.

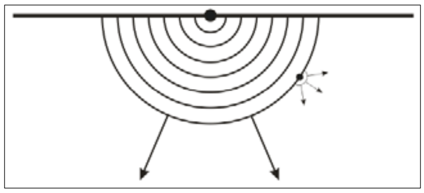

FIGURE 9: Huygen's principle

3. SEISMIC REFRACTION

The seismic refraction method uses the seismic wave that is received on the surface after travelling through a refracted ray path. The results give a generalized expressions relating travel time, offset distance and velocity to thickness of subsurface layers. This is used in the determination of the thickness of layers/earth structures below the surface.

Seismic refraction is based on Snell's law (Equation 14) which relates the refraction of seismic wave at the boundary between subsurface/geological layers of different velocity (Figure 10).

$$\frac{\sin i}{\sin r} = \frac{V_1}{V_2} \qquad (14)$$

where i = Incident angle of ray from source;
r = Angle of refraction; and
V_1 and V_2 = Velocities in media 1, 2 respectively.

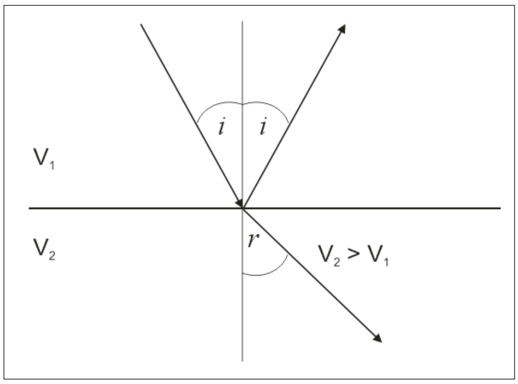

FIGURE 10: Wave refraction and reflection

At the Critical Angle of incidence i_c, the angle of refraction is $r = 90^o$.

Assuming a simple two layered model, seismic wave propagated from a source, will either be: a direct wave, refracted/head wave or reflected wave. Successive positions of the expanding wave-fronts for direct and refracted waves through a two-layer model with only the wave-front of the first arrival phase shown in Figure 11 (Kearey et al., 2002).

The time a wave is received by a receiver is called the arrival time. This is used in the computation of travel time (time taken by a wave to travel from a source to the receiver). Together with the distances between receivers, a distance travel time curve is plotted for both the direct and the refracted wave and related to the velocity model illustrated in Figure 12 below.

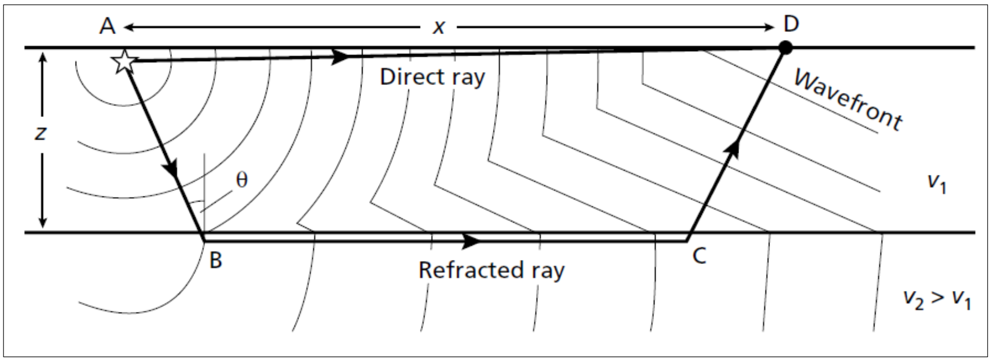

FIGURE 11: Individual ray paths from source A to detector D are drawn as solid lines (Kearey et al., 2002)

The travel time curve can be used to compute the velocities to the different refractors as illustrated in Figure 12.

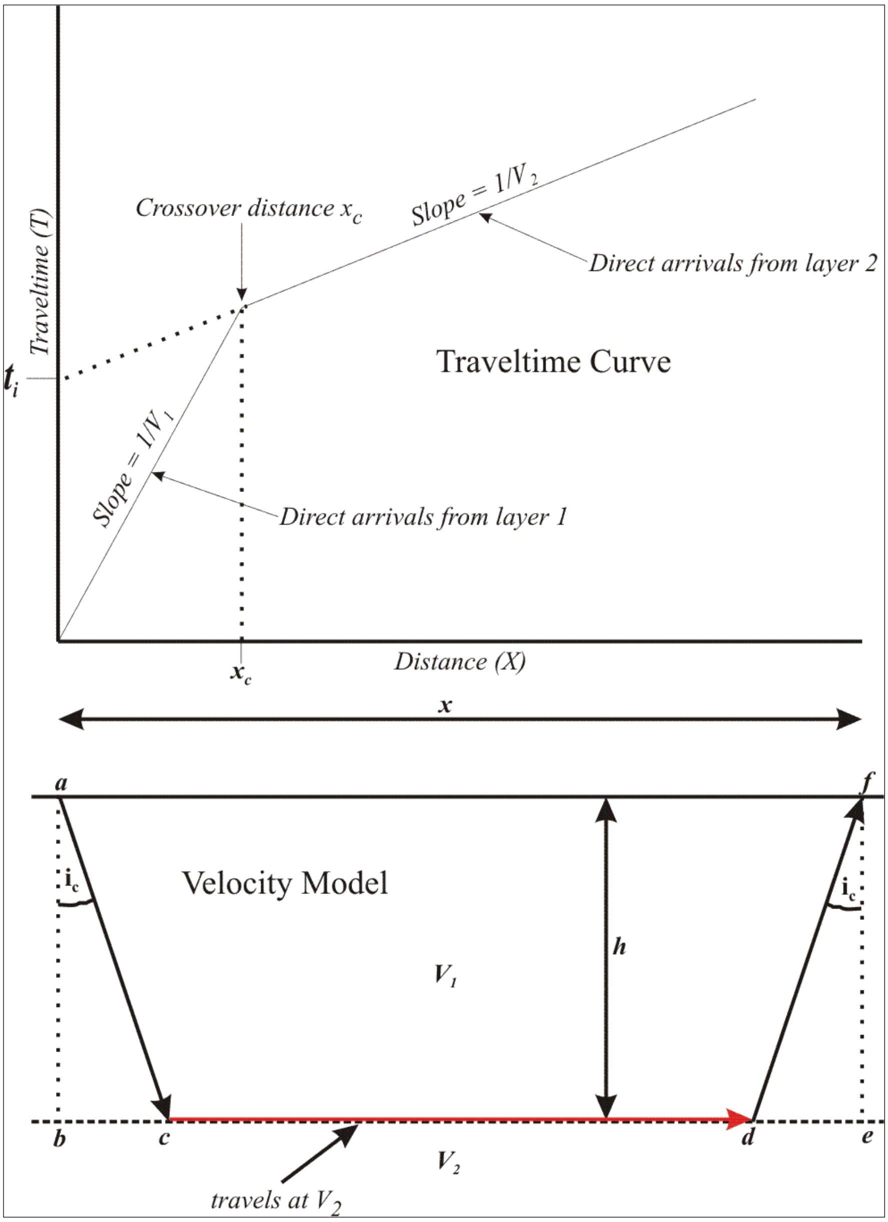

FIGURE 12: Travel time curve and velocity model

With the velocity model, the depth to the different layers (Equation 15) is also computed (Kearey et al., 2002; Shearer, 2009; Stein and Wysession, 2003; Telford et al., 1990).

$$h = \frac{T_{2i}V_1}{2\cos\left(sin^{-1}\frac{V_1}{V_2}\right)} \tag{15}$$

The simplified two layer medium gives general approximation for determining the seismic velocities and the depth to the second layer. This forms a practical introduction to more advance and complex geological conditions with much work illustrated in literature (Kearey et al., 2002; Milsom, 2003; Shearer, 2009; Stein and Wysession, 2003; Telford et al., 1990).

Refraction seismology is applied to a very wide range of scientific and technical problems, from engineering site investigation surveys to large-scale experiments designed to study the structure of the entire crust or lithosphere (Kearey et al., 2002). Where, the results will reveal:

a) Velocity structure used to infer rock type. Different rock types have a characteristic velocity also used to their densities.
b) Depth to interfaces due to changes in the lithology.

In geothermal exploration, seismic refraction has been applied in the various fields to determine the velocity structure/ geological/lithological models of the prospect areas. Where, the location of the main parameter of a geothermal system are mapped. In Öxarfjördur area in NE-Iceland (Georgsson et al., 2000), geothermal activity is found in sedimentary surroundings, refraction and gravity measurements where jointly interpreted to show the extent to formations layers (Figure 13) and inferring the subsurface lithological structure .

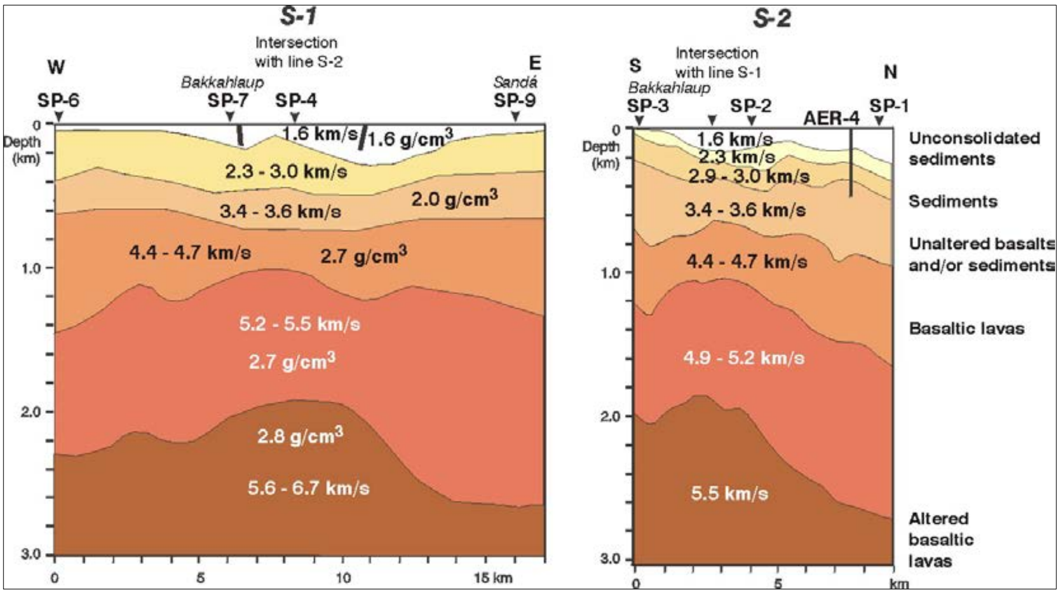

FIGURE 13: Refraction and gravity measurements of Öxarfjördur Area in NE-Iceland

4. SEISMIC MONITORING

Geothermal prospect areas are generally located in regions of high volcanic/tectonic activities which are associated with elevated earthquake activities of varying magnitudes. Where, seismicity of geothermal areas is often higher than the surrounding areas (Foulger, 1982; Ward, 1969). Thus, seismic monitoring has been established as a useful tool in exploration whose focus, is to study earthquake distribution and wave properties (Foulger, 1982; Simiyu, 2000; Simiyu and Malin, 2000).

Seismic monitoring (Foulger, 1982; Onacha et al., 2005, 2007, 2009, 2010; Simiyu, 2000; Simiyu and Malin, 2000; Ward, 1969) has aided in acquisition of earthquake data within a geothermal field which is a useful tool for:

a) Mapping the location of heat sources

The crust's strength increases linearly with increasing pressure but decreases exponentially with increasing temperature, with the peak strength occurring at the boundary zone between the brittle ductile zone (Simiyu and Malin, 2000). This transition zone is characterised by a high brittle deformation resulting in earthquakes. Mapping of this spatial seismic intensity, hypocenter distribution, shear wave attenuation, and wave reflection aid in mapping the location of the heat sources in geothermal areas (Figure 14).

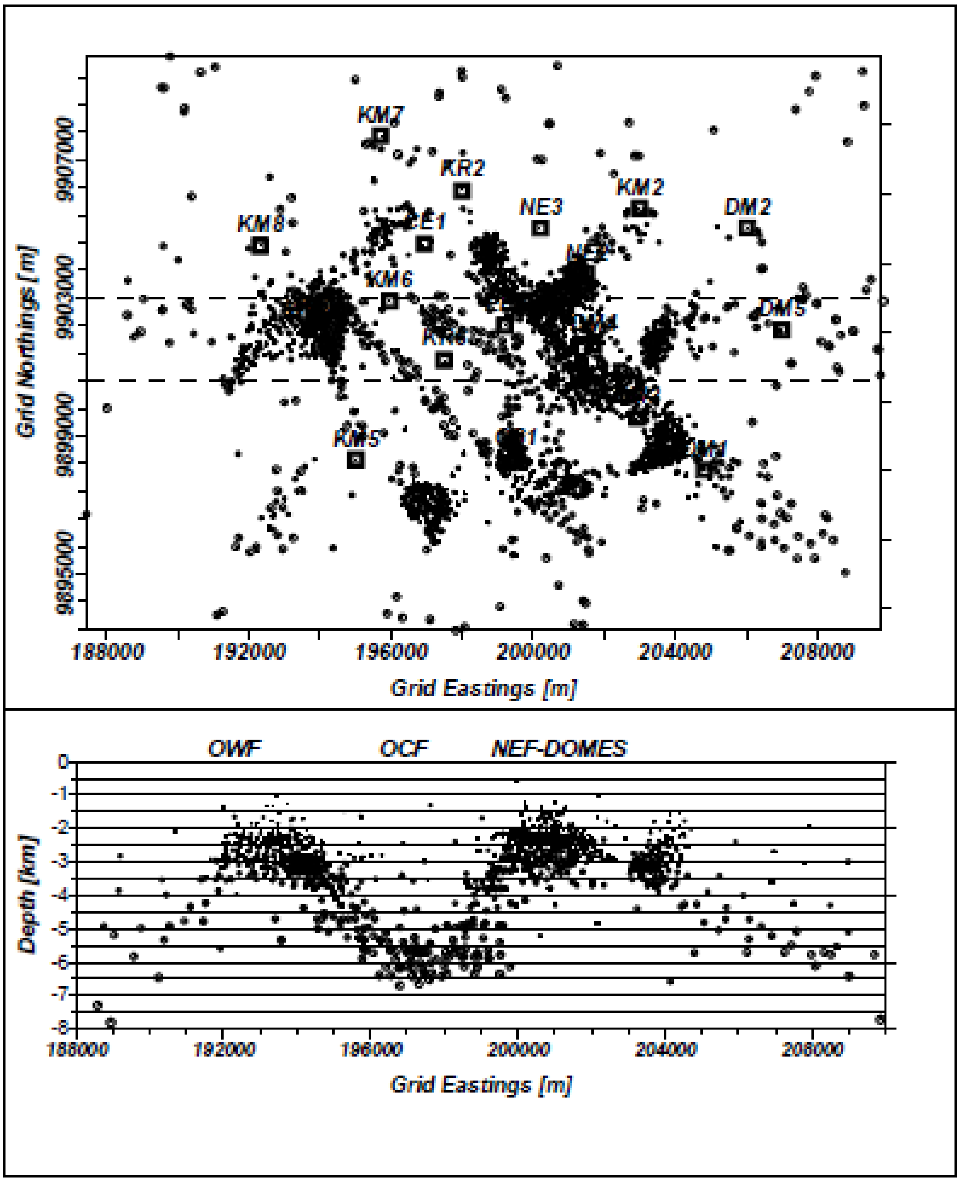

FIGURE 14: Earthquake location map and cross-section of Olkaria Geothermal Field (Simiyu and Malin, 2000)

b) Mapping high fracture density zones
High fracture density zones are paths of fluid movement and permeability in a geothermal area. This zones are interpreted from shear wave splitting analysis and crack density inversion results (Onacha et al., 2009, 2010; Simiyu and Malin, 2000). This are high potential zones for productive geothermal wells.

c) Determining the fluid phase, reservoir size and characteristics
Phase separation in a geothermal system is a function of temperature, pressure, fluid saturation, and the composition of advancing fluids. Phase separation produces different classes of reservoirs such as fluid dominated or vapour dominated systems (Simiyu and Malin, 2000). Measuring the variation in seismic velocity (Vp/Vs ratio) within the field, which is a function of water saturation (Figure 15). This is important in delineating geological structures with contrasting physical properties within the reservoir (Foulger, 1982; Simiyu and Malin; 2000).

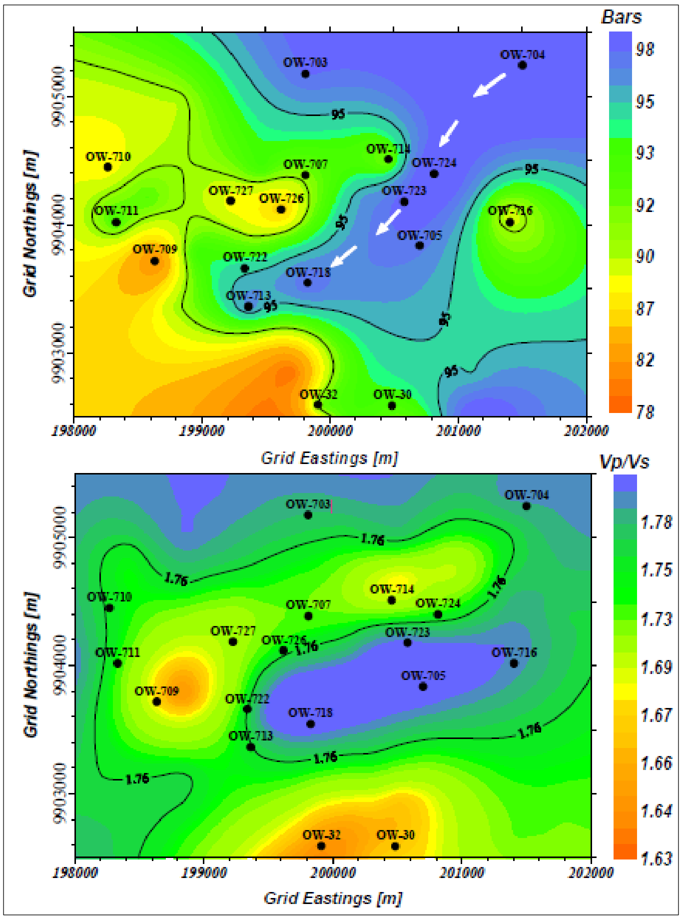

FIGURE 15: Map of the Olkaria North East area showing a comparison of Vp/Vs against reservoir pressure at 500 masl. The white arrows show the direction of water flow into the field (Simiyu and Malin, 2000)

5. CONCLUSION

Understanding of seismic methods and their applications aid in enriching our choice employing the most efficient geophysical techniques for geothermal resource exploration and monitoring. It has been noted, seismic methods are an instrumental geophysical techniques in evaluating the geothermal resource potential of a geothermal prospect. Where, the techniques have been found to be useful in mapping of the geological and lithological structure from interpreted velocity models, and evaluating the geothermal parameter including location of heat source, suitable reservoir, and permeability therby targeting potentially productive wells.

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