

The Application of the Shallow Seismic Reflection Method and AVO Analysis to Identify the Water Table Reflection

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By

Mahmud Mustain

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وَمَا أُوتِيتُمْ مِّنَ الْعِلْمِ إِلَّا قَلِيلًا

.... And of knowledge, you (mankind) have been given only a little.

(Al-Qur'an : Al-Isro' : 85)

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By,
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Abstract

A simple mathematical model of a sandstone aquifer has been constructed based on a local example, the Sherwood Sandstone of the East Midlands, UK. Simple seismic reflectivity calculations show that the air-water interface should theoretically produce a detectable seismic reflected wave for sandstone porosities as low as 10%. A synthetic seismic reflection dataset was constructed for a typical field survey geometry, and processed using the Promax system to produce a stacked section. The final section clearly shows the water table reflector. A field dataset from a subsequent survey has also been processed using the same sequence which also imaged a clear reflector at 30m depth. This is important evidence that the method has uses in identifying water table as a part of progress in shallow seismic reflection survey. The methods currently employed are (1) to define the optimum field, and (2) to define the optimum processing sequence, so that water table reflection can be imaged in a variety of geological situations.

The application of Amplitude versus Offset (AVO) analysis to CMP gathers from the field data shows a characteristic increase of amplitude with increasing angle of incidence for super-critical reflection. In this way the water table reflector is clearly identified with the amplitude increasing by 30% over the range of incident angle from 28° to 34°. AVO analysis has also been applied to other field data that has a similar geological setting, but with a lithological reflector over the same super-critical angle. The resulting AVO curve shows a decrease in amplitude of over 90% with increasing offset, clearly differentiating from the water table reflection. Both water table and lithological results closely agree with theoretical predictions.

The results of the field survey, and its interpretation, are supported by electrical resistivity soundings, seismic P and S-wave refraction studies, and nearby borehole evidence. This research will also be useful for hydrogeological investigation.

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In the name of Allah SWT, price is to Allah SWT. May peach and blessing be upon the Seal of the Prophets, specially Muhammad SAW, his family, all his companions, and his followers, amin.

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Finally, I would like to repeat the do'a of Prophet Musa AS., who said : “ O my Lord ! Expand me my breast. Easy my task for me, and remove the impediment from my speech, so they may understand what I say”. (Al-Qur'an, Toha: 25-28) Amin.

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Chapter 1

INTRODUCTION

1.1. Background

Generally, developments in seismology applications are categorised into two systems known as “deep survey for oil exploration” and “shallow surveys for mineral mining and groundwater investigation”. The former are very effective in developing deep surveys for hydrocarbon exploration, while shallow survey, although effective is rarely applied nowadays.

In recent years, improvements in seismographs, field methods and processing techniques have increased the signal to noise ratio and resolution of shallow seismic reflection surveys (Brabham & McDonald, 1992; Bredewout & Gaulty, 1986). Many publications (e.g. McGuire & Iron, 1997; Woodward, 1994; Davies et al, 1992; Meekes and van Will, 1991; Meekes et al, 1990; Geissler, 1989; Dobecki & Romig, 1985; Goulty, 1983) have demonstrated the potential of the reflection method for groundwater study. By careful attention to detail in the field and the processing, their data could be interpreted in terms of aquifer systems. Differences in computed interval velocities can be used to estimate the properties of porous formations using standard formulae (Davies et al, 1992). Therefore, the correlation of seismo-stratigraphic and geological information may be used to derive a hydrogeological interpretation relating to the porosity of the rock. However, there are very few published works that refer to the estimation of hydrogeological parameters from seismic reflection data.

1.2. Purpose of Study

The general purpose of this research is to progress the development of the reflection method for identification of the water table. Clement et al (1997) states that “*Seismic refraction is the only technique to image the water table*”. Theoretically, the

acoustic impedance contrast of the water table in pure sandstone (typically 15 % porosity) is more than sufficient to produce a reflected signal at the boundary. Therefore, the reflection method may image the water table. It is assumed that a water table forms a well-defined planar boundary within a formation. This has a relevant reflection coefficient that varies with the porosity of the rock that should be >10 %.

A seismic model of a water table interface has been produced, and synthetic data from this model have been processed to show the resulting seismic section. The next objective is to relate the synthetic model to real field data, for both simple and complex geological structure. The final objective is to assess the contribution which seismic reflection methods can make to the assessment of saline incursion into aquifers.

There are two specific objectives in order to identify a water table. The first is to image the reflector using a reflection survey. For this, it will be necessary to firstly identify the critical weaknesses in acquisition and processing of data, and then to find the solution. The second is to confirm that the reflection is actually from the water table. The discrimination between dry and water saturated rock will be validated by formulation of the correlation between the porosity of rock and its reflection coefficient. Resistivity surveys and/or seismic refraction surveys may be used to confirm this study.

The presence of a water table may potentially be determined using seismic surveys by either of the following methods;

1. S-wave reflection or refraction for comparison with P waves
2. P-wave AVO analysis.

The feasibility of using both these methods will be investigated with synthetic models, then as appropriate, with field trials.

1.3. Methodology

This research concentrates on field acquisition techniques and data processing. The methodology of this research is to carefully identify the pitfalls or parameter

ambiguities in the field and the subsequent data processing, then to find ways of solving them. The final result of these improvements will be confirmed by comparison with other geophysical methods e.g. refraction and electrical resistivity survey.

The sequence of research in this study comprises a literature search, use of synthetic models, and collection and processing of field data. The results are presented in the following seven chapters.

Chapter 2

This chapter describes the basic theory of seismic waves and the seismic reflection method. The application of the method to shallow target and vertical-horizontal resolution is discussed.

Chapter 3

This chapter reviews the hydrogeology of the water table and seawater intrusion. This chapter also discusses the field areas where the research will be applied. This includes the general description of geological background of the field; simple structure, complex structure, and coastal vicinity.

Chapter 4

This chapter analyses the use of synthetic seismic models on two computer test applications. The first being travel-time curve, the second being synthetic traces. Based on the same physical parameters to the travel-time curve within the spreadsheet, the synthetic traces have been created using a convolution principle on Promax.

Chapter 5

This chapter describes a practice application of the shallow seismic reflection method as a case history, then uses other geophysical methods to confirm the result. It includes; seismic refraction, resistivity, and S-wave refraction survey. It also covers the processing aspect of the seismic reflection system and technique used to improve data quality for both data acquisition and processing.

Chapter 6

Here, the Amplitude Variation with Offset (AVO) analysis is described. AVO analysis is then applied to two examples, one on isolated water table reflection (Edwinstowe) and the other over layered lithological reflectors, to illustrate the contrast in AVO response.

Chapter 7

This presents the conclusions and recommendation for future work.

Chapter 2

APPLIED SEISMOLOGY: BASIC THEORY AND PRINCIPLE OF SHALLOW SEISMIC REFLECTION

2.1. Seismic Wave

According to the definition, seismic wave is an elastic disturbance which is propagated from point to point through a medium (Sheriff, 1984). The propagation will change at certain places within different parts of the medium, travelled by the wave. The quantities of these changes depend principally on the energy content of the wave and on the physical properties of the medium itself.

2.1.1. Theory of Elasticity

Elastic deformation is a non-permanent deformation that occurs if a body returns to its original shape when the applied stress is released. The principal types of changes due to these deformations are re-distribution of the internal force and modification of the geometrical shape (Al-Sadi, 1980). The theory of elasticity is concerned with the analysis of these two principal effects and their related features.

2.1.1.1. Stress

In the broad definition, stress is represented by forces which act on a finite area occupying an arbitrary position within the medium (Al-Sadi, 1980; Grant & West, 1965). However, in a more practical definition, it is the limiting value of a force acting on an elementary area that is near to zero. In the mathematical formulation, it is given by:

$$P = \lim_{\Delta A \rightarrow 0} \Delta F / \Delta A \quad \text{then} \quad dF = P dA \quad 2.1$$

where dF is a delta force and dA is the elementary area. The stress tensor has nine components those are P_{nm} where $n=m=x,y,z$. It is represented by n that stands for the set area and m for the component direction. Figure 2.1 illustrates all the components of stress.

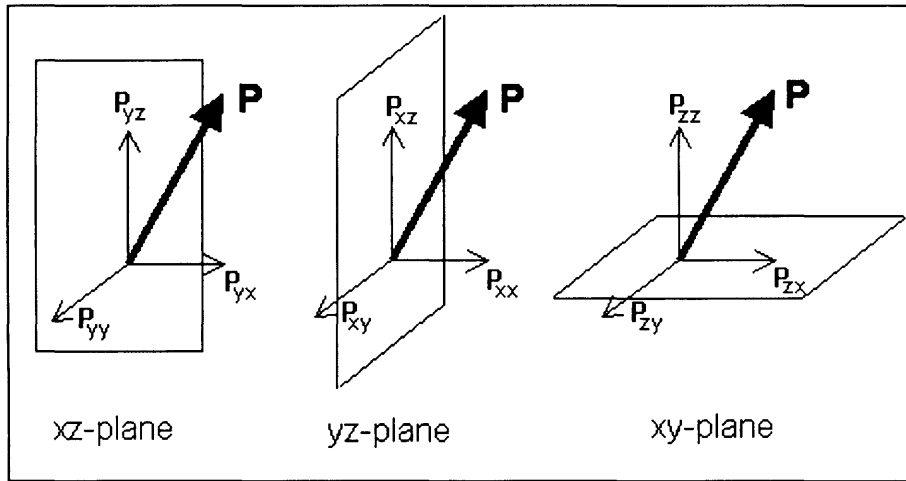


Figure 2.1 The stress tensor

For the small volume of ΔV near to 0 then dP also near to 0 and $P_{ab} = P_{ba}$. Consequently these properties result in reducing the total number of independent component to six (P_{xx} , P_{xy} , P_{xz} , P_{yy} , P_{yz} , and P_{zz}).

2.1.1.2. Strain

Strain is defined as the change of dimensions or shape produced by stress (Sheriff, 1983). If U is defined as stress-produced displacement and has three

components of x , y , and z therefore in Cartesian tensor, this displacement has two components; symmetry or dilatational (e_{ik}) and antisymmetry or rotational (ξ_{ik}).

$$\left. \begin{array}{l} \text{Symmetry } e_{ik} = \frac{1}{2} \left(\frac{\partial U_k}{\partial x_i} + \frac{\partial U_i}{\partial x_k} \right) \\ \text{Antisymmetry } \xi_{ik} = \left(\frac{\partial U_k}{\partial x_i} - \frac{\partial U_i}{\partial x_k} \right) \end{array} \right| \quad \frac{\partial U_i}{\partial x_k} = e_{ik} - \frac{1}{2} \xi_{ik} \quad 2.2$$

Where; $i=1,2,3$ and $k=1,2,3$ for Cartesian co-ordinate $1,2,3 = x,y,z$

There is a special condition for an antisymmetry component that has the properties :

$$\xi_{ik} = 0 \quad \text{if } i=k \quad \text{and} \quad \xi_{ik} = \xi_{ki} \quad \text{if } i \neq k$$

Therefore there are only three independent components: ξ_{yz} , ξ_{zx} , and ξ_{xy} .

2.1.1.3. Stress-Strain Relation

Bullen, 1965 (Al-Sadi 1980), gives the relationship :

$$P_{xx} = \lambda(e_{xx} + e_{yy} + e_{zz}) + 2\mu e_{xx}$$

$$P_{yy} = \lambda(e_{xx} + e_{yy} + e_{zz}) + 2\mu e_{yy}$$

$$P_{zz} = \lambda(e_{xx} + e_{yy} + e_{zz}) + 2\mu e_{zz}$$

$$P_{xy} = 2\mu e_{xy}$$

$$P_{xz} = 2\mu e_{xz}$$

$$P_{yz} = 2\mu e_{yz}$$

where λ and μ are called Lamé's constants. In the general formulation :

$$P_{ij} = \lambda \theta \delta_{ij} + 2\mu e_{ij} \quad 2.3$$

Where; θ = cubic dilatation

$$\delta_{ij} = 1 \text{ for } i = j$$

$$= 0 \text{ for } i \neq j$$

where i and j take the values x , y , and z

2.1.1. 4. Elastic Constant

The elastic constant (also commonly called elastic modulus), represents the specific characteristics of elastic response of a particular medium. The most common ones are the following :

1. Young's Modulus (Y) $P_{xx} = Y \cdot e_{xx}$

2. Bulk Modulus (B)

$P_h = -B \cdot \theta$ where P_h is the hydrostatic pressure and θ is the fractional change as cubical dilatation

3. Rigidity or Shear Modulus (μ)

$P_{xy} = \mu \cdot \phi$ where ϕ is an angle shear, $\mu = 0$ for liquid media

4. Poisson Ratio (σ), $\sigma = \frac{\Delta d / d}{\Delta l / l}$; where l is length and d thickness

5. Lamé's Modulus (λ) $P_{xx} = \lambda \cdot e_{zz}$

The most common relations are :

$$Y = \frac{\mu(3\lambda + 2\mu)}{(\lambda + \mu)} = \frac{9B\mu}{3B + \mu} \quad ; \quad B = \frac{3\lambda + 2\mu}{3} = \frac{Y}{3(1 - 2\sigma)}$$

$$\mu = \frac{2B - \lambda}{2} = \frac{Y}{2(1 + \sigma)} \quad ; \quad \sigma = \frac{\lambda}{2(\lambda + \mu)} = \frac{2B - 2\mu}{6B + 2\mu}$$

$$\lambda = \frac{3B - 2\mu}{3} = \frac{\sigma Y}{(1 + \sigma)(1 - 2\sigma)}$$

2.1.2. Wave Equation

2.1.2.1. Equation of Motion

If a time dependent stress is caused by a wave travelling through a medium that has a Cartesian co-ordinate system then we can write the displacement at the point $P(x,y,z)$ at any instant t as vector $U(x,y,z,t)$. There are three pairs of opposite faces in one component in every direction x , y , or z . For example, the surface force acting on the small volume ΔV in z direction and perpendicular to xy plane for face of $\Delta x \Delta z$ (see figure 2.2) is given by:

$$\frac{\partial P_{yz}}{\partial y} \Delta V = (P_{yz} + \frac{\partial P_{yz}}{\partial y} \Delta y) \Delta z \Delta x - (P_{yz} - \frac{\partial P_{yz}}{\partial y} \Delta y) \Delta z \Delta x \quad 2.4$$

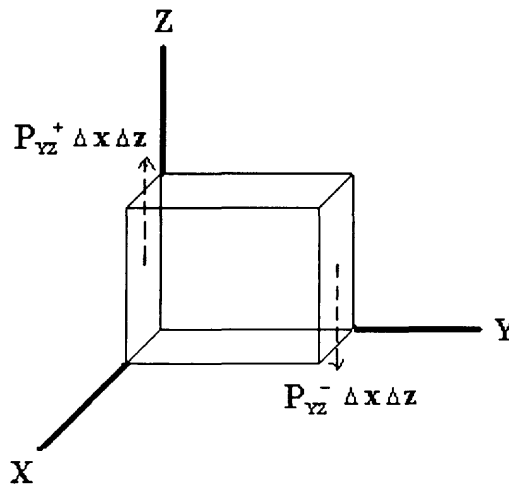


Figure 2.2 a pair of surface forces from the face of $\Delta x \Delta z$ in the z direction

Similar expressions for two other Cartesian coordinates can be obtained. Then we can write the net of the three components:

$$F_z = \left(\frac{\partial P_{xz}}{\partial x} + \frac{\partial P_{yz}}{\partial y} + \frac{\partial P_{zz}}{\partial z} \right) \Delta V \quad 2.5$$

In general term :

$$F_i = \sum_{k=1}^3 \frac{\partial P_{ki}}{\partial x_k} \Delta V = \sum_{k=1}^3 \frac{\partial P_{ik}}{\partial x_k} \Delta V \quad 2.6$$

Therefore the motion of the material at P, neglecting the gravitational force, is given by :

$$\rho \frac{\partial U_i^2}{\partial t^2} = \sum_{k=1}^3 \frac{\partial P_{ik}}{\partial x_k} \Delta V \quad 2.7$$

where :

ρ : density of the medium

d/dt : Lagrangian diff operator

(It shows the total motion of a particular element in the medium)

$\partial/\partial t$: Eulerian operator or partial differentiation

(shows the differentiation of a point fixed in the medium)

The relationship between them is :

$$\frac{d}{dt} = \frac{\partial}{\partial t} + \sum_{i=1}^3 \frac{dU_i}{dt} \frac{\partial}{\partial x_i} \quad 2.8$$

2.1.2.2. Type of Seismic Wave

From the analysis of stress and strain it is apparent that strain generates two deformations, translational and rotational. These deformations become the cause of the

dilatations or longitudinal waves (P) and transverse or shear waves (S) respectively. These are referred to as body waves. On the other hand when the wave travels along a free surface then it is called a surface wave.

1. Longitudinal Wave (P)

The main characteristic of this wave is that the direction of particle motion is parallel to the direction of wave propagation. It is also called compression or primary (P) wave. If the medium is perfectly elastic, homogenous, and isotropic (having the same physical properties in all directions) we can use the formulation of stress-strain relation, giving :

$$\rho \frac{\partial^2 U_i}{\partial t^2} = (\lambda + \mu) \frac{\partial \theta}{\partial x_i} + \mu \nabla^2 U_i \quad 2.9$$

that is,
$$\rho \frac{\partial^2 \theta}{\partial t^2} = (\lambda + 2\mu) \nabla^2 \theta \quad \text{or} \quad \frac{1}{\alpha^2} \frac{\partial^2 \theta}{\partial t^2} = \nabla^2 \theta \quad 2.10$$

where $\alpha^2 = (\lambda + 2\mu) / \rho$, $\alpha = V_p$ as velocity of P-wave. This is the fastest seismic wave,

$$V_p = \sqrt{(\lambda + 2\mu) / \rho} \quad 2.11$$

2. Transverse Wave (S)

This is also called Shear or Secondary wave. The particles move in the direction that is perpendicular to the wave propagation. Therefore, there are two components vertical (SV) and horizontal (SH). By analogy to the equation (2.9) we can write the equation for V and W as y and z respectively:

$$\rho \frac{\partial^2 V_i}{\partial t^2} = (\lambda + \mu) \frac{\partial \theta}{\partial x_i} + \mu \nabla^2 V_i \quad 2.12$$

$$\rho \frac{\partial^2 W_i}{\partial t^2} = (\lambda + \mu) \frac{\partial \theta}{\partial x_i} + \mu \nabla^2 W_i \quad 2.13$$

By subtracting the derivative of equation (2.12) with respect to z from the derivative of equation (2.13) with respect to y we get (see Sheriff and Geldart, 1995 p-40) :

$$\rho \frac{\partial}{\partial t^2} \left(\frac{\partial W}{\partial y} - \frac{\partial V}{\partial z} \right) = \mu \nabla^2 \left(\frac{\partial W}{\partial y} - \frac{\partial V}{\partial z} \right) \quad 2.14$$

that is,
$$\frac{1}{\beta^2} \frac{\partial^2 \theta_x}{\partial t^2} = \nabla^2 \theta_x \quad \text{where } \beta^2 = \mu/\rho \quad 2.15$$

3. Surface Wave

There are two types of surface wave, Rayleigh (R) and Love (L) waves. The particle motion of R-waves is in a vertical plane, which is parallel to the direction of propagation and has a retrograde elliptical orbit. The simple formulation of this wave is given : $V_R = 0.92 V_S$ where V_S is the S-wave velocity in the same medium (Al-Sadi, 1980).

On the other hand, the particle motion of L-waves is transverse and in a horizontal plane. Therefore, this wave can not be detected by any vertical component of the receiver. The velocity of L-waves approaches the S-wave velocity.

4. Seismic Noise

In seismology, noise is commonly defined as an unexpected signal. Practically, it is divided into two types; coherent and random noise. Coherent noises are certain signal trains that bear a systematic phase relation (coherent) between adjacent traces (Sheriff, 1983). Most source-generated seismic noise is coherent, such as ground-roll, shallow refraction, air-wave, and multiple events. Random noise, however, consists of unpredicted signals. The main characteristic of this noise is that both the amplitude and onset are naturally random. That is why it is called random or incoherent noise. It is also called natural ambient noise or microseisms because the source is natural such as wind and sea waves, and other various man-made disturbances.

2.1.2.3. Travel Time

The basis of the seismic method is the measurement of the travel time of seismic waves that have either travelled direct, or been refracted or reflected at subsurface boundaries. From measurement of travel time it is usually possible to analyse the depths and dips of horizons of reflectors and the velocities of the seismic waves. Due to different wave, there are three kinds of travel time for seismic waves; direct, reflection, or refraction waves. For a single horizontal boundary, travel time (t) (see figure 2.3) can be expressed by :

1. The direct wave, $t = x/v_0$ 2.16

2. The reflected wave, $t = \frac{\sqrt{x^2 + 4h^2}}{v_0}$ 2.17

3. The refracted wave, $t = 2h\sqrt{\frac{1}{v_0^2} - \frac{1}{v_1^2}} + \frac{x}{v_1}$ 2.18

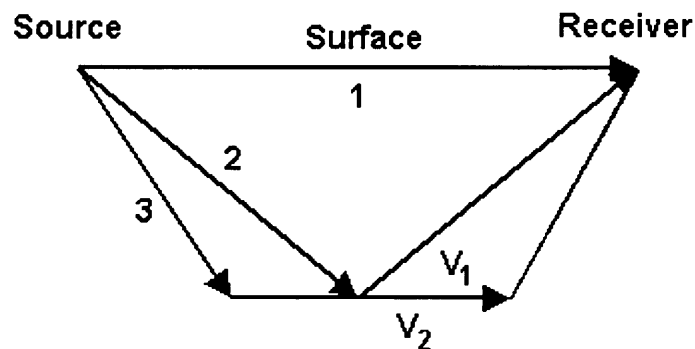


Figure 2.3 Travel time; 1 the direct wave, 2 the reflected wave, 3 the refracted wave

The travel time of the direct wave travelling a certain distance x across the surface, depends only on the velocity of the wave v_0 . On a time-distance graph the direct wave is represented by a straight line through the origin which has a gradient of $1/v_0$ (see figure 2.2).

The reflected wave is represented by a hyperbola on the t - x graph. It intersects the t -axis at t_0 as $2h/v_0$ and approaches $t=x/v$ asymptotically for large offsets. Therefore a reflected wave is always a late arrival. The increase of travel time depends only on x if we assume that seismic waves travel in a homogenous medium and t_0 is constant. This point is very important as a basic concept of normal moveout correction where $t-t_0 = dt$ is the step-up (move-out) time.

For multiple reflection events, where we have n reflectors, still concerning the reflection wave, then we have n travel time curves. For any angle of incidence that is sufficiently small, the travel time of reflection wave in simple horizontal reflectors can be approximated by:

$$t_n = \sum_{i=1}^n 2h_i / v_i \quad 2.19$$

where n is the number of layers and each layer is identified by velocity v and thickness h .

The head-wave only occurs where $v_I > v_0$ and the incident wave reaches the boundary at the critical angle θ_c for that : $\sin \theta_c = v_0/v_I$. The travel time equation shows that the wave essentially propagates at the velocity of the faster second layer but it is delayed as it travels along the inclined parts of the path. A head wave is represented on the graph by a straight line, which intersects the t -axis at a time t_1 :

$$t_1 = 2h \sqrt{\frac{1}{v_0^2} - \frac{1}{v_1^2}} \quad 2.20$$

and has a gradient of $1/v_I$ (see figure 2.4). The refracted arrival can be observed at a distance beyond the angle of critical refraction θ_c . This distance is commonly called the critical range x_c and given :

$$x_c = 2 h \tan \theta_c. \quad 2.21$$

For n layers, Dix 1955 (Sheriff and Geldart, 1995; Amery, 1993), give the formula for the reflected traveltime :

$$t^2 = \frac{x^2}{V_{rms}^2} + t_0^2 \quad 2.22$$

$$t^2 = \frac{x^2}{V_{rms}^2} + 4h^2 / V_{rms}^2$$

$$\text{and } V_{rms}^2 = \frac{\sum_{i=1}^n V_i^2 \Delta t_i}{\sum_{i=1}^n \Delta t_i} \quad 2.23$$

where t_0 is two-way vertical time, V_i is interval velocity Δt_i the interval time.

2.1.3. Boundary Layer

2.1.3.1. Wave equation in boundary layer

According to the Helmholtz method (Grant & West, 1965), the displacement U_i has two displacement potentials; dilatational, (ϕ) , as arbitrary scalar and rotational, (ψ) , as arbitrary vector. It is given by :

$$U_i = \nabla \phi - \nabla \times \psi \quad 2.24$$

If we have additional condition of $\nabla \cdot \psi = 0$, then div. and curl operators can be used to simplify the term to become :

$$\nabla \cdot U_i = \nabla^2 \phi \quad \text{and} \quad \nabla \times U_i = \nabla^2 \psi \quad 2.25$$

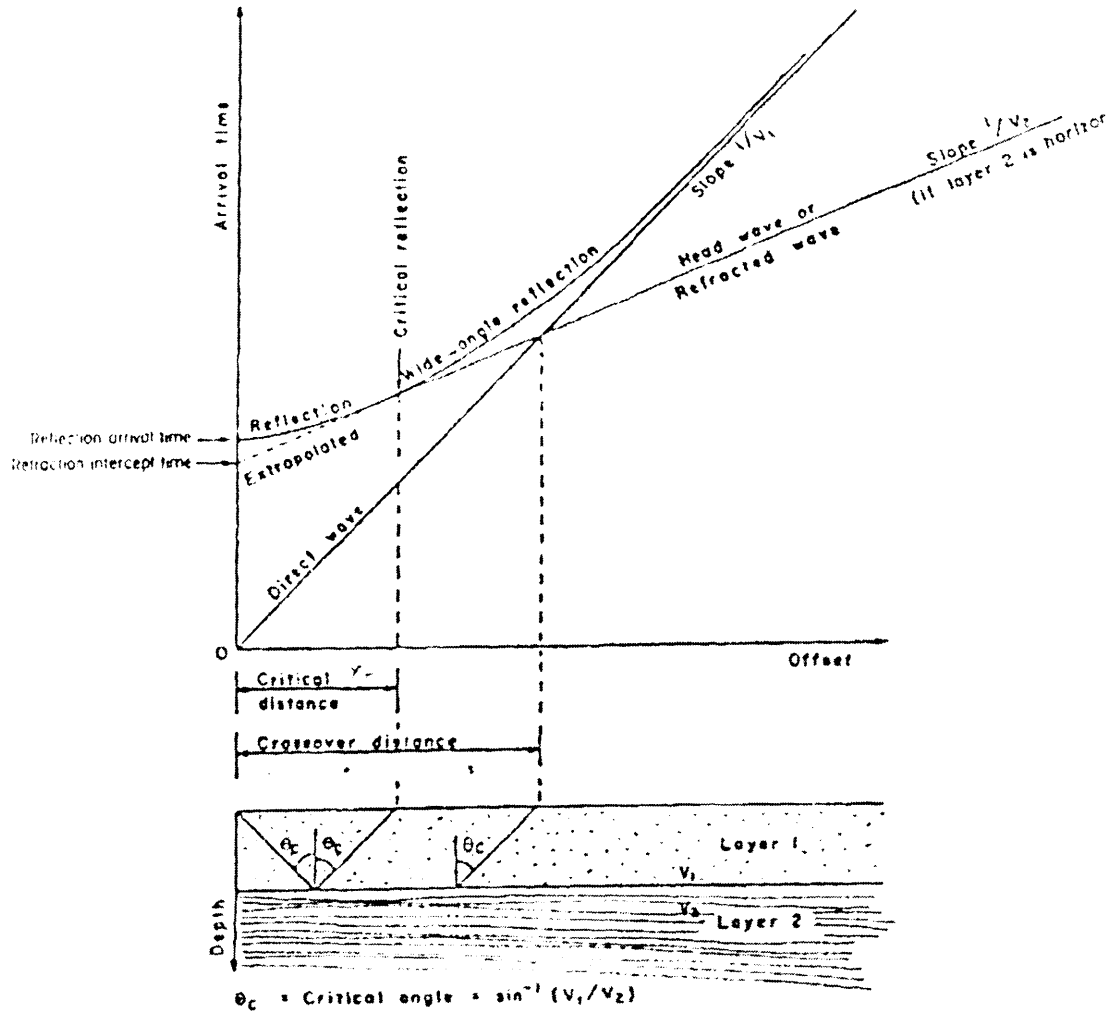


Figure 2.4 Relation between traveltime curves of reflected, critical refracted, and direct wave. The distance from 0 to x_c is a critical distance for the refracted wave (From Sheriff, 1984).

Let us consider the propagation of a plane wave that is characterised by those displacement potentials. When the material is homogeneous and isotropic (having the same physical properties regardless of the direction in which they are measured, Sheriff 1984), the displacement potentials Φ & Ψ and the displacement component v can be written respectively as :

$$\frac{\partial^2 \psi}{\partial t^2} = \beta^2 \nabla^2 \psi, \quad \frac{\partial^2 \phi}{\partial t^2} = \alpha^2 \nabla^2 \phi, \text{ and } \frac{\partial^2 v}{\partial t^2} = \beta^2 \nabla^2 v \quad 2.26$$

Now we start to write in their spectrum form ;

$$\begin{aligned} \nabla^2 \Phi + k_\alpha^2 \Phi &= 0 \\ \nabla^2 \Psi + k_\beta^2 \Psi &= 0 \\ \nabla^2 V + k_\beta^2 V &= 0 \end{aligned} \quad 2.27$$

where Φ , Ψ , and V are Fourier transform. The general solution of these equations that correspond to plane wavefronts travelling in the direction $(l, 0, n)$ are given by:

$$\Phi(x, z) = A(\omega) \exp. i. k_\alpha (lx - nz) \quad 2.28$$

$$\Psi(x, z) = B(\omega) \exp. i. k_\beta (lx - nz) \quad 2.29$$

$$V(x, z) = C(\omega) \exp. i. k_\beta (lx - nz) \quad 2.30$$

Where; $B = C = 0 \rightarrow P$ wave

$A = C = 0 \rightarrow SV$ wave

$A = B = 0 \rightarrow SH$ wave

The minus sign shows that the direction is downward. When a plane wave is incident on a plane boundary layer as a reflector separating two elastic media, then both reflected and transmitted plane waves will propagate away from the boundary. If the direction of the displacement in the incident wavefront is oblique to the interface, shearing and compressive stresses will occur on account of the discontinuity in the elastic properties (Grant & West, 1965). Therefore both transmitted and reflected stress fields will contain both P and S components.

When the media are assumed to be uniform and the wavefronts planar, we can represent the progress of the wavefronts as ray paths. Moreover, if the angle between the incident wavefront and the interface is φ , then φ will also be the angle of incidence of the incident ray path. Therefore for the incident wave (in the plane of x - z) $l = \sin \varphi$ and $n = \cos \varphi$. Figure 2.5 shows the reflection and transmission wave path and plane wave (x, z) that propagate away from the boundary.

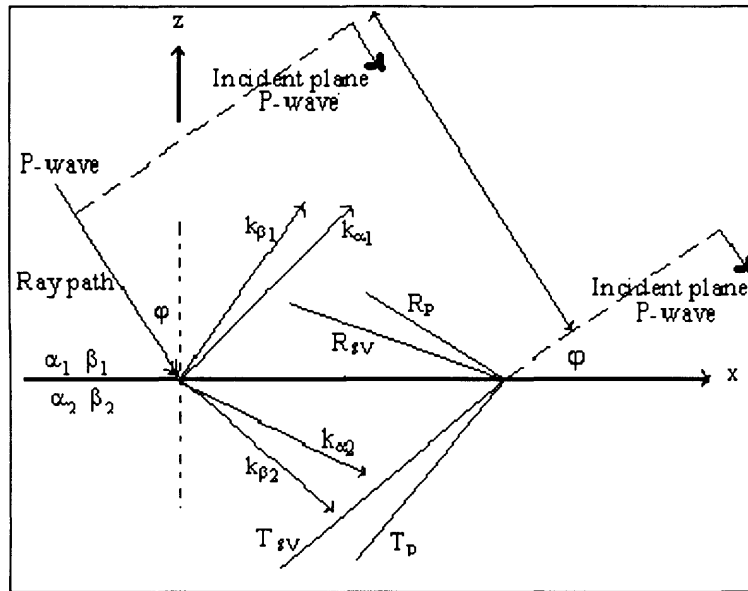


Figure 2.5 Reflection and transmission of P-wave at a boundary for ray path and incident plane (R_p : reflected P-wave, R_{sv} : reflected SV-wave, T_p : transmitted P-wave, and T_{sv} : transmitted SV-wave) (after Grant and West, 1965).

For the case of P waves that incident at angle φ_p in a certain medium, then $B = C = 0$ in that medium. In this case we can write the wave equation of the incident disturbance as;

$$\Phi_i(x, z) = A_1(w) \exp. i.k_{\alpha 1} (l_{\alpha 1}x - n_{\alpha 1}z) \quad 2.31$$

where wave number $k_{\alpha 1} = \omega/\alpha_1$; $l_{\alpha 1} = \sin \varphi_p$; and $n_{\alpha 1} = \cos \varphi_p$.

The reflected and transmitted displacement field due to that P wave are respectively given;

$$\Phi_r(x, z) = D_1(\omega) \exp. i.k_{\alpha 1}(l_{\alpha 1}x + n_{\alpha 1}z) \quad 2.32$$

$$\Psi_r(x, z) = E_1(\omega) \exp. i.k_{\beta 1}(l_{\beta 1}x + n_{\beta 1}z) \quad 2.33$$

$$V_r(x, z) = F_1(\omega) \exp. i.k_{\beta 1}(l_{\beta 1}x + n_{\beta 1}z) \quad 2.34$$

and

$$\Phi_t(x, z) = A_2(\omega) \exp. i.k_{\alpha 2}(l_{\alpha 2}x - n_{\alpha 2}z) \quad 2.35$$

$$\Psi_t(x, z) = B_2(\omega) \exp. i.k_{\beta 2}(l_{\beta 2}x - n_{\beta 2}z) \quad 2.36$$

$$V_t(x, z) = C_2(\omega) \exp. i.k_{\beta 2}(l_{\beta 2}x - n_{\beta 2}z) \quad 2.37$$

Now we have to determine six coefficients of these wave function; D_1, E_1, F_1, A_2, B_2 , and C_2 for solving them. Then we need six boundary conditions in the stress and displacement at the interface. The stress and displacement components are respectively given;

$$\begin{aligned} P_{zx} &= 2\mu e_{zx} = \mu(2\frac{\partial^2 \Phi}{\partial z \partial x} - \frac{\partial^2 \psi}{\partial x^2} + \frac{\partial^2 \psi}{\partial z^2}) \\ P_{zy} &= 2\mu e_{zy} = \mu \frac{\partial V}{\partial z} \end{aligned} \quad 2.38$$

$$\begin{aligned} P_{zz} &= \lambda \theta + 2\mu e_{zz} = \lambda \nabla^2 \Phi + 2\mu(\frac{\partial^2 \Phi}{\partial z^2} - \frac{\partial^2 \psi}{\partial z \partial x}) \\ \text{and } U &= \frac{\partial \Phi}{\partial x} + \frac{\partial \Psi}{\partial z}; V; W = \frac{\partial \Phi}{\partial x} - \frac{\partial \Psi}{\partial z} \end{aligned} \quad 2.39$$

Knott, 1899 (Sheriff and Geldart, 1995), solved the wave equation that wave energy is divided among reflected and transmitted (refracted) waves. Zoeppritz' equations (1919) give the amplitude of the reflected and refracted waves at a plane boundary layer for an incident P wave in terms of displacement while Knott's equation give the calculation in terms of potential (Sheriff and Geldart, 1995).

Knott's equations are given in terms of amplitudes of the displacement potential function Φ and Ψ (see Sheriff and Geldart, 1995, P75) :

$$-a_1 A_0 + a_1 A_1 - B_1 = -a_2 A_2 - B_2 \quad 2.40$$

$$A_0 + A_1 + b_1 B_1 = A_2 - b_2 B_2 \quad 2.41$$

$$\mu_1 c_1 A_0 + \mu_1 c_1 A_1 - 2\mu_1 b_1 B_1 = \mu_2 c_2 A_2 + 2\mu_2 c_2 B_2 \quad 2.42$$

$$-2\mu_1 a_1 A_0 + 2\mu_1 a_1 A_1 + \mu_1 B_1 = -2\mu_2 a_2 A_2 + \mu_2 c_2 B_2 \quad 2.43$$

where A_i and B_i in these equations are the amplitudes of the potential displacement functions Φ and Ψ , and not of the displacement. Whereas the Zoeppritz' equations are given :

$$(-A_0 + A_1) \cos \varphi_1 - B_1 \sin \delta_1 = -A_2 \cos \varphi_2 - B_2 \sin \delta_2 \quad 2.44$$

$$(A_0 + A_1) \sin \varphi_1 + B_1 \cos \delta_1 = A_2 \sin \varphi_2 - B_2 \cos \delta_2 \quad 2.45$$

$$(A_0 + A_1) Z_1 \cos 2\delta_1 - B_1 W_1 \sin 2\delta_1 = A_2 Z_2 \cos 2\delta_2 - B_2 W_2 \sin 2\delta_2 \quad 2.46$$

$$(-A_0 + A_1)(\beta_1 / \alpha_1) W_1 \sin 2\varphi_1 + B_1 W_1 \cos 2\delta_1 = -A_2 (\beta_2 / \alpha_2) W_2 \sin 2\varphi_2 - B_2 W_2 \cos 2\delta_2 \quad 2.47$$

where A_0 , A_1 , and A_2 are the displacement amplitudes of the incident, reflected, and refracted P-wave and B_1 & B_2 for the reflected and refracted S-waves, φ_1 incident angle, φ_2 refraction angle of P-wave, δ_1 reflection angle of S-wave, δ_2 refraction angle of S-wave, $Z_i = \rho_i \alpha_i$ and $W = \rho_i \beta_i$ are called acoustic impedance of P and S-wave respectively. This equation indicates that for a P-wave of given amplitude A_0 incident at the angle φ_1 on a plane boundary separating two media with given values of ρ , μ , α , and β , Snell's law determines the angles θ_i and δ_i , whereas Zoeppritz' equations fix the reflected and refracted amplitudes A_i and B_i (Sheriff and Geldart, 1995). Similar equations can be derived for an incident S-wave, a fluid medium $B_i = 0$ because only P-wave are propagated. Figure 2.5 shows waves generated at solid-solid boundary by incident P wave.

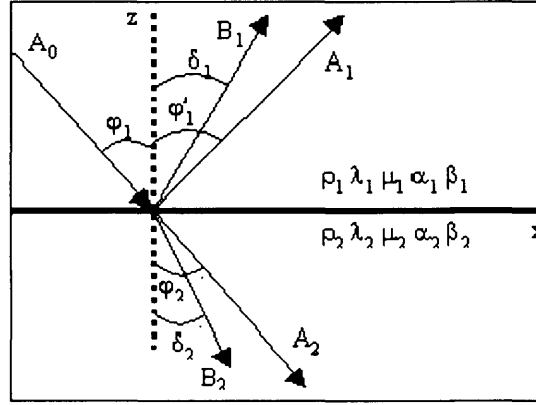


Figure 2.6 wave generated at solid-solid boundary by incident P wave.

Zoeppritz' equations reduce to a very simple form for normal incidence (assuming up to 15°). If a P-wave is at normal incidence then there are no tangential stresses and displacements, hence $B_1 = B_2 = 0$ (see Sheriff and Geldart, 1995 p-76). Therefore, the solution of the equations are found as the reflection coefficient R and transmission coefficient T :

$$R = \frac{A_1}{A_0} = \frac{\alpha_2 \rho_2 - \alpha_1 \rho_1}{\alpha_2 \rho_2 + \alpha_1 \rho_1} = \frac{Z_2 - Z_1}{Z_2 + Z_1} \approx \frac{\Delta Z}{2Z} \quad 2.48$$

$$R \approx \frac{1}{2} \Delta(\ln Z) \approx \frac{1}{2} (\Delta\alpha / \alpha + \Delta\rho / \rho)$$

$$T = \frac{A_2}{A_0} = \frac{2\alpha_1 \rho_1}{\alpha_2 \rho_2 + \alpha_1 \rho_1} = \frac{2Z_1}{Z_2 + Z_1} \quad 2.49$$

The fractions of energy reflected and transmitted are given by E_R and E_T respectively (assuming conservation of energy, $R+T=1$, as mention before):

$$E_R = \frac{\alpha_1 \rho_1 \omega^2 A_1^2}{\alpha_1 \rho_1 \omega^2 A_0^2} = \left(\frac{Z_2 - Z_1}{Z_2 + Z_1} \right)^2 = R^2 \quad 2.50$$

$$E_T = \frac{\alpha_2 \rho_2 \omega^2 A_2^2}{\alpha_1 \rho_1 \omega^2 A_0^2} = \frac{4Z_2 Z_1}{(Z_2 + Z_1)^2} = \frac{Z_2}{Z_1} T^2 \quad 2.51$$

These equations illustrate that reflected energy varies for impedance contrasts. For most of the interfaces encountered, only a small portion of energy is reflected because both density and velocity contrast are small (see table 2.1). The table shows the shallow interface and water table possibility by coefficient reflection (R_c).

Table 2.1 Energy reflected at interface between two media (After Sheriff and Geldart, 1995)

Interface	First Medium		Second Medium		R_c
	Velocity	Density	Velocity	Density	
Shallow Interface	2.1	2.4	2.3	2.4	0.045
Base of Weathering	0.5	1.5	2	2	0.68
Gas sand over water sand	2.2	1.8	2.5	2.3	0.18

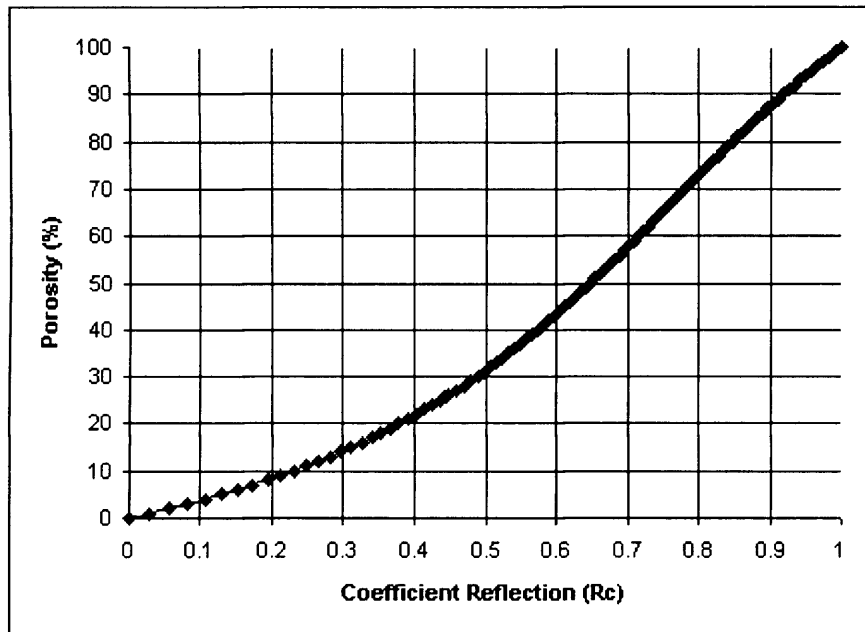


Figure 2.7 Coefficient Reflection (R_c) of seismic P-wave Vs Porosity (%) of Pure Sandstone
(The calculation is in the Appendix A)

In this specific case, the water table is expressed by a boundary of gas sand overlying a water sand. The porosity of the rock is not specified. The reflection coefficient of a water table interface in a given lithology is dependent on the porosity

(Bachrach & Nur, 1998; Jones et al, 1998; Sheriff & Geldart, 1983). Figure 2.7 shows the Porosity Vs Reflection coefficient (Rc) for pure sandstone. Typically, the porosity of aquiferous rock is about 15%. The Rc value of 0.18 that is mentioned in table 2.1 implies porosity of about 8%.

2.1.3.2. Amplitude Record of Seismic Wave

A seismic wave is a travelling signal wave-front that spreads out from its source to its receiver. The recorded amplitude of a seismic wave is very important in signal analysis. The amplitude depends mainly on the source type, wave type, the wave path, and distance from its source.

Regarding with wave type, the effect of geometrical spreading may cause variation of the wave amplitude. The intensity (I), spherical cap area (ϑ) or energy (E) of a wave at a distance (r) from the source (see figure 2.8) is given by :

$$\vartheta_1 \backslash \vartheta_2 = I_2/I_1 = E_2/E_1 = (r_1/r_2)^n \quad 2.52$$

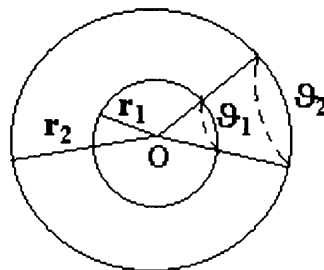


Figure 2.8 Geometrical spreading of wave
(after Sheriff and Geldart, 1995)

where the subscript 1 and 2 show the specific place in the medium, n is the constant and equals 2 for spherical waves. Spherical divergence causes the amplitude of both P and S-waves to decrease proportionally with distance. Whereas cylindrical divergence causes the amplitude of surface waves to decrease proportionally with the square root of the distance (Evans, 1997). For this reason Surface waves are always larger than body waves if it is assumed that the source is equally efficient for all wave types.

Concerning with wave path, energy of the wave is lost during travelling because of its transit through the earth. The lost energy is transformed into other forms of energy, usually heat due to absorption process by inelastic and inhomogeneous mediums (mentioned in the previous section). During the passage of the wave, heat is generated during the compressive phase and absorbed during the expansive phase (Telford et al, 1990).

The loss of energy by absorption can be expressed in terms of the change of amplitude. It is given by :

$$A = A_0 e^{-\eta x} \quad 2.53$$

where A_0 is the initial amplitude of the wave at the source, A is the amplitude at a distance x , and η is the absorption coefficient in dB /wavelength (cycle).

The absorption can also be seen from the quality factor Q that is inversely proportional to the wavelength and absorption coefficient. It has been proved by experimental evidence that the absorption coefficient is proportional to frequency. The evidence shows that Q and η are roughly constant for any particular rock (Telford et al, 1990). This indicates that the increase in absorption with frequency provides one mechanism for the loss of high frequencies with distance.

Telford et al (1990) calculated the relative energy losses by absorption and spreading. These illustrate that (for low frequencies and short distances from source) the loss of energy by spreading is more important than that by absorption. Losses by

absorption are much more important for high frequencies and larger distances from source.

The partitioning process also reduces the wave energy. When a wave reaches the boundary, scattering, reflection, or refraction, and transmission will occur and each outgoing wave then becomes a fraction of the incident energy. As mentioned in the previous section the total energy of all new waves (except the scattering wave because it is very difficult to catch) is generally less than the incident wave. However, this simple assumption neglects that this loss of energy is using the principle of conservation of energy. Zoeppritz's equation (Sheriff and Geldart, 1995) expresses the partition of energy when a plane wave impinges on an acoustic impedance contrast (Sheriff, 1984). In the general case for an interface between two solid media, when the incident angle is not zero, four waves will be generated ; reflected P-waves & S-waves and transmitted P-waves & S-waves. The coefficients of reflection (R) and transmission (T) waves for normal incidence are respectively given:

$$R = (Z_2 - Z_1) / (Z_2 + Z_1) \quad \text{and} \quad T = 2Z_1 / (Z_2 + Z_1) \quad 2.54$$

where Z_i is the acoustic impedance and the index $i = 1$ or 2 indicates the top or bottom interface/layer. Assuming that there is no loss of energy, $R + T = 1$ (energy conservation) and fractions of energy reflected E_R and transmitted E_T are given ;

$$E_R = R^2 \quad \text{and} \quad E_T = (Z_2/Z_1).T^2 \quad 2.55$$

therefore $E_R + E_T = \text{Energy of incident wave}$. The larger the acoustic impedance contrast over the boundary the more energy is reflected and vice versa for the refracted wave. More detail will be described in the next section.

2.2. Shallow Seismic Reflection Method

2.2.1. Shallow Reflection

2.2.1.1. General Consideration

Seismic reflection surveys have several advantages over the seismic refraction method. Blind layer and thin-bed (hidden layer) problems do not exist, and greater resolution is possible as mentioned in the introduction. In the case of shallow applications, there are two classic problems; attenuation of high-frequency energy at the near surface and the suppression of coherent noise. The recorded dominant frequency may only be of the order of 20-50Hz due to the attenuation of high-frequency energy by the surface layer. Therefore, the technique has not been able to detect layers less than 5 to 25 m thick or to find structures smaller than these dimensions.

Recent developments in the shallow seismic reflection method have led to many successful applications (Steeple, 1998; Miller and Steeple, 1994). The improvement in instrumentation, field methods and data processing have increased both the resolution and the signal to noise ratio of processed stacked sections.

In addition, the shallow seismic reflection technique is inexpensive relative to drilling. It can increase the horizontal resolution and decrease the number of drillholes (Miller et al, 1995). The technique is extremely effective for detecting faults and interpreting stratigraphic relationships. Furthermore, it can estimate the depth to the latest and infer lithologies without drilling confirmation.

The optimum conditions for shallow reflection methods occur when the surface materials are fine-grained and water-saturated. In optimal situations reflections with dominant frequencies of 300-500Hz can be obtained to about 30m depth (from the surface). Therefore, in unconsolidated sediments which have a typical velocity of 1500-

1800 m/s the waves will have approximately 1m of vertical resolution and wavelengths of 3-5m. On the other hand, when the surface materials are coarse grained and dry, the dominant frequencies of reflection data can be less than 100 Hz. In this case, the vertical resolution may not be adequate to obtain the required information. One important method to optimise the resolution is to perform a field test to determine the optimal energy sources and field bandpass filters to use (Ali & Hill, 1991).

2.2.1.2 Depth of Target

The main characteristic of the reflection principle is the measurement of the travel times of longitudinal waves that have been reflected at subsurface boundaries. From measurement of reflection travel times it is usually possible to determine the depths and dips of reflecting horizons and the velocities of the seismic waves.

The basic concept of target depth is determined by the geological objective. The quality of the reflector as a target will be determined by the acoustic impedance contrast which is dependent on sonic velocity or density contrast. There are three issues that are closely tied to the depth and nature of the target reflector (Evans, 1997). Firstly, the energy source must have enough power to produce frequencies which image the target reflections. When the source is too strong, the dynamic range of the recording instruments may be saturated, ruining the fidelity of the data. However, if the source is too weak, a poor S/N ratio may result and the target will be imaged poorly.

Secondly, the fold of cover. The fold of cover should be sufficient to obtain a target signal-to-noise (S/N) ratio large enough for a correct interpretation. It relates to the common midpoint (CMP) method that one point has duplication of reflections from a single reflection point, but with different travel paths. This method is the well-known technique used to improve S/N ratios. Generally, the more traces that are stacked, the better the S/N ratio becomes, because the signal is additive while random noise is not. The fold of cover is an important survey parameter to estimate the sufficient S/N ratio for a

given depth. Signal processing theory states that signal-to-noise ratio improvement is proportional to the square root of the fold of cover, if all noise is truly random. However, for common geological conditions, an increase in fold of cover above a certain amount may make no noticeable improvement in data quality (Evans, 1997). Fold of cover is defined by:

$$F = (N \times S_R / S_S) / 2 \quad 2.56$$

where; F : Fold of cover

N : number of spread receiver station

S_R : receiver station interval

S_S : shot-point interval.

The value of 1/2 is necessary because adjacent CMP points are separated by one-half of the receiver station interval. The formula is valid for a shot point interval greater or equal to half the receiver station interval. If the shot point interval is less than the half station interval, the number of reflection points increases but the fold of each of them remains the same as when the shot point interval is half the station interval.

Finally, the source/receiver geometry must also be in an appropriate offset to optimise velocity calculations in order to image the target sufficiently. If the offset is too great, the apparent velocity (V_a) will be large. On the other hand, if it is too short we will find unlimited V_a . Ideally, the offset is decided by the angle of the wavefront around the critical angle.

2.2.2. High Resolution

We know that delineation of shallow subsurface structures is necessary for many engineering and environmental applications (palmer et al, 1997). For instance, GPR (ground penetrating radar) and seismic reflection are often used for those purposes (Kaida

et al, 1995). The main advantage of GPR is its ability to use the high frequency vibrator system, which will increase the resolution. However, there is still a need to increase the resolution further in order to produce the optimal high quality survey result. The main constrain of GPR is that the depth of penetration in soils is extremely limited (Kaida et al, 1995). The big problem of seismic reflection is that the wavelength of signals generated by seismic source is too large for sufficient resolution of small targets due to attenuated high frequency signals by near surface material. Here, the only concern is the resolution of the shallow seismic reflection technique.

Both vertical and horizontal resolution must be considered. With respect to seismic waves, vertical resolution is how far apart (in space or time) two interfaces must be to show-up as two separate reflectors. While the horizontal resolution is how far apart two features involving a single interface must be separated to show as two distinct features (Sheriff and Geldart, 1995). It is clear that the ability to see and distinguish features depends on the S/N ratio and the knowledge and experience of the interpreter.

2.2.2.1. Vertical Resolution

Vertical resolution is controlled by the frequency content of the recorded reflected seismic signal. The wider the bandwidth and the higher the frequencies recorded, the greater the resolution of the final stacked profile and the greater the definition and imaging of geologic boundaries (Evans, 1997; Narbutovskin et al, 1995). If noise is constant in amplitude, then the recorded reflection amplitude and S/N ratio will be greater, and the final stack more coherent. Increased bandwidth also helps lithological interpretations that depend on detailed knowledge of the amplitude and phase of reflection events. Therefore, it is important that the survey recording parameters do not compromise the survey objective by either temporally sampling the data too far apart or band limiting the recorded frequencies.

For a reflected pulse represented by a simple wavelet, the maximum resolution possible is generally between one quarter and one eighth of the dominant wavelength of the pulse (Sheriff and Geldart, 1995). Thus, for a reflection survey recording a signal with a dominant frequency of 100Hz propagating in sedimentary strata with a velocity of 2000m/s, the dominant wavelength would be 20m. Therefore, the vertical resolution may be no better than about 5m. Vertical resolution also decreases as a function of depth due to absorption and the increase of velocity with depth.

2.2.2.2. Horizontal Resolution

Horizontal resolution is described by the Fresnel zone. It is often taken as limiting horizontal resolution on unmigrated seismic data although other factors such as signal-to-noise ratio, and trace spacing (sampling), also affect how far apart features have to be, to be distinguished as separate features (Ramanantoandro, 1995). The Fresnel zone is the area on a reflection from which reflected energy arriving at a detector has phases differing by no more than a half cycle; thus, this energy interferes constructively (Sheriff and Geldart, 1995). The first Fresnel zone is often taken as a measure of the horizontal resolution of unmigrated seismic data. The radius of the first Fresnel zone (R) is given by the formula :

$$R = (\lambda h/2)^{0.5} = (V/2)(t/v)^{0.5} \quad 2.57$$

where h is the depth, t the arrival time, V the average velocity, λ the dominant wavelength, and v frequency. Thus, for a typical reflection survey: a signal with a dominant frequency of 100 Hz, a velocity of 2000 m/s, and the arrival time 50 ms, the horizontal resolution may be no better than about 7 m.

The other considerations of horizontal resolution are determined by detector or geophone spacing. In this case, the horizontal resolution is clearly determined by the spacing of the individual depth estimates from which the reflector geometry is reconstructed. For a flat-lying reflector, the horizontal resolution is equal to half the

geophone spacing as a Common-midpoint (CMP) spacing. This is a controllable field parameter, rather than the pure physical limit given by the Fresnel zone. The field parameters should be chosen so that resolution is limited only by the Fresnel zone.

2.2.2.3. Limiting Parameters

There are two commonly used techniques to increase the quality of shallow seismic data; the optimum-window, and the Common-Mid-Point (CMP) technique. Both techniques are now widely and routinely used in engineering, environmental, and ground-water application. Regarding the CMP shallow reflection interpretation (Steeple & Miller, 1994), there are some limiting parameters, which must be selected carefully:

1. Spatial aliasing of ground roll.
Decreasing the receiver station interval will improve the coherency of true reflectors and reduce the danger of spatially aliased ground roll.
2. Enhancing or not attenuating ground roll during CMP processing
Ground roll noise always becomes a major problem in shallow seismic reflection
3. Ground-coupled air wave as true seismic wave.
Air wave noise emitted from sources can often cause serious problems in processing.
4. Refraction as reflection on stacked CMP section.
It is often difficult to distinguish a shallow reflection event from a shallow refraction event during processing. Separating the reflection signal from a refraction signal is the major limitation of the shallow seismic reflection method.
5. Not recognising processing artifacts.
This may result from insufficient velocity analysis and inaccurate static correction, and becomes a new problem.

2.3. Shallow Seismic Reflection Survey

For shallow (<100 m depth target) seismic reflections, surveys are carried out on a linear traverse in order to build up a two-dimensional picture of the subsurface. This section will describe the basic consideration of practice operation that include; data acquisition, data processing and interpretation.

2.3.1. Acquisition

2.3.1.1. General Consideration

The basic requirement of data acquisition is high-quality field work (Davies and King, 1992). A fundamental factor affecting the quality of final shallow seismic reflection data is the nature of the near-surface geology (Kragh, et al 1992). They state that the best source coupling seismic profiling was found where the overburden was fine-grained and water-saturated. They recommend choosing ground conditions with no weathered layer and, with sufficient ground resistance to surface impacts or small explosions so that high-frequency signals can be easily generated.

Furthermore, the common midpoint (CMP) method is the field method used almost exclusively today (Sheriff and Geldart, 1995). Its objective being to increase signal to noise ratio (S/N) by using the multi-fold recording where one point on the subsurface is sampled several times and then the coverage is called "X-fold" recording (sometimes called X00%). Correct choice of source type, geophone pattern, and group spacing is required to optimise the quality of reflected signals. Generally, the selection of field parameters depends on both geologic objective and local noise conditions.

2.3.1.2. Field Parameter Improvement

To improve field parameters the following factors must be given consideration along with the recording equipment used, such as number of available recovery channels and geophones, how cables and geophones are wired, etc (Steeple & Miller, 1998; Keiswetter & Steeples, 1995; Hill, 1992b; Iverson & Smithson, 1982). For a single receiver recording system, there are four points to be considered, in order to improve shallow reflection data quality for identifying water table reflections:

1. The maximum offset, the distance from source to the furthest group, should be comparable to the depth of the deepest zone of interest. This distance is the critical distance discussed in section 2.1.2.3. This usually results in large enough normal-moveout differences to distinguish primary reflections from multiples and other coherent noise. However, the offset should not be so large that reflection coefficients change appreciably, or that conversion to shear wave becomes a problem, or approximations of the CMP method become invalid (Sherrif and Geldart, 1995). If data quality in the deepest zone of interest is sufficiently good, the maximum offset may be increased up to the value of the basement depth as the deepest reflector of interest.

2. The minimum offset, the distance from source to the nearest group, ideally should be no shorter than the depth of the shallowest section of interest. Getting sufficiently far from source-generated noise sometimes needs a greater distance, but this may cause a loss of useful shallow data. On the other hand, a fundamental difficulty with shallow seismic reflection surveying on land is that for receivers close to the source, there can be substantial interference from refracted arrivals and high amplitude surface waves.

3. The receiver station interval should be no more than double the desired horizontal resolution, thus providing subsurface spacing equal to the desired resolution. High-frequency (100 Hz) geophones are required to record the high-frequency energy necessary for the desired vertical resolution (Knapp & Steeples, 1988).

4. The minimum charge size or source effort is determined by the ambient noise observed late on the record. Random noise should not affect repeated records until after a depth below the deepest section of interest. It is essential that the seismic source is of adequate power to produce enough reflected seismic wave energy. However, more powerful sources usually lower frequency signals (Hill, 1992a).

It should be remembered that line length, line orientation, and line spacing are also field parameters (Roberts, 1992). In the case of a shallow target, minimum offset is essential to avoid recording both refraction and surface wave occur.

2.3.1.3. Field Test

Before starting a seismic reflection survey, it is necessary to perform field tests in order to choose and set-up optimal field parameters. Regarding noise analysis, we take a set of geophones and a seismic source (e.g. 5kg sledge hammer and a plate) and set the field geometry as planned. By using different geophone station spacings, the shot records should be different so we have a selection of possible filter ranges. The seismograph executes and analyses the records to verify that the instruments are functioning properly. It checks the entire system, including geophones.

It is well known that the best coupling under most conditions is obtained with geophones firmly planted, usually on long spikes (Knapp & Steeples, 1996a). The geophones are normally fitted with 125 mm spikes that sufficiency improve ground coupling. The relative amplitude of reflection wave compared to ground roll are to some degree a function of coupling to the ground. If geophones are poorly coupled to the ground, ground roll is enhanced relative to reflection because poor coupling favours low frequencies (Steeple & Miller, 1994).

The time (t) Vs distance (x) plot of seismic events using an excel spreadsheet (see section 4.1) can be used as a simple reference for a field test record (Knapp &

Steeple, 1996b). By knowing the shallowest and deepest targets of interest, and the line length of the survey, we can plot and optimise the record window by choosing the correct geophone spacing, record length and minimum & maximum offset. Record length is decided by the deepest target reflection while geophone spacing depends on how much we want to optimise; the horizontal resolution, the spatial aliasing of surface waves and the improvement of the coherency of true signals. Minimum & maximum offsets are decided by two points; (1) the shallowest and deepest targets of interest and (2) the physical properties of near surface material. These produce first arrival waves (direct wave and refraction signal) and surface waves as coherent noise i.e. ground roll and air waves. These field parameters will be used to assess field design by combining with the field test record.

The recording digital-sampling rate is chosen to avoid the loss of information in the recording process with minimising the data volume and subsequent data processing time and cost. The Nyquist sampling theorem states that the sampling frequency must be greater than twice the maximum signal frequency of interest.

Regarding the use of digital frequency filters in the field, we can apply low cut, high cut, or bandpass filters. This depends on what and/or how strong the coherent noise effectively masks the main signal. This coherent noise can be observed in the field test record. For example a low cut filter may be used to remove ground roll as the low frequency energy is below that of the reflection signal. In this case, we should remember that the broader the bandpass window, the higher the result of vertical resolution (see next part in this chapter). However, we also should note that there are some distorted features that are caused by the mechanical characteristics of geophones. The distorted data are commonly in the range below the geophone frequency. For example, the 100Hz geophone will include those informations at $< 100\text{Hz}$. For this reason, using low cut filters around 100 Hz is appropriate.

A sledge hammer is the simplest impulsive source type used for shallow seismic surveys. Commonly, this source will produce coherent noise and we can very easily remove or attenuate noise such as air waves. The last field parameter that we should

decide is the number of shots stacked. Stacking is needed to increase the signal to noise ratio in the shot record. This depends on how strong the main signals are in the single shot field test record. This case, we usually took five shot stacks that make a shot record the square root of five (or about 2.24) times greater than the single shot field test record.

2.3.2. Data Processing

It is clear that the object of seismic data processing is to take seismic shot records from the field. These records in-turn produce a coherent cross-section, indicating significant geological horizons in the subsurface. The author will improve the quality of processing based on the limiting parameters as mentioned in section 2.2.2.3. This is a special process that solves some, but not all weaknesses in shallow reflection. Additionally, it helps decrease the negative effect of those limiting parameters. This includes the common problem of near surface records; ground roll, air waves, and direct and refraction waves.

It is very good if coherent noise effects such as refraction and ground roll in both velocity and frequency, can be recognised in the shot record. Removing these phases completely requires, either top or surgical muting (Doornenbal & Helbig, 1983). However, misidentifying non-reflective phases, or failing to attenuate or eliminate them, may lead to artifacts in the next step of process (Steeple et al, 1997). Appendix B describes the use of Promax version 7.2. for the data processing in this research.

Commonly, there are great variations in near surface velocities, and in surface elevation changes. Both these factors require, extremely important static corrections and velocity analyses. For instance, the near surface often has extremely high velocity gradients, so that velocity may vary by an order of ten within a few meters. Also, the near surface sometimes exhibits heterogeneity on the scale of a few high-frequency seismic wavelengths (Steeple et al, 1997; Lericolais et al, 1990). Coherency filtering must be applied carefully, and spatial aliasing of air blast and ground roll can

contaminate close traces when f-k filtering is not applied carefully. Finally, when processing is complete, the final stack should be consistent with the initial brute stack.

2.3.2.1. Pre-Processing

Pre-processing includes recording field data records, data transfer from the seismograph to the processing software, system geometry input, and trace editing. It is important to understand the recorder system before attempting processing. All field files of this research were recorded on the Bison 9048, downloaded to a network PC using the Bison “menu” program, then transformed to the Promax Sun Computer, Sun6 by FTP running a binary transfer. A Promax flow was run using “Floppy input” to read the Bison files and convert them to one SEG Y equivalent file.

The system geometry input is the set of field parameters named “geometry input”. This is a flow that contains 2D or 3D Land Geometry Spreadsheets. The flow should be executed the first time processing is started in Promax systems for a line. It is used to enter all the required field geometries such as receivers, sources, position and possibly any pattern of spreadsheets. For example, to set the receiver we need all geophone positions for every shooting.

Editing includes trace editing that follows format verification and trace kill/reverse that is commonly used for killing or reversing traces. The data are rearranged. Field data are usually time-sequential so that the first sample for each channel is recorded before the second channel for any one channel. Editing may involve detecting dead or exceptionally noisy traces. Bad data may be zeroed out or replaced with interpolated values. Anomalously high amplitudes, which are probably noise, may be reduced to zero or to the level of the surrounding data. The list of traces for editing may be retrieved from database entries created by earlier interactive screen selection. Alternatively we can use the editor to specify trace numbers for editing. This step can

also be used to find traces in which all samples are zero (dead) or have reverse polarity. Scanning for dead traces immediately upon input is strongly suggested.

Trace muting is also classified as part of data editing. The term is commonly applied for the process of zeroing the unwanted part (or parts) of a trace. In mathematical terms, every trace is multiplied by zero for each muted part and by one for each non-muted part. Using a ramp in a time scale is as a slope for the area in between muted and non-muted part.

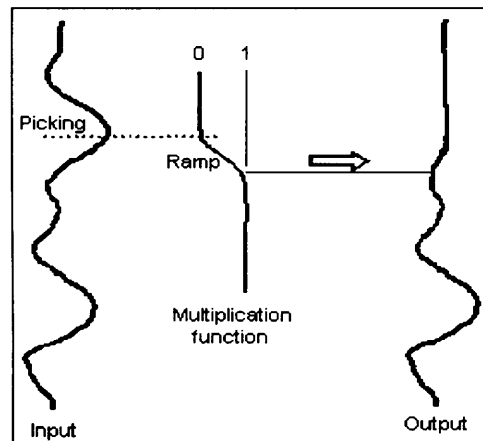


Figure 2.8 illustration of the top mute process

Figure 2.9 illustrates the top mute process. To avoid stacking non-reflection events, such as first arrival and refraction arrivals, with reflection then the first part of the trace is normally muted before the next step to the stacking process. This is obviously also done to avoid degrading the quality of shallow reflections (Sheriff and Geldart, 1995). Normally, surgical mute is applied to remove the airwave while bottom mute is used for ground roll. However, in many cases air blast attenuation and f-k filtering are better than both mutes. The main point is that the seismologist does not want to miss any information due to the zeroing process of mutes.

2.3.2.2. Static Correction

Static corrections are used to correct for surface elevation and weathering velocity variation. Refraction static correction is commonly applied to seismic data to compensate for the effect of these variations. Operationally, it includes three steps; first break picking, elevation static, and refraction static.

First break picking is based on first-break refraction arrival time, and provides a means of dealing with long-wavelength variations. The objective of this static is to determine the refraction arrival times, which would have been observed if all measurements had been made on a (usually) flat plane with no weathering or low-velocity material present. In Promax this flow automatically picks first breaks (see figure 2.10). The only potential problem with this type of picker is the occurrence of false picks in the pre-first break noise such as footsteps or wind gusts. The system has a stabilized power ratio to avoid these false picks. Using the certain lengths of the leading and trailing gates it can automatically determine our estimate of the first break.

Elevation static is applied to fix the datum as a reference or final datum which this research uses above the surface. This is crucial because the seismic data is typically shifted to a reference datum plane to correct for travel-time effects of weathered layer variation and surface topography. Using the supplied elevation data this process not only calculates the thickness of air-column (V_0), but also replaces V_0 using the replacement velocity of V_2 . It calculates and applies shot and receiver statics to the final datum (see figure 2.10).

The objective of refraction static is to replace certain velocities, commonly using V_2 (see figure 2.10) in the weathering layer, where boundaries have been supplied by elevation statics. Therefore the main aim of this is to correct for variations in the velocity and thickness of the weathering layer. Commonly the process of shallow seismic survey uses the final datum above the surface elevation and $V_2 > V_1 > V_0$. Therefore, the result will reduce the time travel and consequently the figure as the whole of the signal rises proportional to those variations. The technical process of this correction can be seen in the appendix B.

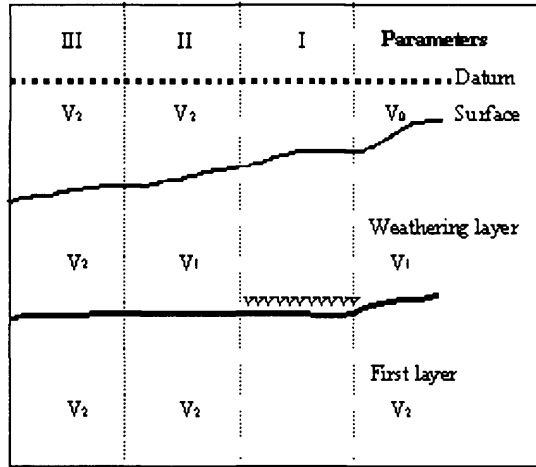


Figure 2.10 three steps of static correction; I first break picking to pick first arrival, II Elevation static to fix datum and replace V_0 using V_2 , III Refraction static to replace V_1 using V_2

Figure 2.10 also illustrates a method of obtaining total static correction, Δt . It is made up of two parts, the source correction Δt_s and the receiver correction Δt_r . The formulation is given as:

$$\Delta t = \Delta t_s + \Delta t_r \quad 2.58$$

$$\Delta t_s = \frac{h-s}{V_1} + \frac{s+E}{V_2} \quad 2.59$$

$$\Delta t_r = \Delta t_s - t_v \quad 2.60$$

where : E is the datum elevation, h is shot position at surface elevation, s is a thickness of weathering layer (a distance between a surface to the first layer), and t_v is vertical travel-time that is always positive (see Al-sadi, 1980, p-171).

2.3.2.3. Filtering (Various Type)

In seismology, filtering is nearly always taken as meaning a modification of the spectrum of the seismogram (Hatton & Warthinton, 1986). In other words, filtering applies the defined criteria that known as frequency filtering. In application it is should be qualified, for example, as band-pass filtering, f-k filtering, and so on. Filters play an important role in processing seismic data events, and the recording instruments used in data acquisition can also apply filtering. Described here are only the filters relevant to this research.

2.3.2.3.1. Band-pass filtering

Frequency filters are classified as band-pass, band-reject, low-pass (high-cut), and high-pass (low-cut) filters according what they discriminate against above or below a certain limiting frequency outside or inside of a given band of frequencies. All of these filters are based on the same principle: construct a zero-phase wavelete with an amplitude spectrum that meets one of the four specifications (Yilmaz, 1988). These filters apply a frequency filter(s) to each input trace. This allows us to perform three types of band-pass filtering; single band-pass filter, time variant filter, and time & spatial variant filter. In common process, a single band-pass filter is chosen. A single filter is usually applied to all traces all the time.

Theoretically, the goal of band pass filtering is to pass a certain bandwidth and to suppress the remaining part of the spectrum as much as is practical. It appears that this goal can be met by defining the required amplitude spectrum for the filter operator as given :

$$A(f) = \begin{cases} 1, & f_l < f < f_h \\ 0, & elsewhere \end{cases} \quad 2.61$$

where f_l and f_h are the low-high as cut-off frequencies. This is known as boxcar amplitude spectrum. To smooth the result it needs ramp function at both low and high edges. Practically, this filter inputs via the keyboard and then the four corner (f_1 , f_2 , f_3 , f_4 as shown in figure 2.11) frequencies need be specified for this.

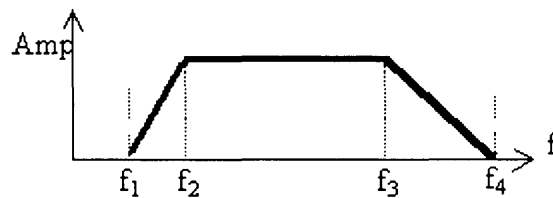


Figure 2.11 window of band-pass filter, f_1 - f_2 and f_3 - f_4 as ramps

Regarding the vertical (temporal) resolution, choosing a bandwidth as frequency filter is the key factor in improving the quality of seismic data. There is a common misunderstanding that only high frequencies are needed to increase resolution. This is not true, as Yilmaz (1988) demonstrates that having only low or high frequencies does not improve vertical resolution. Both low and high frequencies are needed to increase that resolution. He concluded that closely situated reflectors can be resolved by using increasingly broader bandwidths. For example, 10 to 50Hz bandwidth is sufficient to resolve reflectors with 24ms separation while the 10 to 100Hz is needed to resolve reflectors that are separated by 12ms.

2.3.2.3.2. Air Blast Attenuation

Shot-generated air blasts are strong energy sources recorded as broad band noise. They may contaminate seismic data with high amplitudes, especially when using surface sources such as sledge hammer. This noise must be muted (mentioned previously) but unfortunately, such surgical mutes may produce undesired and unexpected results. Air Blast Attenuation attempts to overcome this problem by providing an automated air-blast attenuation tool. It will automatically seek out air blast energy on a trace-by-trace basis, given a pilot velocity, a relative noise amplitude threshold, and an approximate energy envelope width. It can also be used to attenuate other linear noise trains that are strong relative to the surrounding data. Commonly, air blast attenuation is used to remove/attenuate airwave signal.

2.3.2.3.3. F-k filtering

The f-k filter is a frequency-wave number filter applied to the data in the frequency-wave number domain. The objective of this filter is to remove the effect of back-scattering and apparent velocity or ground roll. Data is converted from time and space sampled traces to the f-k domain by a two-dimensional Fourier Transform. After the f-k filter is applied the data is converted back by the inverse Fourier Transform in the form of the seismic traces (Promax manual, 1997).

A significant parameter of this step is the window or range of the filter (Harris and White, 1997; Jeng, 1995). Figure 2.12 illustrates the area of reflection energy used to decide this filter windows. It is the range of velocity and frequency that should be decided in f-k analysis. An interactive process of f-k analysis is operated to decide this range. The value of velocity appears immediately when the cursor in the f-k section is picked. In this way we can chose the appropriate range in pair of velocity (usually it is symmetry for minus and plus value) and frequency (f_1 and f_2 see figure 2.12).

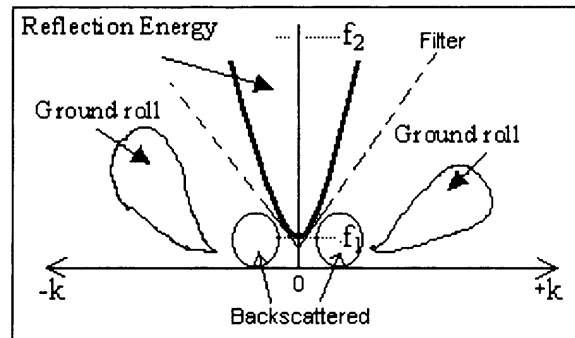


Figure 2.12 the simple f-k diagram for f-k filter of symmetrical split spread.

2.3.2.4. Velocity Analysis and Dynamic Correction

Velocity analysis is the most vital process in seismic data processing sequence (Tsvankin, 1997). It is used to determine the seismic velocities that correspond to the type or scale of application of seismic surveying. These velocities will be applied to the Normal MoveOut (NMO) correction and also for Time-to-Depth conversion. The highest quality of the stack is critically decided by the accuracy of determination of the seismic velocity. Consequently, this determination becomes the most critical parameter.

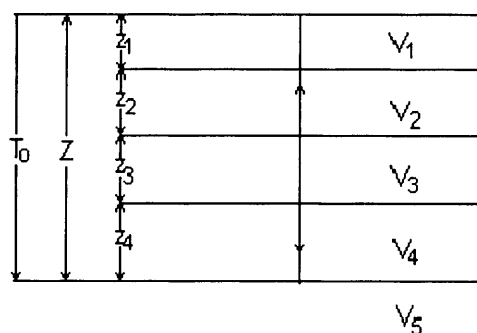


Figure 2.13 Simple schematic of seismic velocity analysis term, the total number of layers (n) = 5 and so $n-1$ refers to fourth layer (after Reynolds)

Figure 2.13 shows the horizontally layered sequence, where each layer has an interval velocity of V_{int} (V_1, V_2, V_3, V_4 , and V_5) and obviously also has interval transit time $t_{\text{int}} = z_i/V_{\text{int}}$. The average velocity V' is total raypath length (Z) divided by total travel time T_o . The weighted-average velocity is termed to Root-mean-square (RMS) velocity (V_{RMS}) :

$$V_{\text{RMS}} = \left[\sum V_i^2 t_i / \sum t_i \right]^{1/2} \quad 2.62$$

This velocity is normally used to approximate the seismic velocity in order to design the velocity function of NMO correction. When x is the offset distance, two-way travel time of a ray reflected from the n th at a depth z is :

$$t_n = (x^2 + 4z^2)^{1/2} / V_{\text{RMS}} \quad 2.63$$

It is well known that Dix formula is an interval velocity (V_{int}) using V_{RMS} , for V_{int} over the n th interval is given :

$$V_{\text{int}} = \left[\frac{(V_{\text{RMS}.n})^2 t_n - (V_{\text{RMS}.n-1})^2 t_{n-1}}{(t_n - t_{n-1})} \right]^{1/2} \quad 2.64$$

where $V_{\text{RMS}.n}$, t_n and $V_{\text{RMS}.n-1}$, t_{n-1} are the RMS velocity and reflected ray two-way travel times to the n th and $(n-1)$ th reflectors respectively.

Dynamic correction is Normal Move Out (NMO). It is normally termed ΔT , as the difference between the actual travel time T_x and the corresponding two-way vertical time T_o . The usual expression for ΔT (for a horizontal reflector) is given :

$$\Delta T = T_x - T_o \approx x^2 / 2V^2 T_o \quad 2.65$$

$$\text{where } T_x = \frac{2}{V} \sqrt{\left(\frac{x}{2}\right)^2 + z^2} \quad \text{and} \quad T_o = \frac{2z}{V}$$

Figure 2.13 shows the horizontally layered sequence, where each layer has an interval velocity of V_{int} (V_1, V_2, V_3, V_4 , and V_5) and obviously also has interval transit time $t_{\text{int}} = z_i/V_{\text{int}}$. The average velocity V' is total raypath length (Z) divided by total travel time T_o . The weighted-average velocity is termed to Root-mean-square (RMS) velocity (V_{RMS}) :

$$V_{\text{RMS}} = \left[\sum V_i^2 t_i / \sum t_i \right]^{1/2} \quad 2.62$$

This velocity is normally used to approximate the seismic velocity in order to design the velocity function of NMO correction. When x is the offset distance, two-way travel time of a ray reflected from the n th at a depth z is :

$$t_n = (x^2 + 4z^2)^{1/2} / V_{\text{RMS}} \quad 2.63$$

It is well known that Dix formula is an interval velocity (V_{int}) using V_{RMS} , for V_{int} over the n th interval is given :

$$V_{\text{int}} = \left[\frac{(V_{\text{RMS},n})^2 t_n - (V_{\text{RMS},n-1})^2 t_{n-1}}{(t_n - t_{n-1})} \right]^{1/2} \quad 2.64$$

where $V_{\text{RMS},n}$, t_n and $V_{\text{RMS},n-1}$, t_{n-1} are the RMS velocity and reflected ray two-way travel times to the n th and $(n-1)$ th reflectors respectively.

Dynamic correction is Normal Move Out (NMO). It is normally termed ΔT , as the difference between the actual travel time T_x and the corresponding two-way vertical time T_o . The usual expression for ΔT (for a horizontal reflector) is given :

$$\Delta T = T_x - T_o \approx x^2 / 2V^2 T_o \quad 2.65$$

$$\text{where } T_x = \frac{2}{V} \sqrt{\left(\frac{x}{2}\right)^2 + z^2} \quad \text{and} \quad T_o = \frac{2z}{V}$$

We can see that the offset-dependent NMO is a hyperbolic function, quadratic offset distance. It is clear that the larger the offset will be the greater the NMO correction. As a general rule a velocity tends to increase with depth. Equation 2.65 indicates that the greater the two-way travel time or velocity, the smaller the correction for any offset. Figure 2.14 illustrates the term of reflection NMO. The application of NMO may cause a change in wavelet shape called stretching. This can be controlled or anticipated by a NMO test for certain CMP gather before NMO processing.

In order to process data the NMO correction it is required to compute ΔT as accurately as possible. Normally both x and T_0 are known as we can calculate from shot record. Therefore the accuracy of ΔT -determination depends largely on how successful we are in estimating V . The error in this estimation incurs a corresponding error in ΔT commonly referred to as the residual NMO, $\Delta\Delta T$ or $\Delta^2 T$.

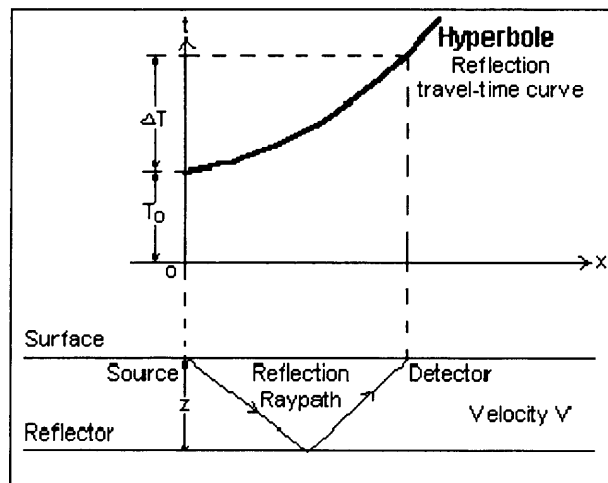


Figure 2.14 illustration of term of reflection NMO

2.3.2.5. Stacking

The simple law of reflection states that the reflection point is located under the source-detector mid-point. It is possible to have several traces that have the same mid-point as common mid-point (CMP), from different shot records. The basic conception of stacking is to sum together (gather) all CMP traces to generate a new simple trace. The main purpose of stacking is to improve the signal to noise ratio and to increase the resolution (vertical) of reflection events as the reduction of incoherent noise. The improvement in the reflection signal due to stacking depends largely on the accuracy of the NMO correction.

2.3.3. Interpretation

Seismic data is usually interpreted by a geophysicist or a geologist. The ideal interpreter combines training in both fields (McQuillin et al, 1979). It is necessary to fully understand the process involved in the generation and transmission of seismic waves, the effect of the recording equipment and data processing, and also the physical significance of the seismic data.

2.3.3.1. Geophysical Interpretation

The primary object of geophysical interpretation, in the whole seismic exploration or investigation, is usually to prepare contour maps showing depths for a series of reflectors, which have been picked out on the seismic section. This interpretation is based on physical properties such as wave velocity, rock density, frequency and wavelength. For example, variation of seismic reflections is caused by the variation of reflection coefficient where acoustic impedance changes. Often the geophysicist is involved in seismic

investigation from the first stage through to the last stage, from survey planning until the interpretation of the data.

Semblance velocity has been used to apply the normal moveout correction and to convert time to depth. We can easily identify the reflection from the semblance velocity section, and find the high quality of reflector and the value of semblance velocity. We can also identify the appropriate reflector and estimate velocity errors. For example to calculate an error, we estimate the maximum and minimum velocity (V_1 and V_2) from one selected point where the line of picking has approximately no change. Then we can calculate the error estimation from those values.

2.3.3.2. Geological Interpretation

The main purpose of geological interpretation is to show the geological meaning of seismic reflection patterns to reflectors as boundaries. These may be boundaries marked as a faults and/or stratigraphic contacts between two layers. Furthermore, it is to show that we can distinguish features that are not marked by sharp boundaries. The most importance key is to properly identify the reflection and reflector position. A seismologist can easily interpret the seismic section after using time/depth conversion. Therefore, a detailed geological structure such as a fault, undulation, and/or dips of reflectors can be shown and/or calculated.

Regarding the identification of stratigraphic boundaries, formations can be described in terms of age, thickness, and lithology of constituent layers. Some formation boundaries mark distinct changes in lithology such as an abrupt change from shale to limestone (Robinson & Coruh, 1988). In addition, to distinguish different geologic formations that have very weak boundaries marking of seismic reflections is achieved by comparing the position of stratigraphic boundary of a special well geophone or sonic logging with the reflection in the seismic sections.

Chapter 3

HYDRO-GEOLOGICAL BACKGROUND

3.1. Introduction

It is not easy to understand the way in which water occurs underground without considering the sub-surface geological structure. Groundwater, not soil water, may be defined as the subsurface water in rocks that are fully saturated (Ward and Robinson, 1990). The level below which all the pores in the rock are totally filled with water, or the upper limit of the saturated zone, is called the water table (Price, 1985). This is also defined as the level where pore-water pressure is equal to atmospheric pressure. This is the boundary between the unsaturated zone and the saturated zone. It is well known that the water table tends to follow the contour of the overlaying ground surface, although in a more simple form.

An aquifer, in geological terminology, is defined as a geological formation comprising of layers of rock or unconsolidated deposits that contain sufficient saturated material to yield significant quantities of water. While, other formations that are much less permeable and can only transmit water at much lower rates than the adjacent aquifers are commonly known as aquitards (Freeze, 1979; Price, 1985). Most of the major aquifers are composed of sedimentary deposits that formed from the erosion and deposition of other rocks (Ward and Robinson, 1990). In contrast, igneous and metamorphic rocks are formed under conditions of high temperature and pressure, and hence generally only have little interconnected pore space. Therefore, consequently most have only limited capacities for water saturation.

There are three types of aquifers; perched, confined, and unconfined aquifer. Figure 3.1. shows how the presence of an aquitard as a confining layer, such as clay, can form the structure of an aquifer and may support a perched water table above the main water table. While figure 3.2. shows how both confined and unconfined condition can

occur in the same aquifer. In zone A, the aquifer is fully confined by the overlying clay and is fully water saturated. In zone B, the aquifer is overlain by the clay but is not fully saturated. The aquifer in zone C is unconfined. Seasonal fluctuations in the water table level, are likely to give a minimum at the end of dry season and at its greatest extent in early dry season (Brassington, 1988). Recharge before dry season causes groundwater level to rise back to the highest level.

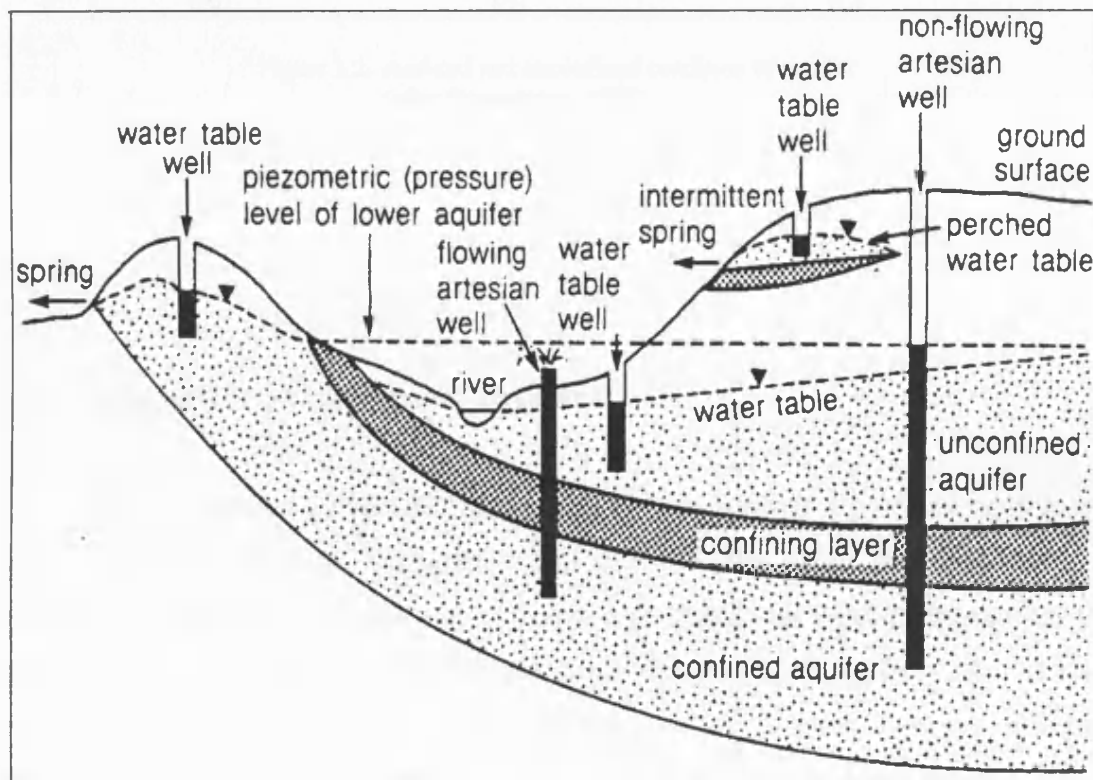


Figure 3.1. the type of aquifers (after Brassington, 1988)

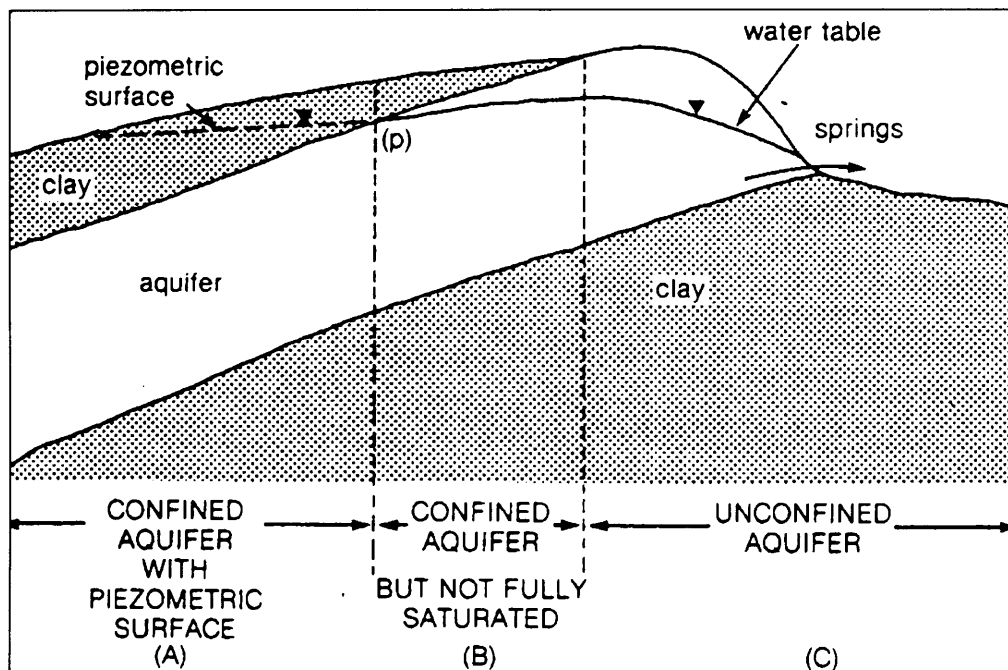


Figure 3.2. confined and unconfined condition of aquifer
(after Brassington, 1988)

3.1.1. Lithology, Stratigraphy, and Structure

The purpose of a hydrogeological survey is to identify a potential aquifer and then examine its lithology, stratigraphy, and structure. The nature and distribution of aquifers and aquitards in a geologic system are controlled by those parameters of the geological deposits and formation. Examination of topographic maps of the area are useful in identifying spring lines. Where springs are used for water supply, one should be aware of spring collection chambers and storage tanks (Brassington, 1988).

The following definitions are common parameters used in groundwater terminology adopted from Freeze (1979). The *lithology* is the physical makeup, including the mineral composition, grain size, and grain packing, of the sediment or rocks that make up the geological system. The *stratigraphy* describes the geometrical and age relations between the various lenses, beds, and formations in geological

systems of sedimentary origin. *Structural features*, such as cleavages, fractures, folds, and faults are the geometrical properties of the geologic systems produced by deformation after deposition or crystallisation. In most regions knowledge of lithology, stratigraphy, and structure directly lead to an understanding of the distribution of aquifers and aquitards.

Figure 3.3 illustrates the situation in which the stratigraphy and structure control the occurrence of aquifers and aquitards. Gently dipping sandstone aquifers (figure 3.3.a) outcrop along the mountain front. Interfingering sand and gravel aquifers (figure 3.3.b) extend from the upland into the intermountain region. Faulted and folded aquifers (figure 3.3.c) occur in desert regions e.g. in Sahara region of Africa. The structure controls the occurrence of water surfaces. All of these cases indicate that surface water bodies reflect structural features.

Aquifers are commonly associated with unconformities, either in the weathered or in the fractured zone immediately below the surface of a buried landscape (Linsley et al, 1988). In terrain that has been deformed by folding and faulting, aquifers can be difficult to discern because of the geologic complexity. In this situation the main objective in groundwater investigation is often a large-scale structural analysis (Freeze, 1979).

Sedimentary rocks are formed as a result of the deposition of particles derived from the weathering or/and erosion of rocks. The deposition commonly takes place under water i.e. the seabed, lake or riverbed. Particles will be sorted and deposited depending on their size with the smallest grains deposited after larger grains. Mineral grains may then be cemented by another material and compacted as a result of burial beneath other layers of sediment during the process of consolidation. Essentially, unconsolidated sediment becomes consolidated sediment while the physical properties of rock are changing. More detail will be discussed in the next section on physical properties of rocks.

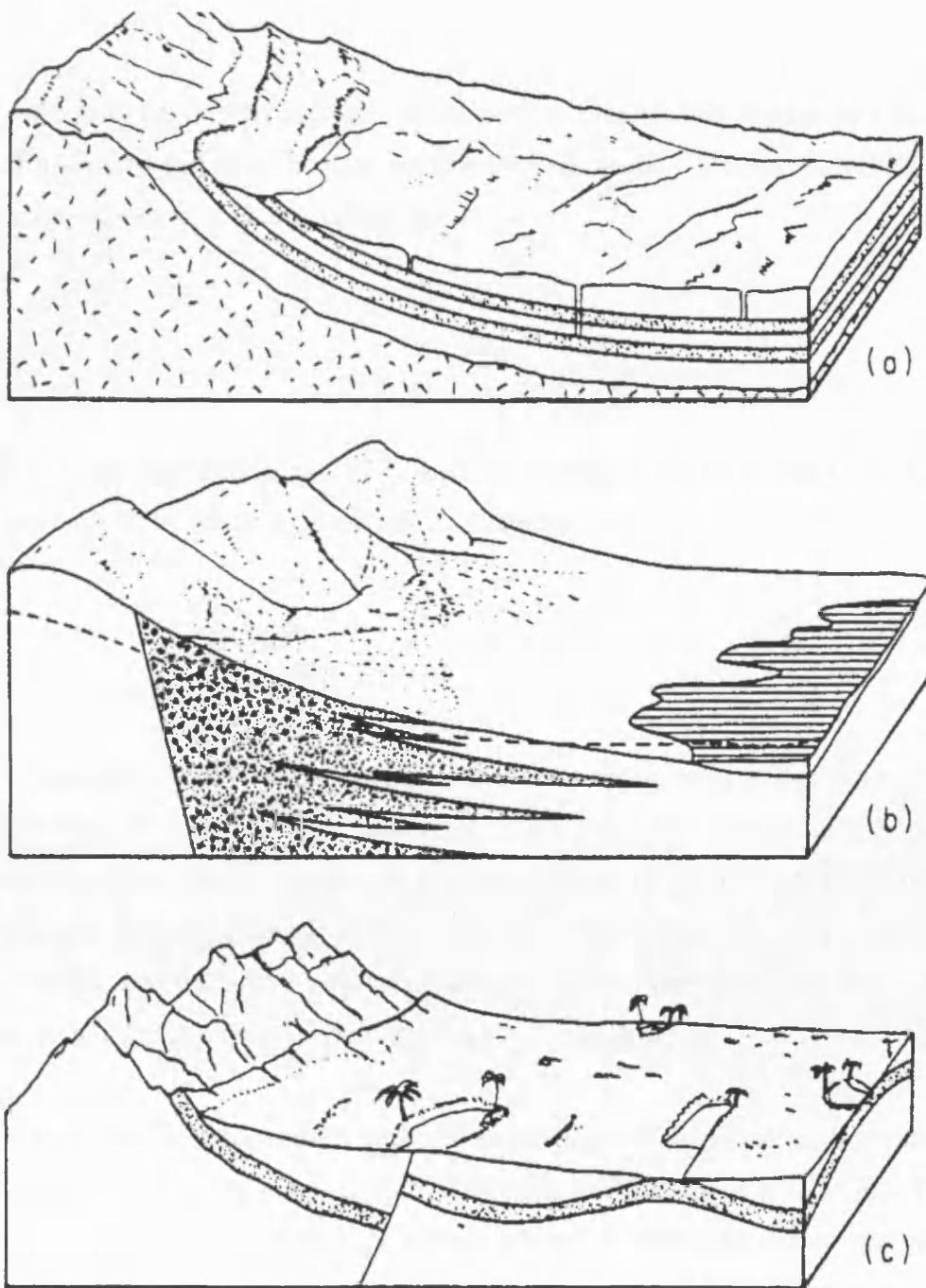


Figure 3.3 Influence of stratigraphy and structure on regional aquifer occurrence. (a) Gently dipping sandstone aquifers; (b) interfingering sand and gravel aquifers; (c) faulted and folded aquifer (after Freeze and Cherry, 1979)

3.1.2. Physical properties: Porosity and Permeability

Porosity (ϕ) directly indicates the amount of groundwater storage in a saturated rock. An adopted definition of total porosity from Domenico & Schwartz (1990) is the percentage volume of rock that is void space:

$$\phi = \frac{V_v}{V_T} \quad (3.1)$$

where V_v is the void volume and V_T is the total volume. It can be written in the term of void ratio, $e = V_v/V_s$ where V_s is the solid volume, as :

$$e = \frac{\phi}{1 - \phi} \quad \text{or} \quad \phi = \frac{e}{1 + e} \quad (3.2)$$

Generally, smaller particle-sized sedimentary rocks, have higher porosities but typically very low permeabilities, due to the larger amount of impermeable cementing material listed in table 3.1 (Ward & Robinson, 1990; Davis & DeWiest, 1966). As mentioned in the last paragraph of 3.1.1, the physical properties of rock change during consolidation, and the porosity of fine-grained sediments decreases with burial depth. Compaction will expel the pore fluid such as water, oil, or gas.

Most fine-grained detrital rocks have relatively high porosities but very low permeabilities (Price, 1985; and Davis & DeWiest, 1966) as listed at table 3.1. This is because very small pores have high surface tension or molecular forces and prevent water movement. On the other hand sediment that contains both large and small grains (poorly sorted) will have low porosities because small grains tend to occupy the voids between the larger grains.

Table 3.1 List of indicative porosity and permeability for consolidated and unconsolidated sedimentary rocks (after Brassington, 1988, p-55)

Geological Material	Grain size (mm)	Porosity (%)	Permeability (meters per day)
Unconsolidated sediments			
Clay	0.0005-0.002	45-60	$< 10^{-2}$
Silt	0.002-0.06	40-50	10^{-2} -1
Alluvial Sand	0.06-2	30-40	1-500
Alluvial Gravel	2-64	25-35	500-10 000
Consolidated sedimentary rocks			
Shale	Small	5-15	5×10^{-8} - 5×10^{-6}
Sandstone	Medium	5-30	10^{-4} -10
Limestone	Variable	0.1-30	10^{-5} -10

3.2. Morphology and Classification of Aquifer

It is well known that the type of aquifer (perched, confined, and unconfined) depends on the structure or geological condition and these aquifers are directly controlled by stratigraphy and geological structure. This section will describe the classification and morphology of both simple and complex aquifers.

3.2.1. Simple Morphology

The simplest morphology occurs when a homogeneous rock of high permeability extends beneath the weathering layer of the land surface to some depth, and water table will then occur in unconfined condition. These aquifers are uniform in composition and the horizontal water table is controlled by topography of land surface. In other word, the flat layered heterogeneity is very simple; weathering layer and

unsaturated – saturated layer separated by water table boundary. This will be discussed in detail in chapter five.

3.2.2. Complex Morphology

Referring to the type and occurrence of aquifers, and the strong control exerted by stratigraphy and geological structure, it is easy to imagine how complex morphologies can occur. For example, a confined aquifer in a dipping layer of a syncline.

Figure 3.3 (a) shows a dipping sandstone aquifer that is adopted as our complex geological structure. In small scale, figure 3.2 zone B illustrates that the water table has an angle to the clay which acts as confining layer. This type of water table has an interesting morphology which can be investigated using reflection seismic method. P-wave surveying will image both the water table and the very low permeability of clay. However, the S-wave method will only image the lower clay as a geological boundary, without imaging the water table, due to the physical properties of this wave (see section 5.5.4).

The other complex structure we will discuss is the occurrence of a perched water table caused by a lens structure in an unconfined aquifer above a regional water table (figure 3.1). Figure 3.4 shows in detail a schematic presentation of such a water table. The shape of water table is controlled by both lens morphology and land surface. This is also an interesting morphology that could be investigated by shallow seismic reflection methods in order to image the water table.

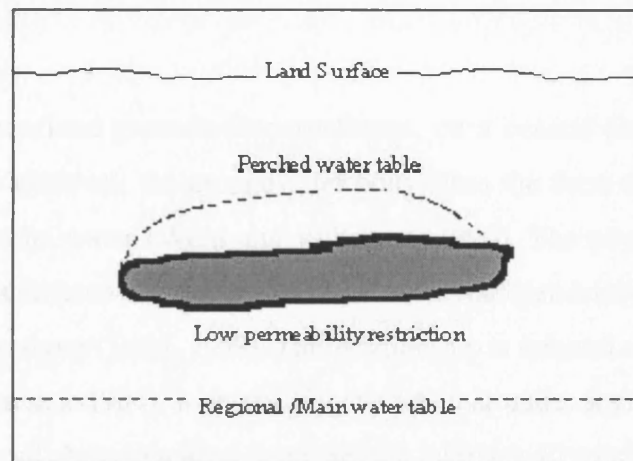


Figure 3.4 schematic presentation of a perched water table
(after Domenico and Schwartz, 1990)

3.4. Water Table

As mentioning in section 3.1 the water table (adopted definition from Price, 1985) is the surface on which the fluid pressure p in the pores of a porous medium is exactly atmospheric (Freeze, 1979). This surface is the lower of the range that separates between the saturated and unsaturated zones. Figure 3.5 indicates the water table position. Shallow seismic reflection method may image the base of the capillary fringe as the water table. This assumes that this base forms a well-defined planar boundary within a formation, which has a relevant coefficient reflection that varies with the porosity of the rock that should be $> 10\%$.

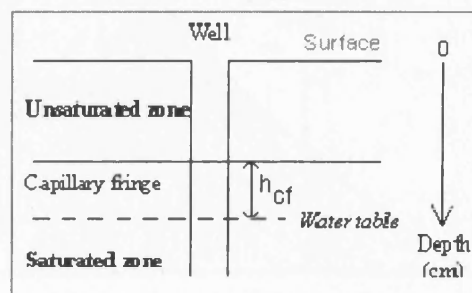


Figure 3.5 schematic of the water table position
(after Freeze and Charry 1979).

3.5. Coastal Aquifer

In normal unconfined groundwater conditions, on a coastal plain with a water table sloping towards sea level, the groundwater body takes the form of a lens of fresh water 'floating' on saline water (Ward and Robinson, 1990). The physical controls on the position of the saltwater-freshwater interface were independently established by Badon Ghyben and Herzberg (Todd, 1959). This relationship is referred to as the Ghyben-Herzberg principle (LaFleur-1984). It shows (Figure 3.6) that under static condition each meter of fresh water head above sea level depresses the interface 40 meter below sea level according to the relationship:

$$\rho_s g z_s = \rho_f g (z_s + z_w)$$

$$z_s = \frac{\rho_f}{\rho_s - \rho_f} z_w \quad 3.3$$

Where ; g : gravity constant, z_w : freshwater head above sea level

z_s : distance of interface, ρ_f : freshwater density = 1000

ρ_s : sea water density = 1025, $z_s = 40 z_w$

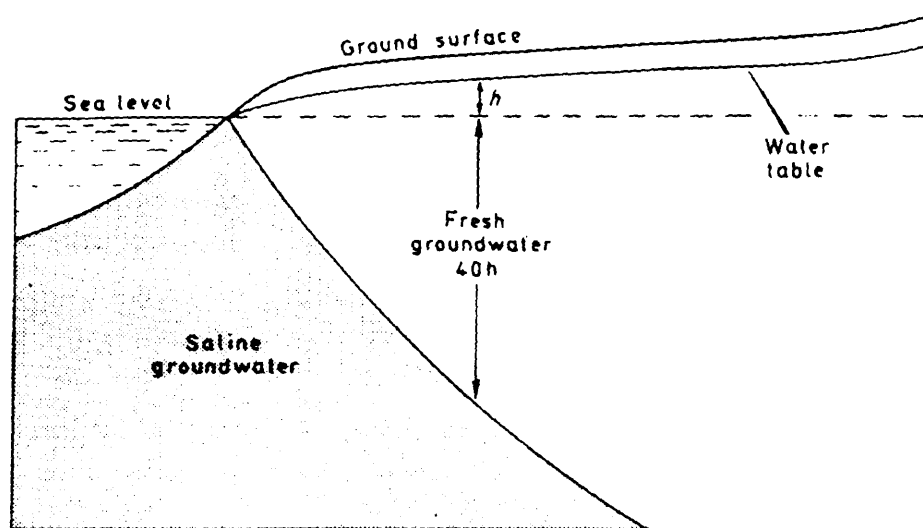


Figure 3.6 simplified diagram showing the position of freshwater-saltwater interface according to the Ghyben-Herzberg principle as hydrostatic relationship (after Ward and Robinson, 1990)

Freeze (1979) expanded that concept, stating that the interface line is not static as assumed by Ghyben-Herzberg principle but is a dynamic relationship with the position of the interface controlled by the head distribution (Figure 3.7). The position of the interface can be determined by flow-net analysis in which the change in freshwater head, Δh , between two adjacent flow lines is the control. Therefore, the interface is deeper than it would be if a static condition existed. This analysis assumes that the interface is a sharp boundary and that the saltwater is not flowing.

This also implicitly assumes that the fresh-salt water interface slopes downward from the coast (Todd, 1959). The interface shape and interface slope can only be inferred for the case where flow occurs in the fresh-water zone. In this case Darcy's law shows:

$$\sin \delta = \frac{dh}{ds} = \frac{v}{K} \quad 3.4$$

where; δ = water table slope,

v = velocity of flow, K = Permeability

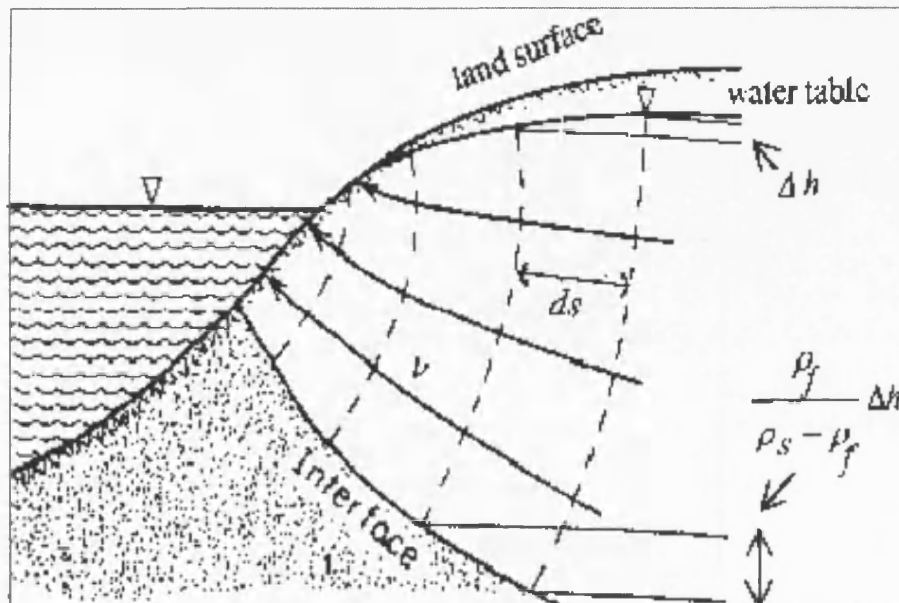


Figure 3.7 saltwater-freshwater interface under conditions steady state seaward flow (after David and DeWiest, 1966)

Water table elevation decreases in the direction of flow along the slope. Consequently, according to the equation 3.3 the fresh-salt water boundary (interface) must rise. The slope of this boundary is:

$$\sin \epsilon = \frac{\rho_f}{\rho_s - \rho_f} \frac{dh}{ds} = \frac{\rho_f}{\rho_s - \rho_f} \frac{v}{K} \quad 3.5$$

Velocity of flow, v , increases with distance, because the slope of interface is convergent, and therefore, the magnitude of the slope increases. The result is a concave interface with respect to fresh water, as shown in figure 3.7.

The length of the intruding sea water wedge, L , is related to Ghyben-Herzberg principle. A salt-water wedge must exist at the intersection of an aquifer with the ocean. It is assumed that if a seaward fresh water flow, q , per cubic metre of ocean front exists, then the approximate formula for a confined aquifer is given from Darcy's law:

$$q = \frac{1}{2} \left(\frac{\rho_f}{\rho_s - \rho_f} \right) \frac{K b^2}{L} \quad 3.6$$

where b is the thickness of aquifer as shown in figure 3.8.

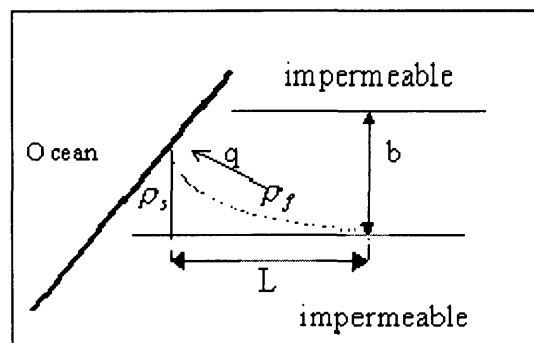


Figure 3.8. salt water wedge in a confined aquifer

This formula indicates for uniform aquifers and fluid conditions that the length of the intruded wedge is inversely proportional to the fresh water flow. This formula can also be applied to unconfined aquifers by replacing b by the saturated thickness, providing the flow does not deviate greatly from the horizontal.

There are five methods used to control the sea water intrusion (Todd, 1959):

1. Reduction and/or rearrangement of pattern of pumping draft.
2. Direct recharge.
3. Increasing of pumping from the area of saline-water through paralleling the coast.
4. Maintenance of fresh water ridges above sea level along the coast.
5. Construction of artificial subsurface barriers.

3.6. Sea Water Intrusion

Accidental contaminations can occur in coastal areas (Karahanoglu, 1997). In coastal areas, fresh water derived from infiltration can overlie saline water in such a way, it causes flow to occur from land to sea (Mahesha & Nagaraja, 1995). That is, the pressure of fresh water exceeds the pressure of the denser, salt water at the interface line.

If water is pumped, the lowering of the water table results in the upcoming of the interface, and salt water may be drawn into the well (figure 3.9). When the pumping takes place, the lowering of water table induces a corresponding rise of the interface. Therefore saline water migrates inland and may eventually reach the well. In practice the interface surface is not sharp, but it is affected by both dispersion and diffusion.

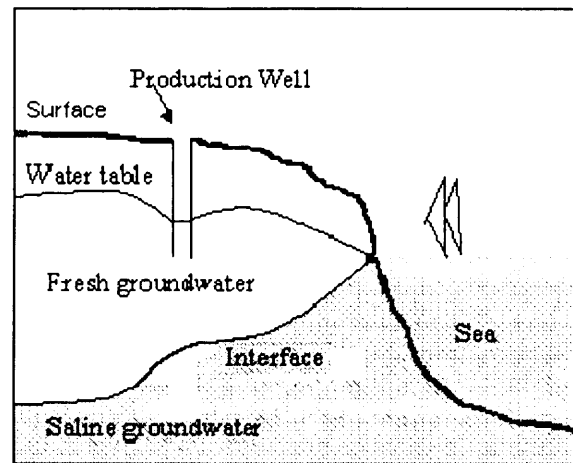


Figure 3.9 production well can change the water table and interface line

Sea water intrusion is becoming a significant problem in coastal plains, but no case studies, have been carried out to accurately detect sea water intrusion problems. Overmeeren (1989) made an equivalent study case: Aquifer boundaries explored by geoelectrical measurement in the coastal plain of Yemen. He used the resistivity model to provide the distribution of an aquifer at a large scale (0 - 560 m depth range). By using the model and resistivity measurement as input, with fixed aquifer thickness, a transition zone between fresh and saltwater (as called interface line) was produced of about 4 km width. However, this result is too vague, because geoelectrical methods have a low resolution. This method is not appropriate for accurately detecting sea water intrusion problems.

A long term aim of this project is to assess the application of seismic methods to this problem (Paiene et al, 1997). Direct detection of an interface is unlikely, but high resolution imaging of geological structures and direct determination of water tables would be of great assistance to the overall problem, especially when combined with modern multi-electrode resistivity electrical or other EM methods for determining good conductivity.

Chapter 4

SYNTHETIC SEISMIC MODEL

4.1. Introduction

The main idea of this chapter is to construct the ideal model of a seismic record using a simple mathematical formula on a spreadsheet and using a convolution principle on Promax. We also aim to produce the optimal sequence of data processing based on borehole information and field records (Chapter 5). The purpose is to compare between the ideal model and the real field record in order to deal with processing and interpretation problems.

There are two reasons why the Edwinstowe field site was chosen to test the detection of water table. It has simple geology and good geological control from the British Geological Survey (BGS). The borehole record informs us that the geology is near homogenous and beds are horizontal with a weathered sand layer zone that was known to give poor transmission of high-frequency seismic energy. Near-surface geology consists of Sherwood Sandstone. A borehole (approximately 700 m NNE of the site) confirms a thickness of 120 m of sandstone underlain by Mercia Mudstone. The Sherwood Sandstone has porosities ranging from 14 to 36%, the water table in borehole is 31 m bgl. (Below Ground Level). The ground surface is very dry and grassy.

In order to design an optimal survey geometry, synthetic travel-time data was generated using an excel spreadsheet. To determine the seismic processing sequence, synthetic waveform data was generated within the Promax processing system, and processed to a stacked section. Velocities for different wave phases were determined by a preliminary survey. Figure 4.1 illustrates the simple model that was generated from BGS information and interpretation of field survey. Using an interactive process, the interpretation result of processing in chapter five will also be used to construct the synthetic model.

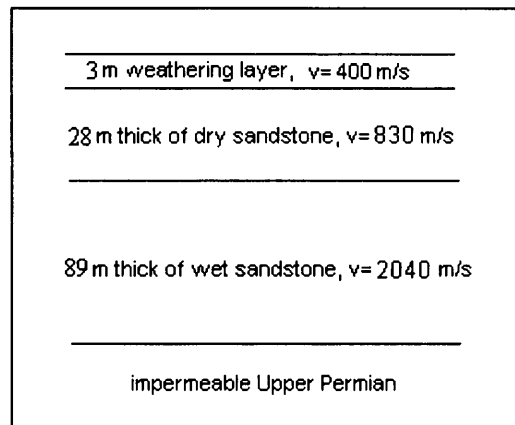


Figure 4.1 illustration of the geology of the field site with 32 % porosity of Sandstone

4.2. T-x Seismic Curve Model

There are at least six kinds of travel time for the simple model required to describe the shot record for one boundary layer; the direct wave and refraction wave as first arrival, reflection wave, surface wave (Rayleigh wave, S_R) as ground roll, and air wave. With a depth target of about 30 m, the window is chosen in range that includes the critical distance (x_c) which has optimum energy. This is illustrated in figure 6.3 (a) no 1 & 2 that the P-wave reflection coefficient decreases for increasing angle of incidence or offset. Although there are some noises in the critical distance, ground roll and air-wave, the f-k filter and Air Blast Attenuation in the processing will overcome these problems.

Referring to the interactive interpretation of the field survey, the P-wave velocity of a water saturated sandstone as V_2 is 2040 m/s (for 32% porosity, see Appendix A). V_1 is 830 m/s (see table 5.1) and is the wave velocity of unsaturated sandstone. The Root mean square (rms) velocity of weathering layer and dry sandstone is 820 m/s for a 3 meter thickness of weathering layer with a velocity of 400 m/s. Using the interval velocity, the critical distance (x_c) has been calculated at around 30m, comparable to the 24° (using formula 2.22). Therefore, the optimal field spread for 48 channels is an End-On spread with a trace interval of < 1.5 meter (in this case we used 1 m. interval).

Figure 4.2 shows the t-x curve model based on figure 4.1. The reflection and refraction record include the reflector of a sandstone base at 120 m depth. First arrival and ground roll, as surface waves, are decided from the field reflection and refraction record. For direct waves the velocity of P-waves are equal to 400 m/s as a travelling wave in the weathering layer. Ground roll is produced using a wave velocity of 170 m/s. Meanwhile, the airwave is generated using a sound wave velocity of 330 m/s. Two reflections and a refracted wave model are produced using the formula mentioned in section 2.3. Field parameter assumptions have been made i.e. the water table location (calculation provided in appendix C).

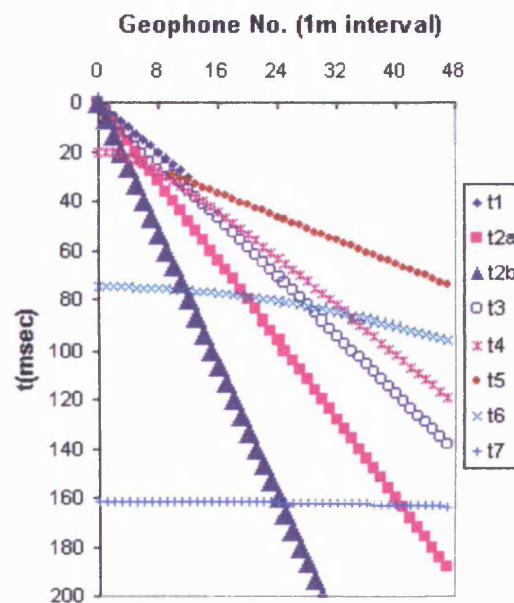


Figure 4.2. T-x seismic curve model of shot record. (t1: Direct wave; t2a: Ground roll, $V_R = 300$ m/s; t2b: Ground roll, $V_L = 170$ m/s; t3 : Air Wave $V = 330$ m/s; t5 : refractor; t4 : first boundary reflector; t6 : second boundary reflector /water table, t7: third reflector as lithological boundary)(see appendix C).

4.3. Seismic Convolution

Convolution is the change in waveshape as a result of passage through a linear filter. If a waveform $g(t)$ is passed into a linear filter with the impulse response $f(t)$, then the output is given by the convolution operation of g with f (Sheriff, 1984). In the seismic reflection method, energy from the source (as input) passes through the Earth as a linear layered system and the output is a signal received by the geophone. Mathematically, there are three components used to produce the signal as the output $s(t)$ of the convolution operation: wavelet $w(t)$ as signal source, reflectivity function $r(t)$ as the Earth system filter, and noise function $n(t)$. This is given by:

$$s(t) = [w(t) * r(t)] + n_1(t) + n_2(t) \quad 4.1$$

Where: $n_1(t)$ = assumed coherent noise as separated process of convolution

$n_2(t)$ = additive noise or random noise

$*$ = convolution operation

4.4. Concept of Seismic Model

One implication of the seismic model is that the process of convolution can be applied as distributive with respect to addition (Robinson, 1980). Therefore, equation 4.1 can be written as :

$$s(t) = [w(t) * n_c(t)] + [w(t) * r(t)] + n_2(t) \quad 4.2$$

where $n_1(t) = w(t) * n_c(t)$, the convolution of coherent noise

$n_c(t)$ = convolution factor of coherent noise

This expresses that the signal is produced by twice ($r(t)$ and $n_c(t)$) process of convolution plus random noise (n_2). For a simple model, we can only use one process for reflection coefficient $r(t)$, and assume $n_2(t) = 0$. Therefore we have a simple

mathematical formulation (Yilmaz, 1988):

$$s(t) = w(t) * r(t) \quad 4.3$$

This is the continuous, or analogue formulation. The other formulation is in the discrete or digital mathematical formulation. In this case, the integral convolution can be seen as a multiple and summation operation, then called discrete convolution:

$$s(J\Delta t) = \sum_{k=1}^N w(k\Delta t)r((J-k)\Delta t) \quad 4.4$$

or

$$s(J) = \sum_{k=1}^N w(k)r(J-k) \quad 4.5$$

where Δt is the sample rate and J the sampling number.

The general assumption in this model is that the wavelet reflected from an acoustic impedance discontinuity has the same waveshape as the incident wavelet. This means that a seismic trace is simply a superposition of individual reflection wavelets. This is a basic feature of the convolution model that reflection processes are being considered as a 'filter'. One-dimensional synthetic seismograms assume that raypaths are vertical and interfaces horizontal. Thus, reflection and transmission coefficients are for normal incidence (equation 3.25 and 3.26). In this case, diffraction and other wave modes such as interference are ignored, but multiples are still included.

4.5. Synthetic Seismic Model

Promax version 7.2 includes a process to produce a synthetic trace by the process of convolution. The idea of generating the synthetic model is to show that the water table acts as a reflector of in the field example and can possibly be identified. The parameters used are those which have been discussed and used in the t-x model.

A set of shot records will be generated as discussed at section 4.2. Data will be processed using the normal sequence used in shallow seismic processing. Finally it will be proven that the water table may be imaged very clearly by this method.

4.5.1. Synthetic Shot Records

The synthetic shot record as a model, is more complex than t-x curve but both have the same principle of wave propagation in that the travel time of each signal is a function of x (offset) and the wave velocity in every layer. The model is very similar to the real record as the shot record. It requires a field parameter such as sample rate, length of trace, frequency window, and so on. The essential difference from the t-x model is using the coefficient reflection (R_c) for reflected signals from each boundary. It expresses the contrast impedance of that boundary. The model uses $R_c = 0.15$ for water table in pure sandstone 6% porosity (figure 2.5). This is the minimum condition where the water table as a boundary can be recognized, larger porosities give a clearer boundary.

Figure 4.3 shows the synthetic shot record. The parameters of the synthetic waveforms have been taken from real records listed in table 4.1. The synthetic geometry is listed in table 4.2.

Table 4.1 synthetic shot record parameters as specified for Promax synthetic trace generator for horizontal layers, additive as random noise with S/N ratio of 2 (V & R_c of water table are as mention in appendix A)

Type	Time (ms)	V (m/s)	Amp. (R_c)	Length (ms)	f1 (Hz)	f2 (Hz)	f3 (Hz)	f4 (Hz)
Direct wave	0	400	0.8	35	25	50	150	300
(layer I & II)	17	830	0.8	35	25	50	150	300
Air wave	0	340	0.4	25	50	150	300	700
Ground Roll (L)	0	170	2	250	20	30	60	70
Ground Roll (R)	0	300	2	150	20	30	60	70
First reflection	20	400	0.54	35	8	40	300	400
Water table	72	790	0.15	25	8	40	300	400
Third reflection	161	1780	0.29	30	8	40	300	400

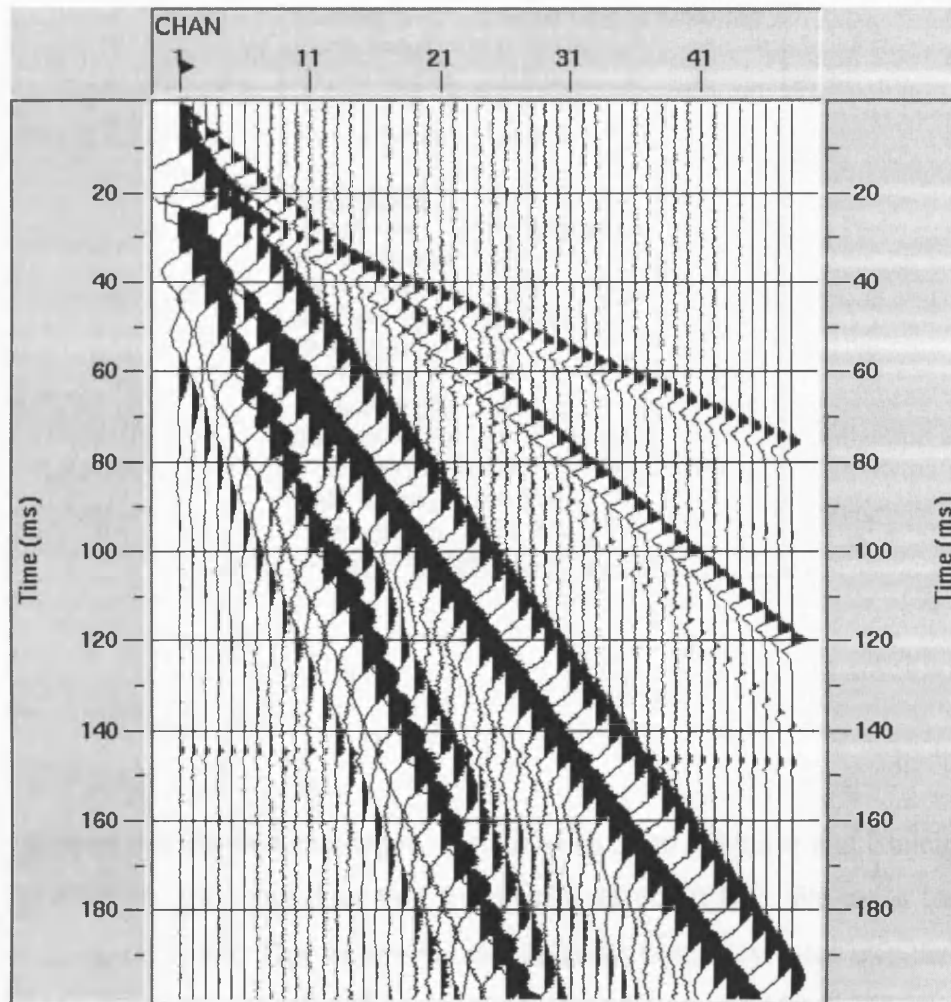


Figure 4.3 Synthetic shot record

Regarding the synthetic shot records produced, there are two types of signal; coherent noise and reflection. Table 4.1 shows that the direct wave, air wave, and ground roll are coherent noise. The travel times of these events should be zero because we used zero offset. All parameters of these noises are taken and modified from the field shot record.

Table 4.2. Simulated shot record parameters

<u>SITE</u>	
Site Name	: Edwinstowe Model
Geology	: Water table in Sandstone
Depth Range	: 0-150 m
Record length	: 200 ms.
Sample interval	: 0.25 ms.
Record Filters	: open
<u>GEOMETRY</u>	
Spread type	: On-end spread
No. of channels	: 48
Ch spacing	: 1 m
Shot spacing	: 0.5 m
Shot depth	: 0
No. of shot	: 87
Shot offset (X)	: 0
Shot offset (Y)	: 0
Nominal Fold	: 48
CMP spacing	: 0.5 m
Length of line	: 91 m
Length of Max.Fold CDP	: 22.5 m

Ground roll displays two types of velocity as Love (L) wave and Raileigh (R) waves (170 & 300 m/s) and dominant frequencies of 40-50 Hz. We use a frequency range of 20-30-60-70 Hz. This means that the dominant frequency value is in the middle of 30-60 Hz. The wave period is about 50 ms. In this case we use 150 & 250 ms as three to five cycles as there is no facility to multiply the cycle. The parameters of direct wave are taken from the field and air waves and are almost the same for all fields.

There are three reflection events. The first reflection expresses the boundary between the weathering layer and dry sandstone. The second is the water table boundary in 6% porosity pure sandstone and the third is boundary between saturated sand and impermeable Upper Permian which is a more compacted rock at depth. All velocities values have been mentioned in figure 4.1. The range of frequency of 8-40-300-400 Hz are normally used as a range of shallow seismic surveys that have dominant frequencies of about 100Hz. The amplitude is the coefficient reflection that is taken and has been discussed at the section 2.

Regarding the synthetic geometry input, table 4.2. shows the technical description of synthetic seismic data. On-End spread at a minimum of zero offset has been used. The number of channels is 48 and a sample rate of 0.25 ms is used to duplicate the field record. Forty-eight channels are adequate to increase the fold of cover to 2400% when using the same interval for the source and receivers. To increase fold of cover and length of line we need a large number of sources.

When we use 87 sources with 0.5 m interval we obtain a 91 m length of line, where the length of CMP is 22.5 m with 24 fold in 0.25 m interval, than it is comparable to 48 fold in 0.5 m interval. This synthetic geometry input is exactly same as the real survey in the next chapter. Figure 4.4 shows that the full-fold CMP number is no. 94 until 172.

Note that the synthetic shot records are produced to show that the field site as expressed by the model can ideally be examined in order to identify the water table. This is clearly sufficient only by one shot record (1 fold of cover). However, we still need further data processing to produce a final section and to acquire complete information.

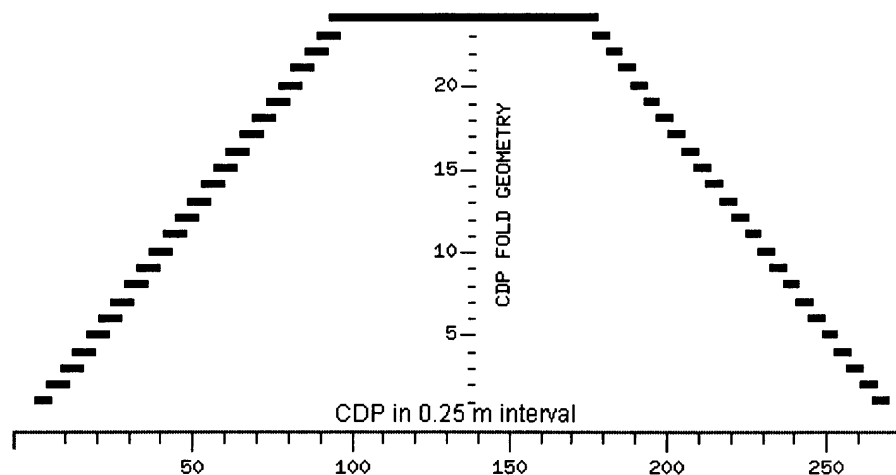


Figure 4.4 the fold of cover in CMP

4.5.2. Processing

4.5.2.1. Pre-processing

In the pre-processing sequence, all steps except the killing trace, as mentioned in section 2.3.2.1 handled very well. This pre-processing will start from geometry input and will finish with First Break Picking. After completing the pre-processing, we can check the graphs in the database such as CDP to show fold of cover profile (figure 4.4).

4.5.2.2. Filtering

The main purposes of data processing are to filter traces and to construct the final seismic section. Filtering includes removing or reducing refraction, airwaves, direct waves, and ground roll. An f-k filter is applied to remove ground roll. Figure 4.6 (b) shows that the 'f-k accept' part is very clean from (c) 'f-k reject' as ground roll. This means that the filter window is very appropriate and is taken from the synthetic shot record parameters. The key stage of this filter choosing the filter window constructed from the maximum and minimum values of wave velocity and frequency. For this process, it is easy to find filter windows from synthetic trace parameters or from doing f-k analysis. Figure 4.5 shows the f-k analysis screen. It is suggested that the energy below 60Hz is mostly noise. Furthermore, there is very clear reflection energy in the middle of wave number (vertical red line).

The primary concern in this data is ground roll, because this is the main coherent noise which must be removed. The correct treatment of this noise is critical in avoiding subsequent artifacts of the processing. When we successfully remove this noise it is likely to reveal boundary layers from within the covered ground roll. As illustrated in the figure 4.6; (a) the original synthetic shot record, reflected signals are covered by ground roll, while (b) and (c) are respectively the signals and the noise after separation.

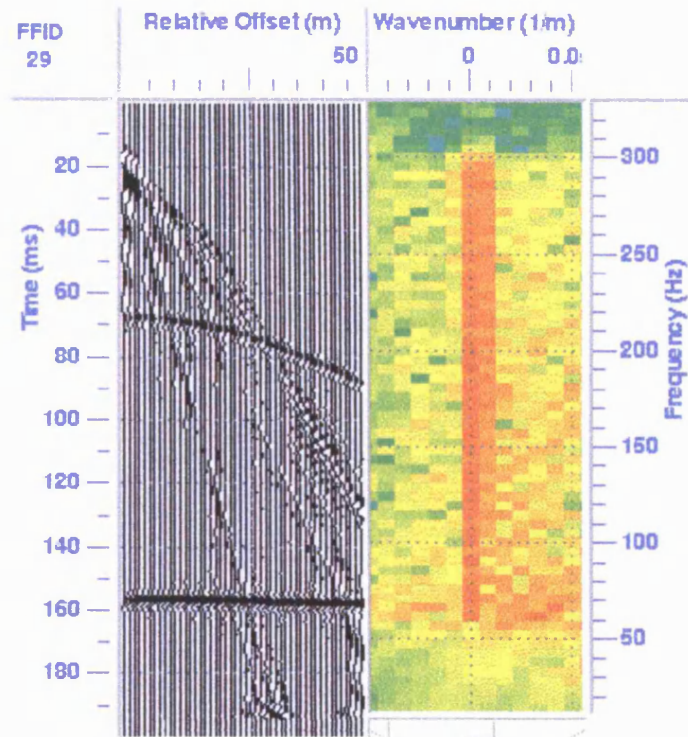


Figure 4.5 the f-k analysis screen.

The result is excellent i.e. ground roll is properly removed. The reflectors appear clearly as seismic boundaries even it is very difficult to identify in the original shot record in small offset. Therefore, in this case (synthetic data) f-k filter is very effective because the model is ideal i.e. the layer is homogenous, and the physical parameters are exactly known and can be used to construct the filter window. For real shot record, the filter window should be analysed and chosen from the f-k analysis and called artifact processing, as one of the limiting parameters mentioned previous section.

Secondly, air wave energy is normally solved by Air Blast Attenuation (ABA). This reduces the amplitude of air waves or the sound that propagates when the plate is beaten by hammer. The important parameters in this process are velocity and the range of frequency. There are easy to find because there are typical of a sound wave in the air. In this case we can refer to airwave parameters of synthetic shot records, where the air wave parameters have been covered by f-k filter parameters.

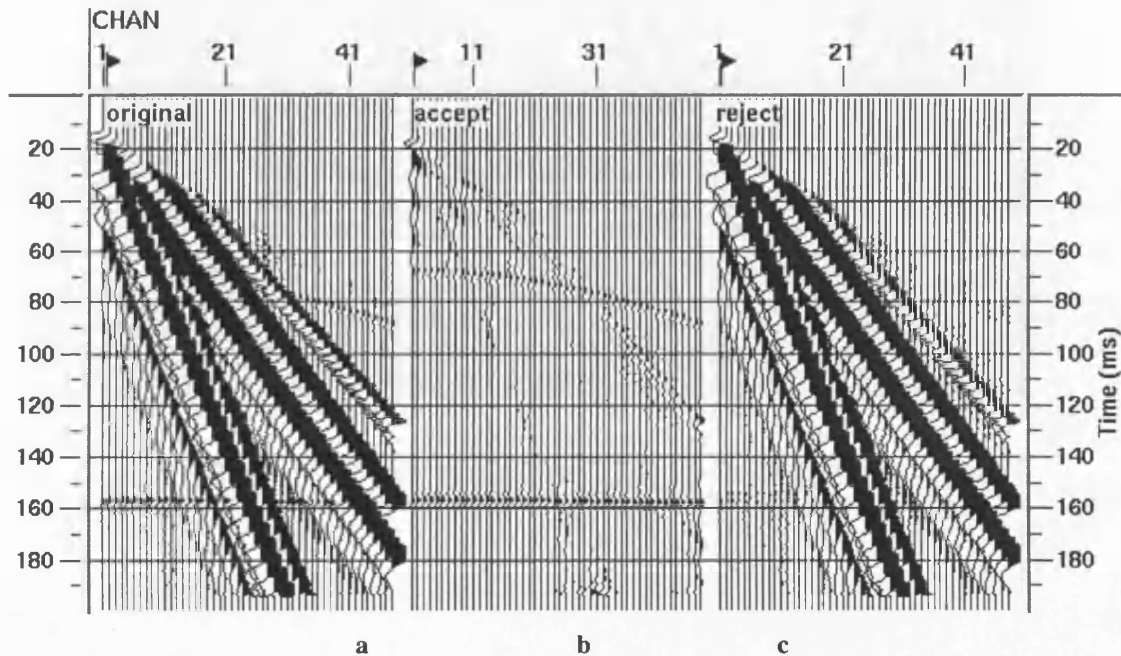


Figure 4.6 (a) is the original shot record and (b) & (c) are after f-k filter $(-1100, 625, 60, 300)$ for accept and reject.

Removing refraction and direct waves is normally performed using top muting in the pre-processing. For synthetic shot records, the noise before first arrival is not significant because of assumed homogenous nature of the surface layer, this can be achieved by using muting in the process of NMO correction. Figure 4.7 (c) shows the clean zone due to removing the refraction and direct wave using top mute.

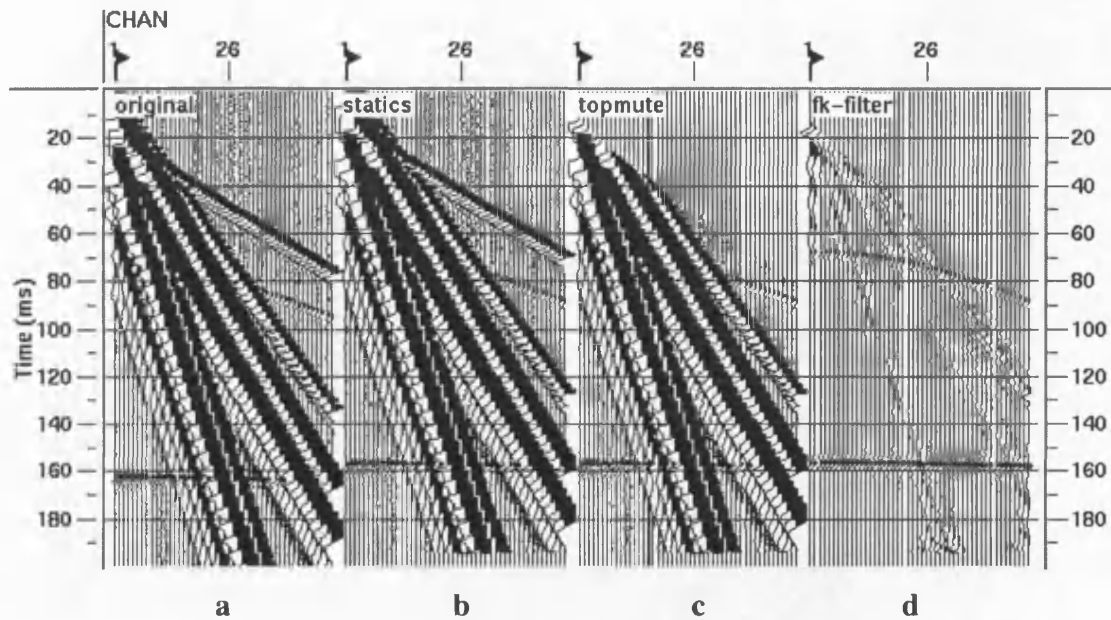


Figure 4.7 (a). synthetic shot record, (b). after refraction static correction, (c). after top muting, (d). after f-k filter (-1100,625,60,300) for accept.

4.5.2.3. Correction

The near surface velocity often has a large and unexpected variation. Therefore, static correction and velocity analyses are extremely important, in reducing the velocity variation effects of the near surface and elevation. Static correction for this model uses the refraction method and the interactive artifact process described in 2.3.2.2 and Appendix B. Again, it also uses similar parameters of static correction of the real processing described in the following chapter. Figure 4.7 (b) shows the result of this static correction. It is clear that most of signal shifts about 7ms. It is reasonable for the medium that the velocity at near surface is very low, and changes dramatically between the weathering layer and dry sandstone. In addition, the existence of random noise contributes to the static error. Therefore, the correction order of about 7ms is a compensation of the replacement of the velocity of weathering layer by dry sandstone velocity.

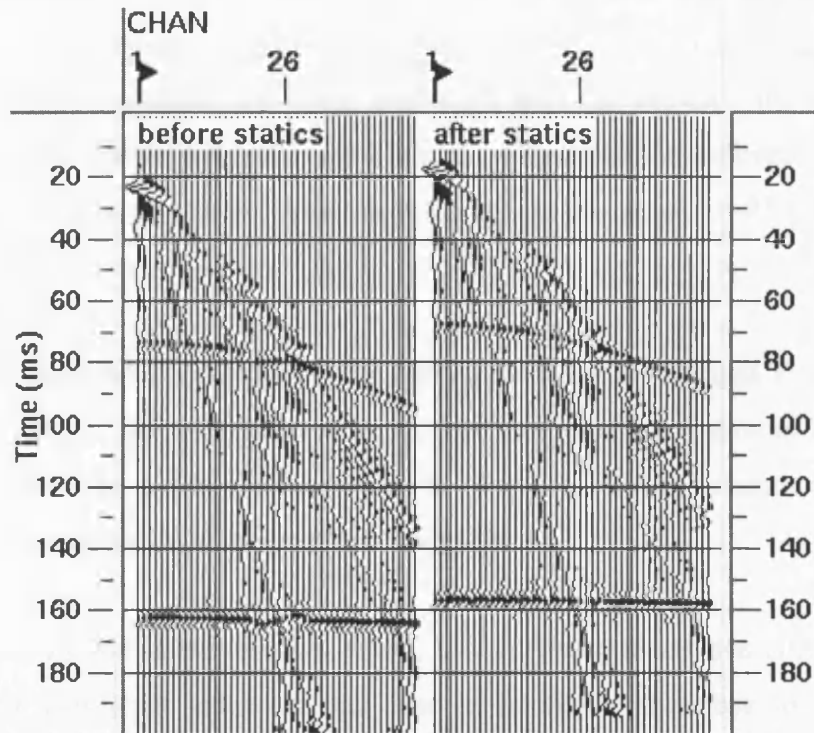


Figure 4.8 the statics correction is very effective to smooth the reflector from the uncertain undulation, adding the Hand Statics error (Trace no. : msec of 21:2 and 28:-2, interpolation between them)

Figure 4.8 shows that the statics correction is very effective in removing these errors. This is illustrated in the panel above after static correction, which followed adding the statics error using the Promax “Hand Statics” processor. This processor is to give several msec to some traces that produce uncertain signals as static error. In this case we put 2 msec for trace no. 21, -2 msec for trace no 28, and interpolation for trace no 22 until 27 (more clear at 161 msec in panel before static). Therefore, the refraction statics have been performed then it is applied as the refraction static correction. The result of this correction is clearly shown by panel after statics that the additional signals were corrected properly.

The Normal Move-Out (NMO) correction is applied. The key factor of this correction is choosing the NMO-velocity. NMO correction for the initial brute stack uses the velocity estimation, while the final stack may use velocity function from the velocity analysis. The velocity estimation is formed based on the parameter of synthetic shot records. This case uses NMO that uses the range velocity of : 0-830, 70-860, 150-1600, 200-1800 m/s.

From the synthetic parameter, the maximum offset is 48m and T_0 for the water table layer is 70ms. The NMO correction (Δt) is very significant, around 20ms for the maximum offset. The equation of correction is given as $x^2/2V^2T_0$, where x for offset, T_0 vertical travel time, and V velocity at boundary layer.

Figure 4.9 shows the result of NMO. It is very easy to say that NMO correction gives a very significant value, in this case to normalize the trace to move up as correction of offset distance that is the most important correction in the seismic processing. However, we still need velocity function for time/depth conversion. There are two options; firstly we can obtain velocity function from boreholes or alternatively, from field records. For the next step, the picked points will be used for depth/time conversion, and stack velocity.

4.5.2.4. Stacking

As mention before, stacking combines all individual CMP traces in each gather into one trace, which will then become part of the final section. Figure 4.10 shows the final stack with CMP line that indicates fold of cover. It is clear that the water table boundary reflection is located at about 30 m. Table 4.3 describes the list of processing of synthetic shot record.

Table 4.3. List of processing synthetic shot records

Trace muting	: Top muting
Gain and Polarity	: True Amplitude Recovery at only before CDP/Ensemble stack
Bining	: Matching pattern number using first live channel and station.
Velocity Analysis	: 0-830, 70-860, 150-1600, 200-1800
Filter:	
-F-k	: F-k analysis; -1100, 625 m/s : 40, 300 Hz. (accept)
No. of CMP's	: 268
CMP Spacing	: 0.25 m
Stack	: Mean stack, all traces.
Output	: Variable area wiggle plot

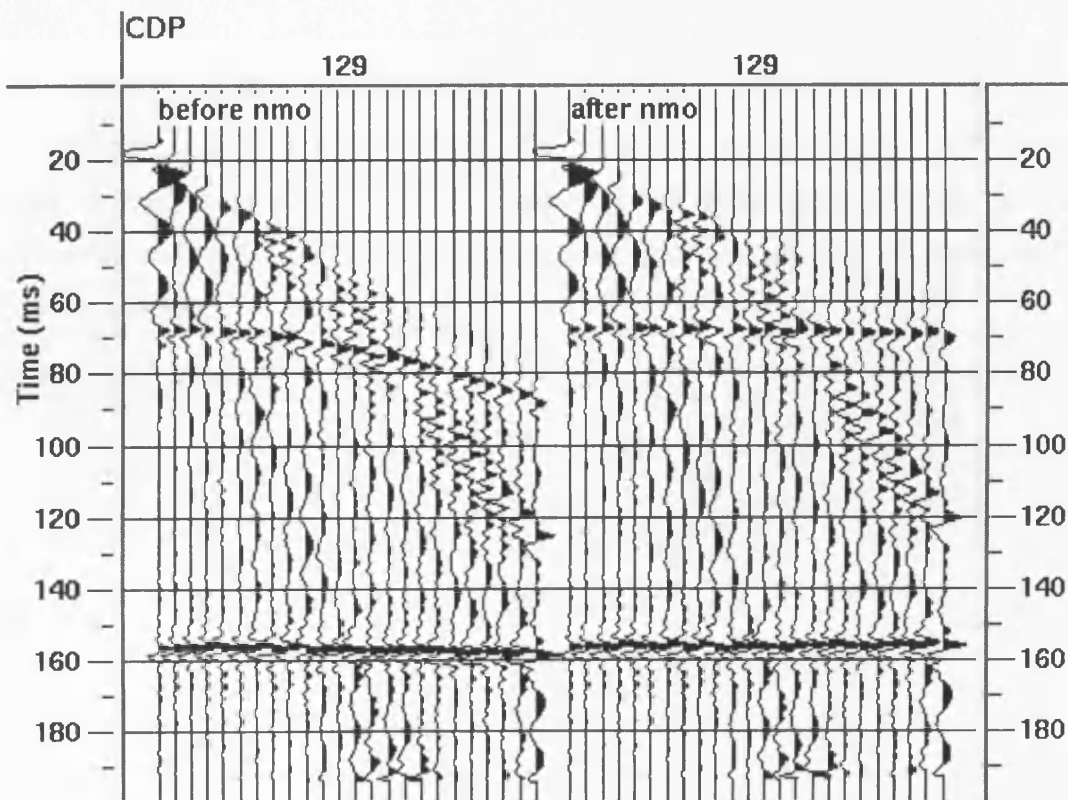


Figure 4.9 The result of nmo

4.6. Discussion

The only point that should be made is that synthetic seismic shows that the water table can be clearly identified as a reflector. Although this needs further assessment on how well synthetic data will actually approximate field data. In particular, there are two assessments dealing with random noise and coherent noise. Stacking from fold of cover of 2400% is effective in reducing the random noise that has been applied to the value of 2 as signal-to-noise ratio. The real data will have variable random noise that will degrade the f-k filter, refraction statics and velocity analysis. Because of this, real data will not process as well as the synthetic data. The amplitude and velocity of ground-roll as coherent noise in synthetic data are estimated and calculated from the dominant parameters of real data.

Synthetic shot records prove that the water table in 6% porosity sandstone can reflect significant energy to the receiver. This result is clear even before CMP stacking. Therefore, the expected use of the processing for real fieldwork is to image the water table reflection in a final stacked section with sufficient of fold of cover and an optimum window.

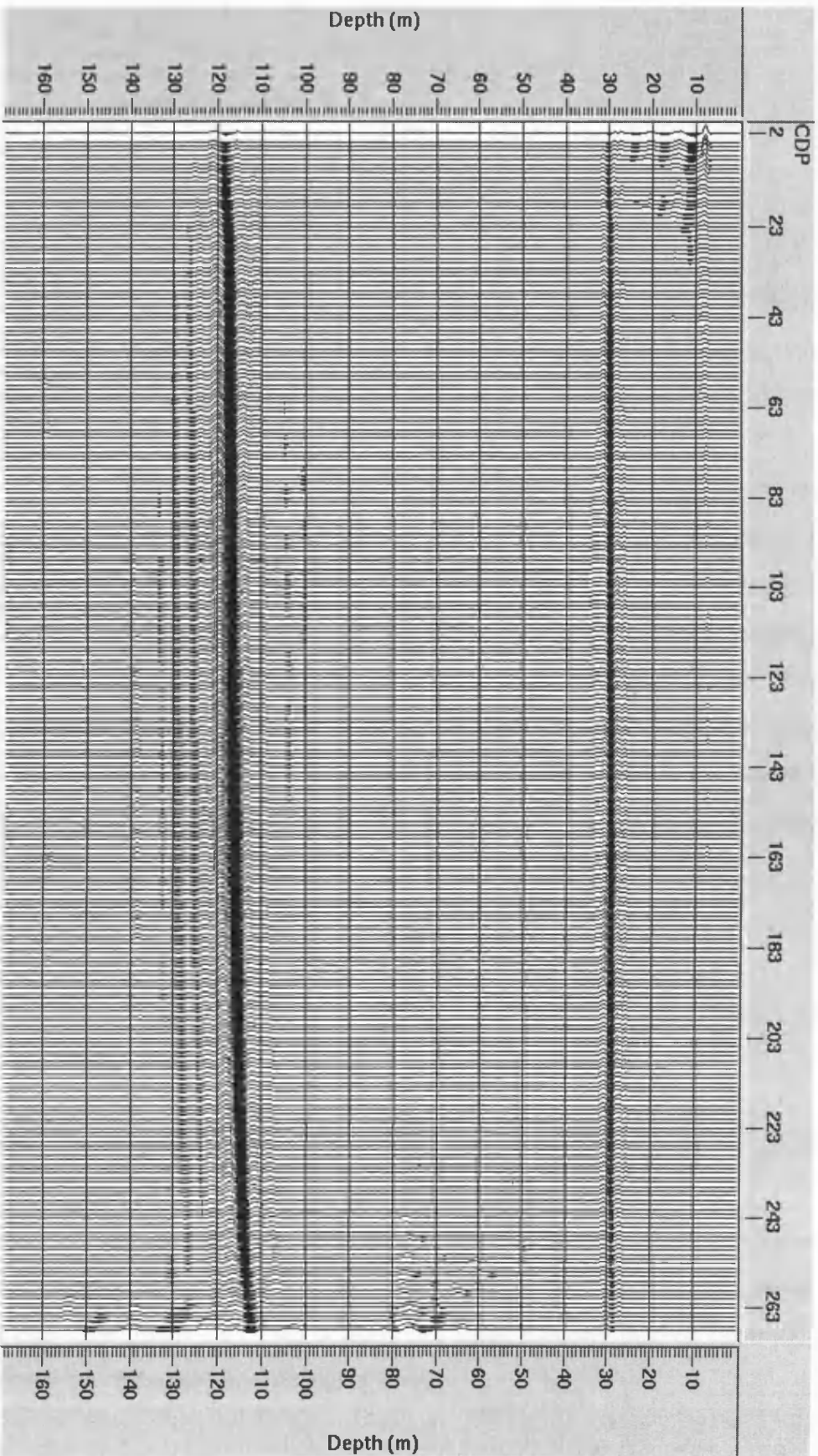


Figure 4.10 the final stack - CDP line in 0.25 m interval

Chapter 5

CASE HISTORY: EDWINSTOWE SURVEY

5.1. Geological Background

This chapter will describe the collection and processing of field data, and examine the method used to detect the water table by seismic reflection. The data was recorded on 3/6/1998 at Edwinstowe, 25km north of Nottingham (SK 630 675).

Figure 5.1 shows the location of field work. The geological background of the location is near homogenous and horizontal, comprising a weathered sand (Sherwood Sandstone) layer that is known to give poor transmission of high-frequency seismic energy. A borehole approximately 700m NNE of the site confirms a thickness of 120m of Sherwood Sandstone underlain by Mercian Mudstone. The Sherwood Sandstone has a porosities of 14 to 36 %. The water table in borehole is 31 m bgl. The ground surface is dry and grassy. Figure 5.2 illustrates the geology derived from the British Geological Survey (BGS) information. Figure 5.3 calculates the arrival times of several kinds of waves for this geological model.

5.2. Acquisition

5.2.1 Field Data

92 shot records were recorded along the seismic line. The line is located parallel to the western boundary of the field, approximately 10 m into the field, along a bearing of 020° (see figure 5.1). This position lies along a poorly marked car track. Table 5.1 gives a list of the technical details of survey.



Figure 5.1 The plan location of field work.

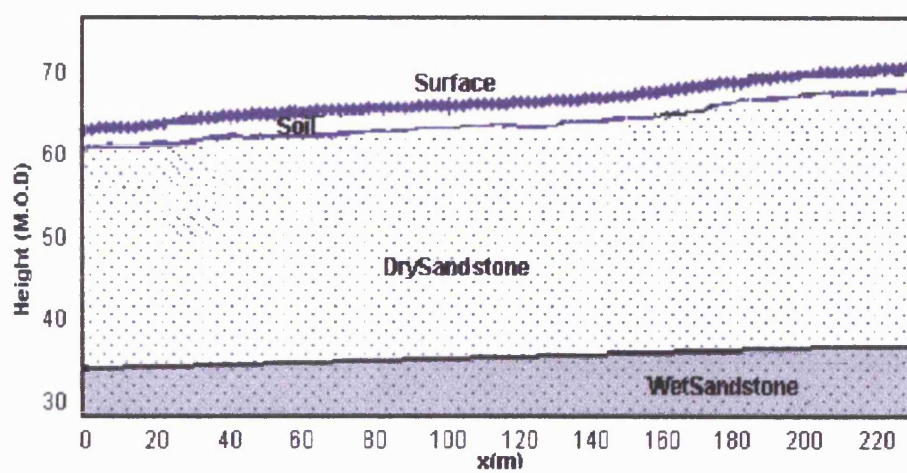


Figure 5.2 Ground model of site

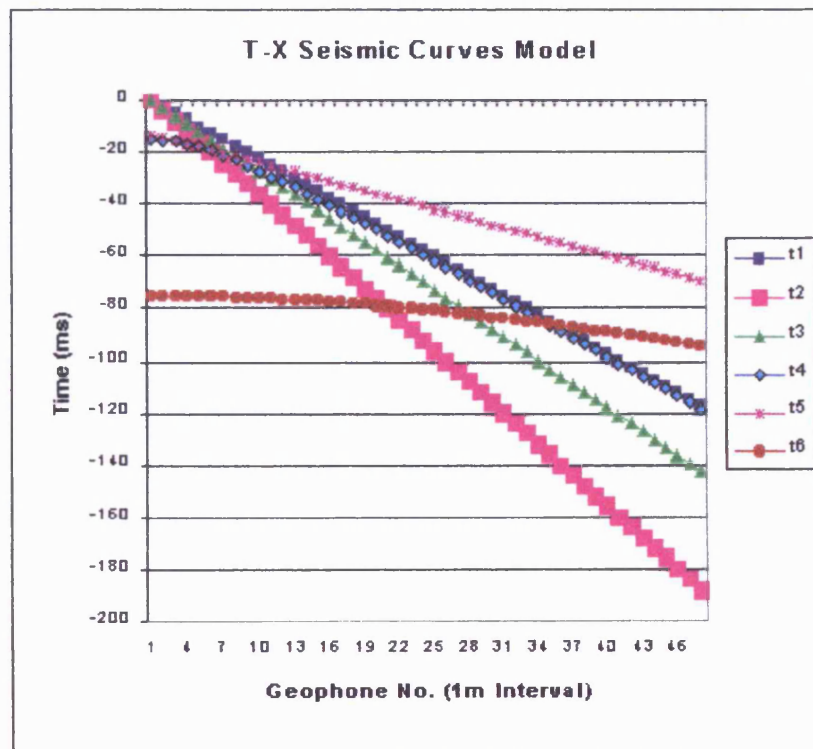


Figure 5.3 t-x seismic curve as the velocity structure interpretation (t1: Direct wave; t2: Ground roll, $V_r = 250$ m/s; t3: Air Wave, $V = 330$ m/s; t5: refractor; t4: first boundary reflector; t6: second boundary reflector /water table)

5 shot records were used for source and equipment tests, and also to record refracted first-arrivals. The 87 remaining shot records were recorded with fixed geometry as shown in Table 5.1 and figure 5.4. Figure 5.5 is a typical example of 87 shot records, no 29. The shot point moved from 0 m to 43 m along the line in a NNE direction, with the geophones leading the shot.

Table 5. 1 Technical description of the seismic data location, Equipment, and acquisition parameters.

SITE	
Site Name	: Edwinstowe
Grid Ref.	: SK 630 675
Geology	: Sherwood Sandstone
Depth Range	: 0-100 m
Date of Survey	: 03/06/1998
EQUIPMENT	
Seismograph	: Bison 9048
Geophone type	: Mark Products 100 Hz., and 125 mm spikes.
Source Type	: Hammer 5 stacks with plastic plate
Record length	: 200 ms.
Sample interval	: 0.2 ms.
Record Filters	: Locut: 128 Hz, Hicut: 1000Hz.
GEOMETRY	
Spread type	: On-end spread
No. of channels	: 48
Length of line	: 91 m
Geophone spacing	: 1 m
Shot spacing	: 0.5 m
No. of shots	: 87
Shot depth	: 0
Shot offset (X)	: 0
Shot offset (Y)	: -0.5 m
Nominal Fold	: 24
Length of Max Fold CDP	: 22.5 m
Nominal CMP spacing	: 0.25 m

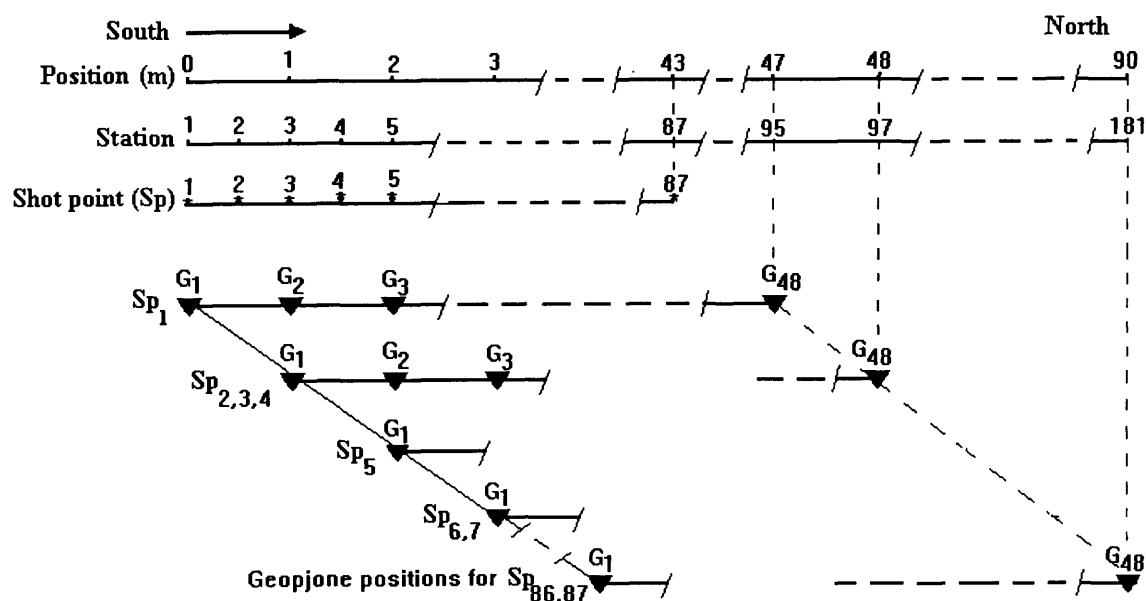


Figure 5.4 Survey lay-out of shallow seismic reflection at Edwinstowe

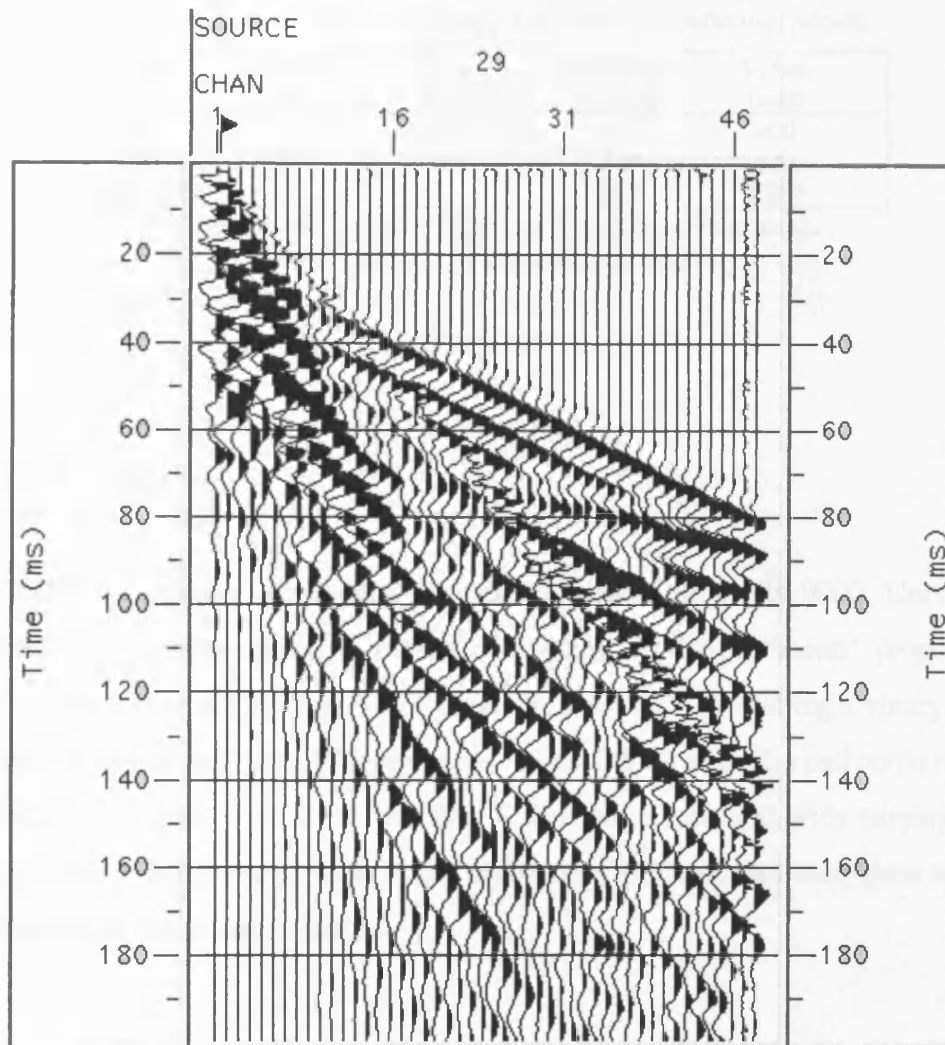


Figure 5.5 the typical example of 87 shot records (no 29).

The reflection data collected is sufficient for the purpose of the survey. The velocity structure was directly determined from 20 records as a representative number of the total 87 records as shown in table 5.2. This velocity structure will be compared to the velocity structures from both refraction and velocity analysis.

Topography was measured by levelling using a Zeiss autolevel to link all station points to a spot height on the adjacent road. The south end of seismic line, $x = 0.00$ m, (starting approximately 40 m from southern boundary of field) is at level 65.60 m while North end ($x = 90.00$ m) is at 67.75 m high which was used as a final datum.

Table 5.2 the velocity structure of the field from reflection records.

Layer	Thickness (m)	Twt (ms)	Depth to top (m)	V. Int. (m/s)
Soil	3.5	20	0	400
Dry Sand	24.7	67	3.5	810
Wet Sand	?	-	28.2	2025*

*From the calculation of Reflectivity Coefficient Vs porosity,
for 34 % porosity of sandstone and RC 0.52

5.2.2. Data Transfer

All field files were recorded on a Bison 9048 (Serial N0: 9000, University of Leicester), then downloaded to a network PC using the Bison “menu” program, then further transferred to the Promax Sun Computer, Sun6 by FTP running a binary transfer. A Promax flow was run using “Floppy input” to read the Bison files and convert them to one SEG Y equivalent file. Since the first 5 files were recorded with varying sample rates, these have been overwritten in the conversion process, and their time scales are now incorrect in the Promax data set.

5.2.3. Data Edit

The first 5 shot records were excluded from the reflection data processing, but they were used to derive a refraction velocity model, shown in table 5.2. Figure 5.6 shows the shot records no 3 and 5 in which the first arrival as refraction waves have been picked (figure 5.7). The apparent velocities have been calculated as listed in table 5.3. All traces of the reflection data set were visually inspected as shot records and about 15 traces with high noise levels were killed. None were of reversed polarity.

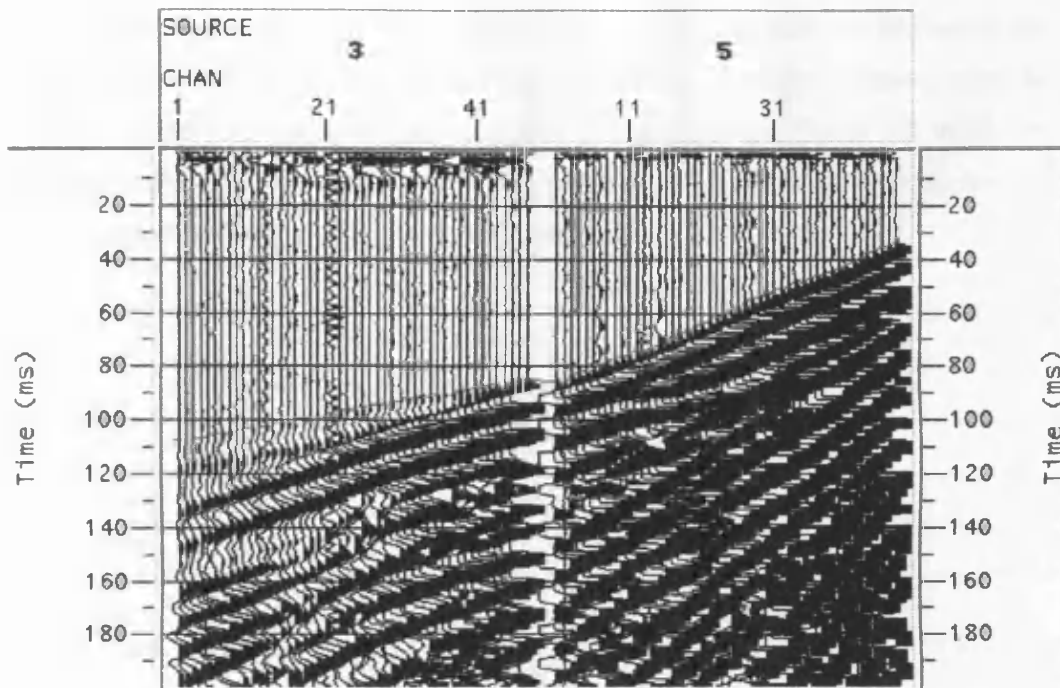


Figure 5.6 shot record no 3 and 5

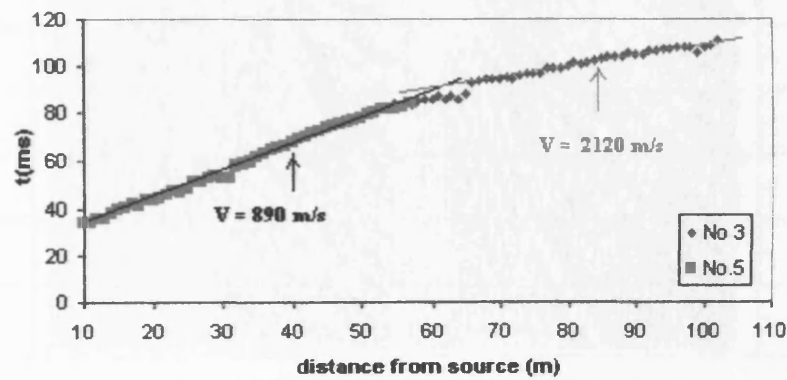


Figure 5.7 t-x curve of shot record no 3 and 5 with apparent velocities

Table 5.3 the velocity structure of the field from refraction records no3 & 5.

Layer	Thickness (m)	Depth to top (m)	App. V. (m/s)	Real V. (m/s)
Dry Sand	22.2	0	890	790
Wet Sand	?	22.2	2120	2015

Real velocity from the calculation of Reflectivity Coefficient Vs porosity give porosity around 35 % of sandstone and RC 0.53

Top muting has also been applied as it is a very effective technique for removing noise before the first arrival time, and also for reducing the strong signal of the first arrival itself as a direct wave and refraction wave. Figure 5.8 show the picking of top mute and the result of this process. There is some remaining energy after top mute that suggested as second and third cycles of first arrival.

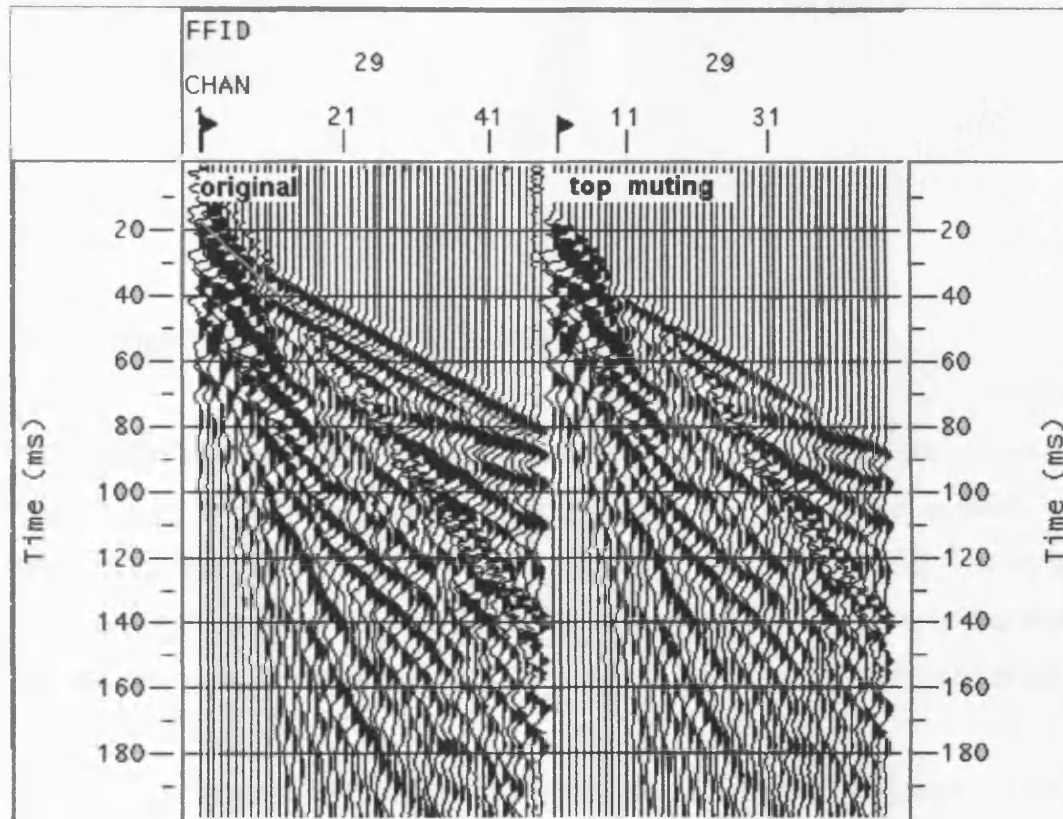


Figure 5.8 the picking and result of top muting process

5.3. Processing

5.3.1. Geometry Input

The seismic line geometry consists of 181 stations at 0.5m intervals (from $x=0$ to $x=90$ m). Shot points extended from 0m to 43m. The geophone spread is shown

Table 5.1 with channel 1 either at the shot point (odd numbered shots) or with offset + 0.5 m (even numbered shot) as the receiver spread rolled along by 1 m every 2 shots.

Using the default binning interval on Promax, resulted in 268 CMP gathers being generated, with maximum fold of 24 at 0.25m intervals and offset increment 0.5m. These were combined for processing to give 48 fold CMPs at 0.5m intervals, where the offset increments in the CMP gather was 1m. This geometry was loaded into the traces header.

5.3.2. First Break Picking and Elevation Static

The general point of static application is to simulate a new set of data from an old set. This is by using a replacement source-receiver surface that is much smoother and usually flat (Hatton,1986). This was described in section 2.3.2.2. The replacement surface is termed the datum, and lies below or above the real surface. In this application we use the datum above the surface, and at the same level as the North end of the line.

Static correction for this data processing uses the refraction first-break method and interactive process. First breaks have been picked automatically by the *First Break Picking* process then edited by hand on an interactive screen. An elevation static has also been applied automatically by using a refractor velocity of 830m/s (velocity from velocity analysis) and a final datum of 67.75m. Figure 5.9 shows the result of the refraction static correction. It is clear that most of signal time shifted towards zero by about 5 to 10ms. This is reasonable, since the velocity in the near surface is very low, and the value changes dramatically between the weathering layer and the dry sandstone.

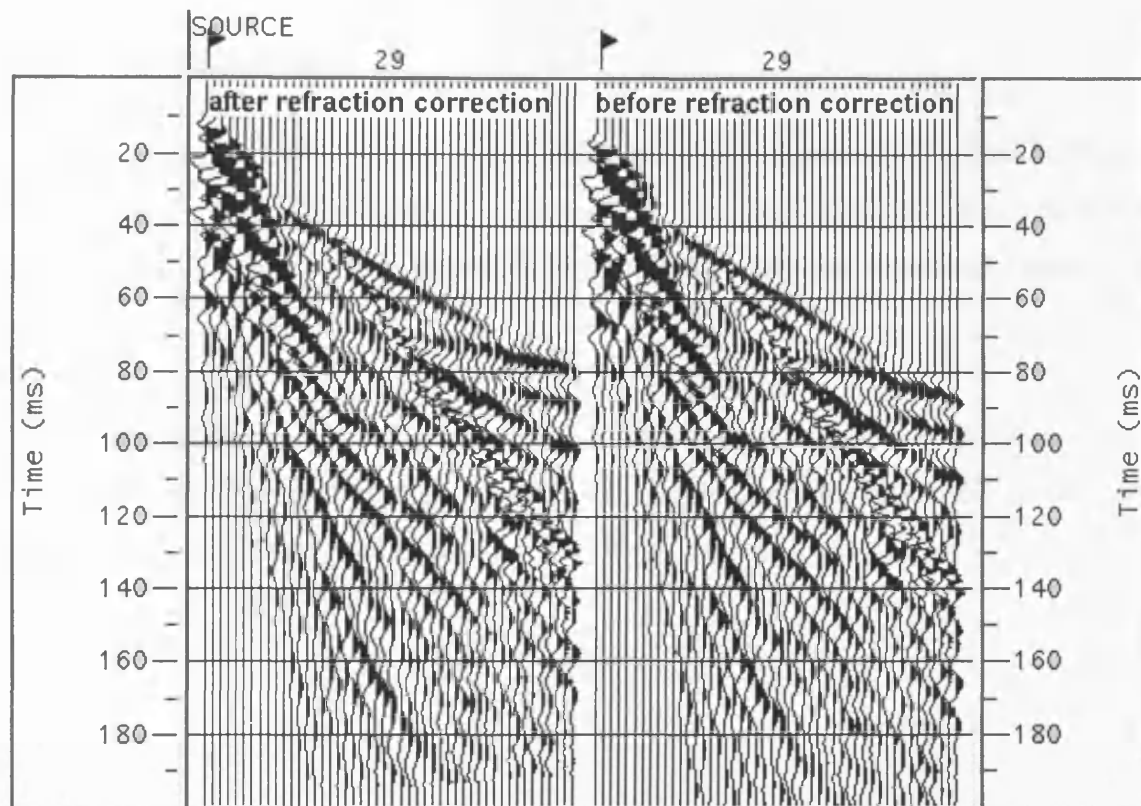


Figure 5.9 The result of the refraction static correction.

5.3.3. Filtering of Shot Records

5.3.3.1. Filtering in the Field

All data was effectively filtered twice in the field, firstly by the 100Hz geophone response and secondly by the 128Hz locut -1000Hz Hicut recording filters. These were chosen based on the field parameter test in order to get the highest dominant frequency in order to improve the vertical resolution. Although there is a general problem with high frequency attenuation at the near surface, the frequency content of the recorded data is high, up to 400Hz.(See figure 5.9).

5.3.3.2 Bandpass Filter

The window of this filter is chosen based on the test parameter illustrated in Figure 5.10, giving a range of frequency on each panel. The panel represents the shot record excluding traces 1-16 (due to very strong ground roll that obscures changes in reflections).

The optimum window is selected as 50,60,300,400. It only reduces the small part of ground roll. This is caused by effective filtering in the field. Figure 5.9 shows very little difference before and after filter.

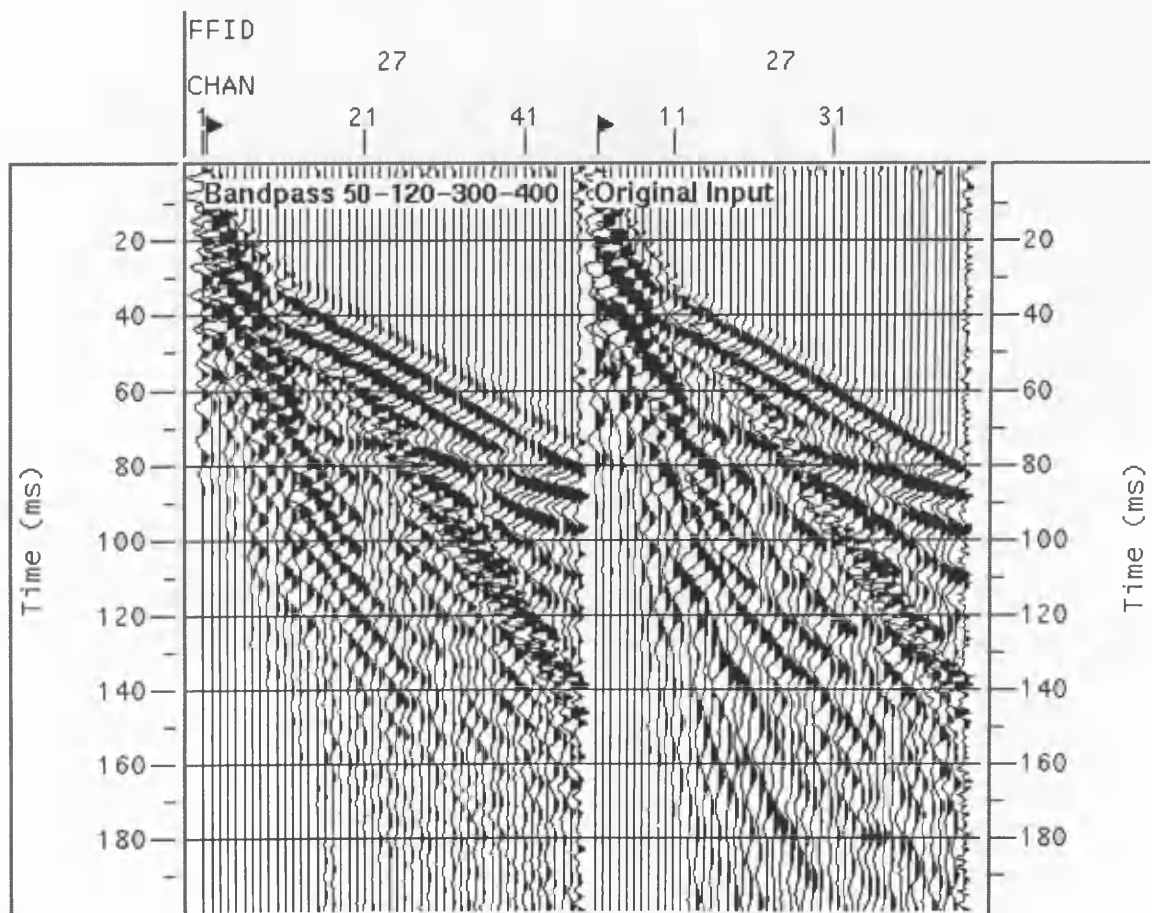


Figure 5.11 Band pass filter of 50,120,300,400 can reduce a little ground roll

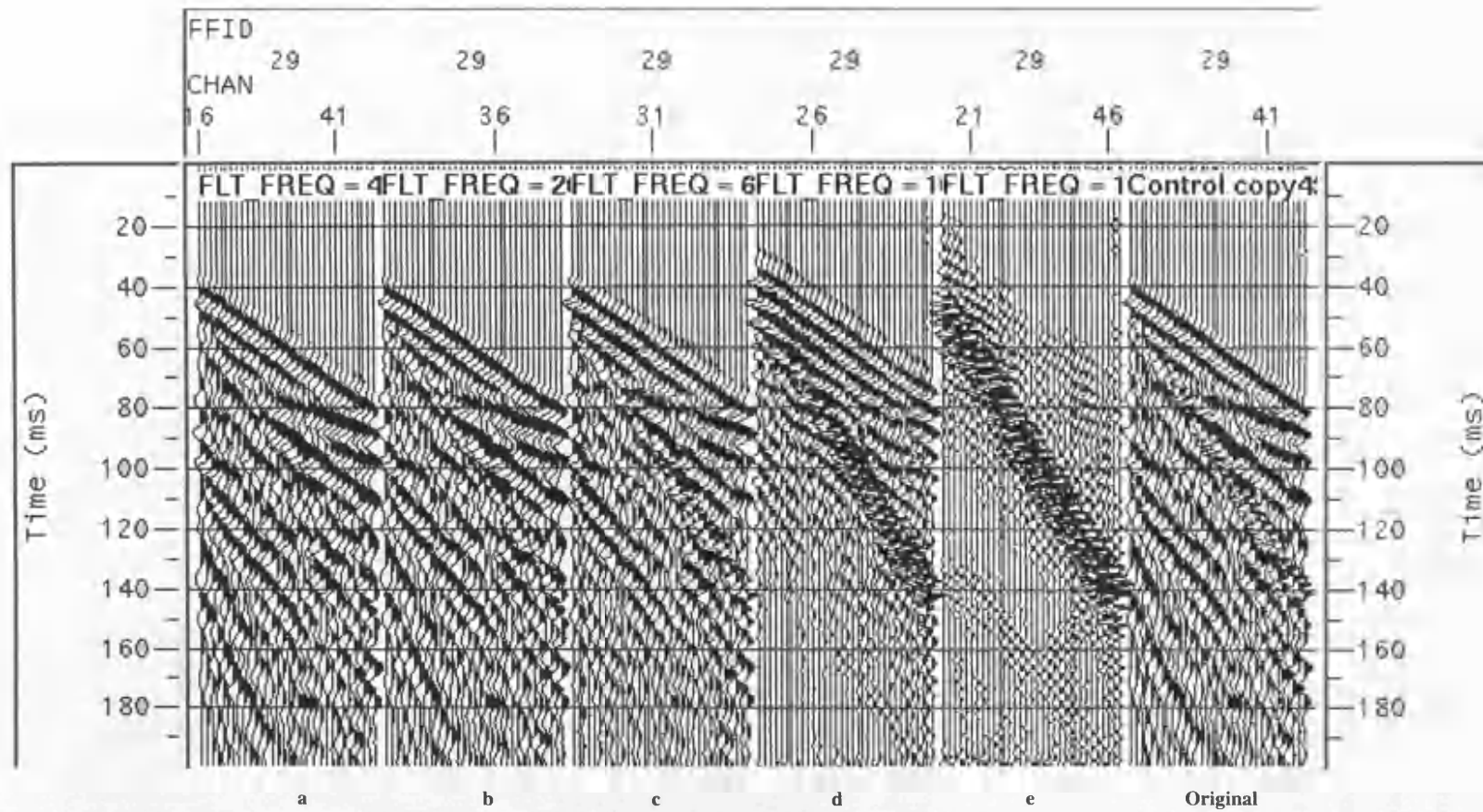


Figure 5.10, A range of Band pass filter window
(a.4-10-100-200, b.20-40-150-250, c.60-100-200-250, d.100-150-250-350, e.170-200-300-450).

5.3.3.3. F-k Analysis and Filter

Traces 1-16 have been excluded to reduce the accumulation of energy near the source. Figure 5.12 shows a f-k analysis screen where we believe that the certain place (red colour in wave-number panel) is reflection energy. Therefore, a f-k window of –850 to 700m/s and 60 to 300Hz has been chosen as the best range. It is a reasonable f-k window to remove the ground roll that has velocity < 700m/s and frequency < 60Hz. (measuring in the shot record). Minus number of –850 is relating to the negative direction of shooting, which we use the South.

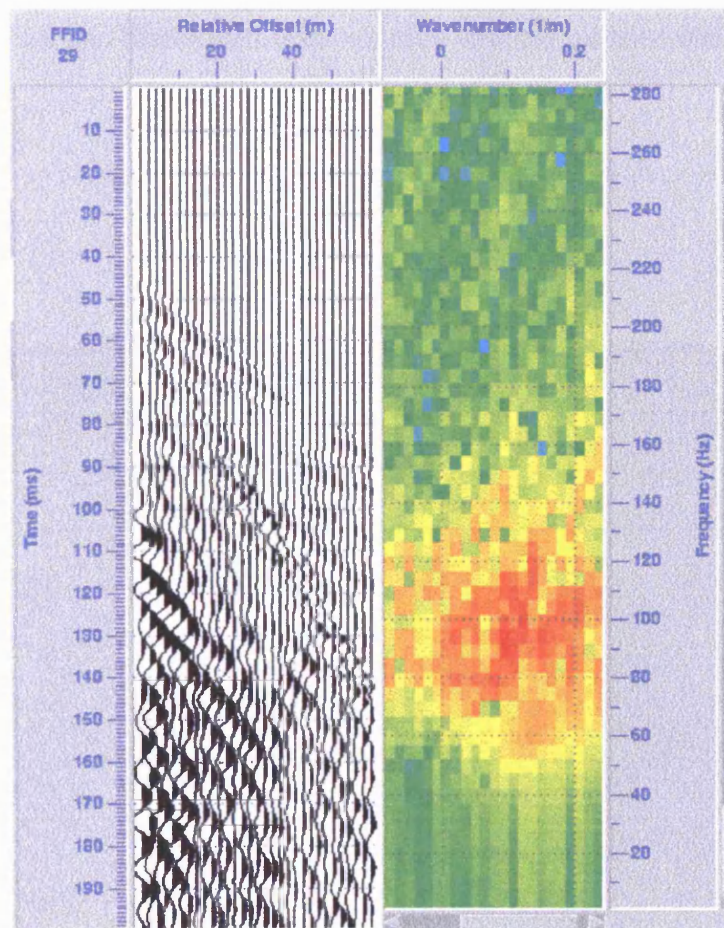


Figure 5.12 F-k analysis screen, the only certain place has the energy that is supposed as reflection energy.

Figure 5.12 clearly shows that reflection energy in wave-number panel is not symmetrical i.e. most of energies are in the positive numbers. This informs us that survey method is End-On and the shooting always goes ahead to the geophones. Therefore, the remaining energy in negative signs (the left side of zero in the wave-number panel, figure 5.12) is assumed to be noise. A f-k window of -850 is used to remove this noise.

Figure 5-13 shows the result of this filter. It is effective in removing ground roll. In the reject panel is the signal that has been removed, consisting of a lot of ground roll and some air wave energy. The accept panel shows much more clearly, a good indication of the water table reflection at 80ms in the middle and 95ms at the right. The remaining signals are multiples below the water table. We can see that the filter can also remove the air-wave and therefore it does not need to apply air blast attenuation.

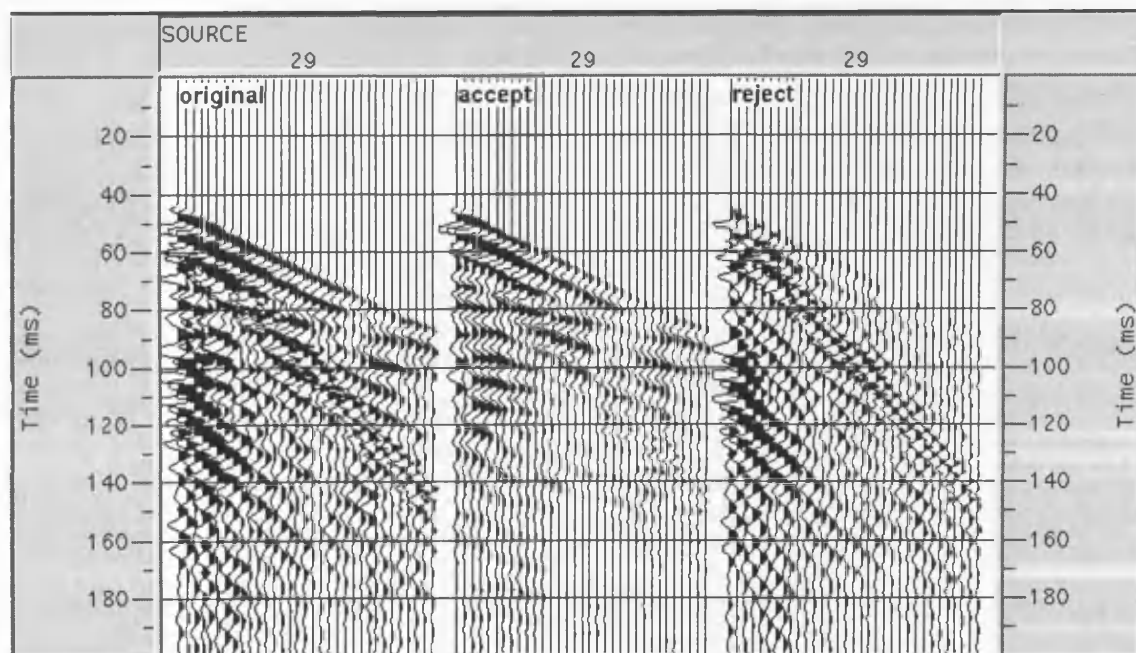


Figure 5.13 The result of f-k filter of -850 to 700 m/s and 60 to 300 Hz

5.3.4. Velocity Analysis and NMO Correction

A velocity function was calculated based on the interval velocity information from unreversed refraction & reflection data (shot record no. 29) and the boundary velocity as assumed by the water table from the part of CMP section. The rms velocity from the surface to a certain depth (200ms), is 1:0-830, 70-900, 150-1500, 200-1800. The important thing is to assume the first layer is dry sandstone with a velocity of 830m/s, and that the soil layer has been replaced using this sub-weathering velocity by the statics correction. The velocity structure beneath this layer is a constant gradient velocity to depth to control the NMO correction. This is the only velocity function that has been used without using an interactive velocity analysis from the screen. This is because the record is too shallow and the maximum offset is too short. Figure 5.14 shows the result of NMO correction.

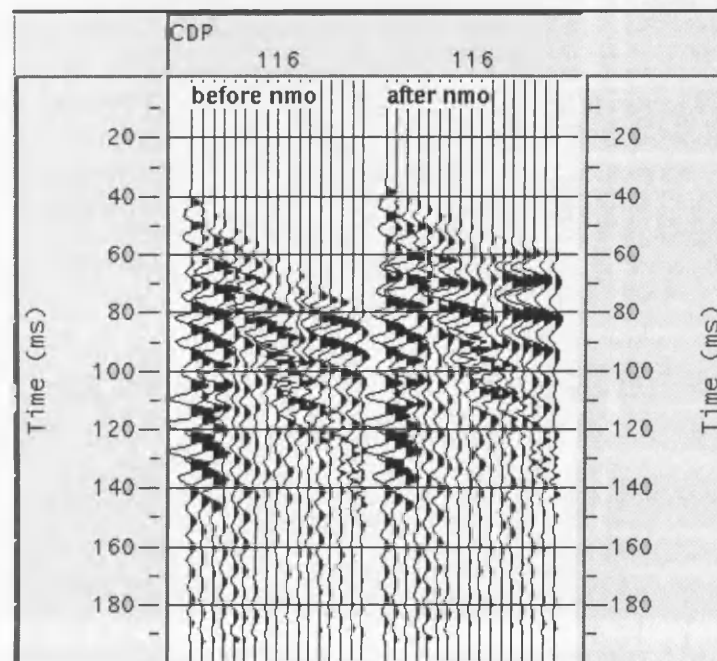


Figure 5.14 The result of NMO correction
(Velocity function of 1:0-830, 70-900, 150-1500,200-1800)

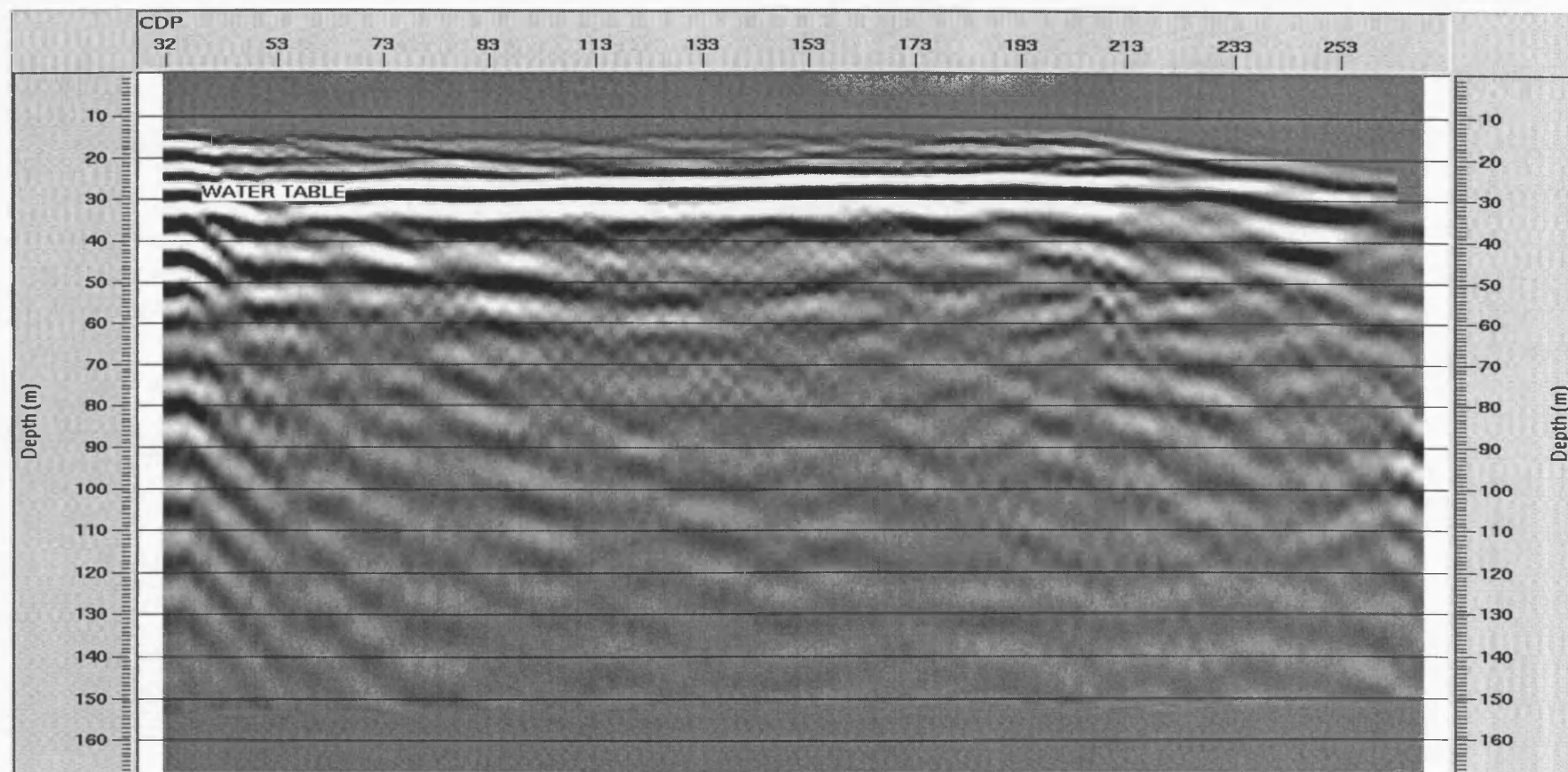


Figure 5.15a The result of stack using variable density plot with Time/depth conversion and CMP spacing of 0.25 m

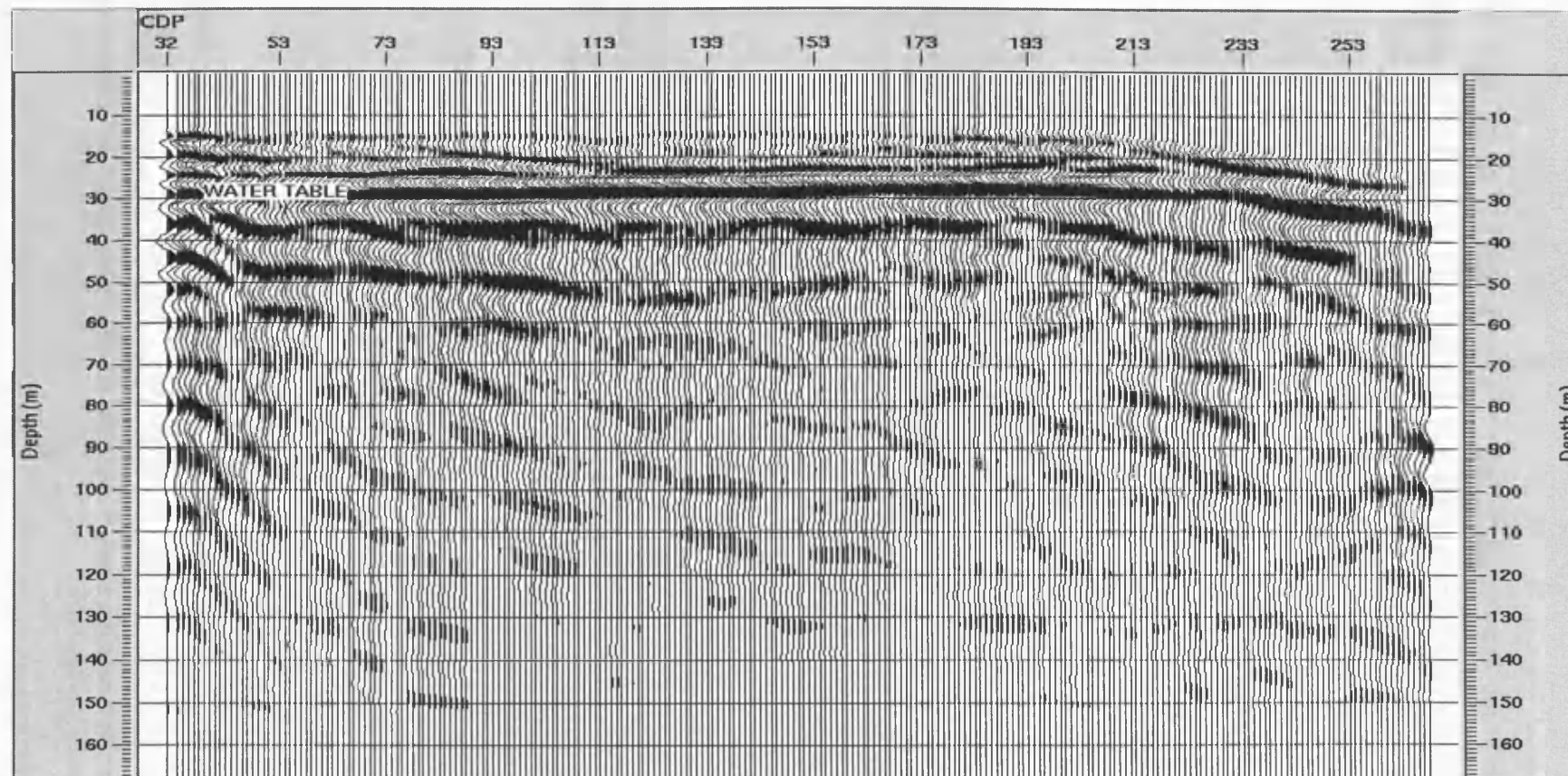


Figure 5.15b The result of stack using variable wiggle plot with Time/depth conversion and CMP spacing of 0.25 m

5.3.5. Stacking

The stack uses the mean or average value of each gather to replace a new CMP. Figure 5.15 shows the results of stacks with reference to CMP lines that indicate maximum of fold of cover in the middle. Table 5.4 summarises the list of processing for the Edwinstowe data set.

Table 5.4 The list of processing for Edwinstowe data set.

Editing :	
- Trace kill/reverse	: Trace kill
- Trace muting	: Top muting
Gain and Polarity	: - Automatic Gain Control at before and after CDP/Ensemble stack - True Amplitude Recovery at only before CDP/Ensemble stack
Bining	: Matching pattern number using first live channel and station.
Velocity Analysis	: Estimation from interval unreversed refraction data and reflector : velocity assumed as water table from the CMP section
Filter:	
-Band pass	: 50,60,300,400 Hz.
-F-k	: F-k analysis; -850, 700 m/s : 60, 300 Hz. (accept)
-NMO	: 0-830, 70-900, 150-1500, 200-1800
No. of CMP's	: 268
CMP Spacing	: 0.5 m
Nominal Fold	: 48
Stack	: Mean stack, all traces.
Output	: Variable density and wiggle plot

5.4. Interpretation

5.4.1. Geophysical Interpretation

According to the calculation of the reflection coefficient from the porosity (figure 2.5), the P-wave velocity of the dry sandstone is 830m/s (velocity analysis) which will give a velocity of 2040m/s for rock and CR of 0.515 for 33% porosity. This is very close to the value of the velocity structure before processing (table 5.2), giving 810m/s to 2025 and CR of 0.524 for 34%. It is also very similar to the result from refraction record of table 5.3 that give 790m/s to 2015 and CR 0.53 for 35%. Table 5.5

shows the result of the velocity structure by a different method. We believe that the best method is the velocity analysis, because it is done very accurately.

As we know the larger distance of offset, the greater refraction velocity we will get due to the greater depth of medium travelled. This is because the greater the travelling depth the greater the P-wave velocity. Therefore, this is a reasonable refraction result. The important interpretation here is that all porosities are in the range of estimation suggested in the introduction as 14-36%. Furthermore, the variation of porosity interpretation is less than 3%.

Table 5.5 the result of velocity structure

Method	Dry Sandstone (m/s)	Wet Sandstone (m/s)	Porosity (%)
Refraction records no 3 & 5	790	2015	35
Manual picking from 20 records	810	2025	34
Velocity analysis	830	2040	33

5.4.2. Geological Interpretation

It is interpreted that the depth from surface to the water table is 29m in the North and 31m in the South with a picking error at the final section of ± 0.5 m. This interpretation is comparable to the value given in table 5.2, that gives an average depth of $28.2\text{m} \pm 2.7\text{m}$, which is about 1.5m deeper than what we have calculated from refraction shot records no 3 & 5. Obviously, that calculation, using only two shot records, is inaccurate in determining depth. However, refraction calculation of this stage is only for making preliminary confirmations, especially for velocity analysis of velocity structure. In the next section, a complete refraction survey will be reported.

The interpretation of water table depth is closer to the borehole information of 31m than previous attempts (table 5.2 and 5.3). The possible reason for a ± 0.5 errors from 31m, is due to a borehole position some 700m NNE off-line. Therefore, an offline borehole position alongside seasonal fluctuations may alter the position of water table.

Results indicate that the water table has a slope that follows the contour of the ground surface, 1.4° down from North to South. In addition, we cannot see any reflector at 120m in the impermeable Upper Permian, suggestive of a Sandstone boundary. One possible reason is that the source does not have sufficient energy to reach such depths. The energy below the reflector does not have any relationship to the geological information, but it is only remaining energy after the filtering processes.

5.5. P-wave Refraction, VES (Vertical Electrical sounding), and S-wave Test Site

5.5.1. General Consideration

It is interpreted using seismic reflection that there is a very strong, clear boundary layer at a North end of line (31m) which agrees with the borehole information as being the water table. But how do we know that the detected reflector is a hydrological boundary (water table) and a not lithological boundary as normally detected by seismic reflection? Another type of geophysical survey was used to confirm this reflection result. Several geophysical surveys methods were performed and will be subsequently described.

5.5.2. P-wave Refraction Survey

5.5.2.1. Introduction

The survey objective is to locate the water table as a water saturated sandstone refractor that was estimated to be at around 30m depth (from the reflection survey). Seismic reflection methods never detect the velocity underneath the reflection boundary, because reflected energy will be reflected back up by the top of the boundary. On the other hand, the head wave in the refractor for refracted energy will include the underneath boundary. However, the larger the offset, the more depth of detection and also the larger velocity. In this case, the refraction method has a good chance of detecting velocities with a large offset. The complete survey; forward and reverse survey, will give the depth interpretation model.

5.5.2.2. Field Data

44 shot records were recorded along the seismic line (which is the same as the line of seismic reflection). Figure 5.1 shows the location of fieldwork, just extended from 90m to 230m.

The shot records were recorded with a fixed geometry as shown in table 5.6 and figure 5.16 as the planned model. The shot point moved from 0 m to 230 m with 46 m spacing for first Geophone at 0,46,92,138 and 184 m along the line in a NNE direction. Ideally total records should consist or at least 30 shots.

The collected data was of good quality, and all data was sufficient for data processing except three shot records that were very poor due to very large offsets where the signals were particularly weak and noisy. There are 14 extra shot records, which after sorting become 30 shot records for the complete configuration. Figure 5.17 shows the T-X curves of the real data. It is picked manually from the screen from an original record with a frequency filter of 40,60,300,400 Hz.

Table 5.6 Technical description of the seismic data location, Equipment, and acquisition parameters.

SITE	
Site Name	: Edwinstowe
Grid Ref.	: SK 630 675
Geology	: Sherwood Sandstone
Depth Range	: 0-60 m
Date of Survey	: 01/11/1998
EQUIPMENT	
Seismograph	: Bison 9048
Geophone type	: Mark Products 8 Hz, Vertical 125 mm spikes.
Source Type	: Hammer with plastic plate some shots with buffalo gun
Record length	: 500 ms.
Sample interval	: 0.5 ms.
Record Filters	: Locut: 4 Hz, Hicut: 500Hz.
Gain	: Medium
GEOMETRY	
Spread type	: On-end spread
No. of channels	: 24
Length of line	: 230 m
Geophone spacing	: 2 m
Shot spacing	: 46 m
No. of shots	: 44
Shot depth	: 0 and 1 m
Shot offset (X)	: 0
Shot offset (Y)	: - 2m to -0.5 m

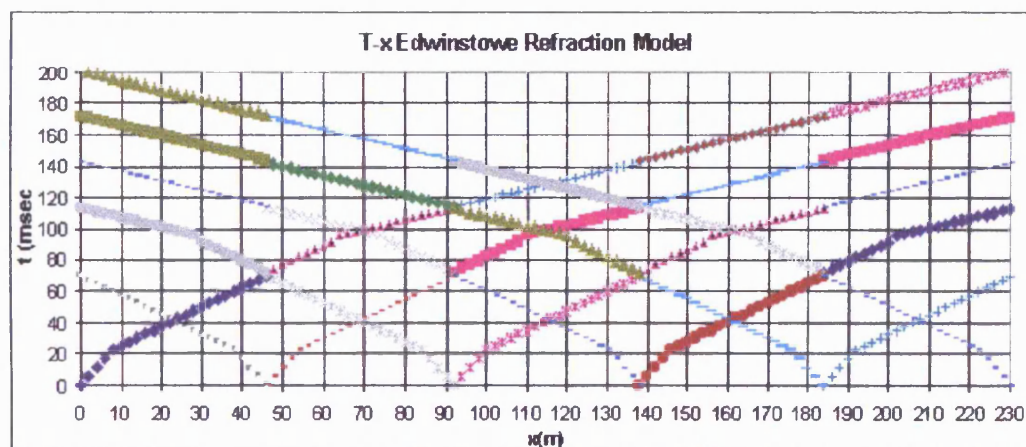


Figure 5.16 T-x curves for three layers as the planned model

Surveying data was taken by levelling using a Zeiss autolevel for linking all station points with 2-meter spacings. The south end of seismic line, $x=0.00\text{m}$, starting from Southern boundary of field (plan barrier) is 63.29 high while North end as $x=230\text{m}$ is at 70.93m. Figure 5.18 shows the elevation profile of the seismic line. It is matched at 68m for $x=0.00\text{m}$ by the previous data (reflection survey) from the South end. This elevation was used to construct the ground model of site on figure 5.2.

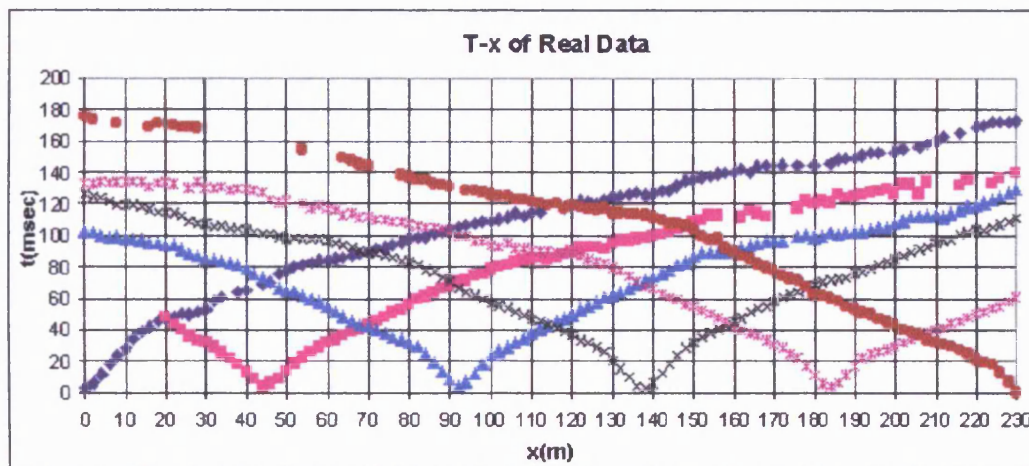


Figure 5.17 T-X curves of the real data

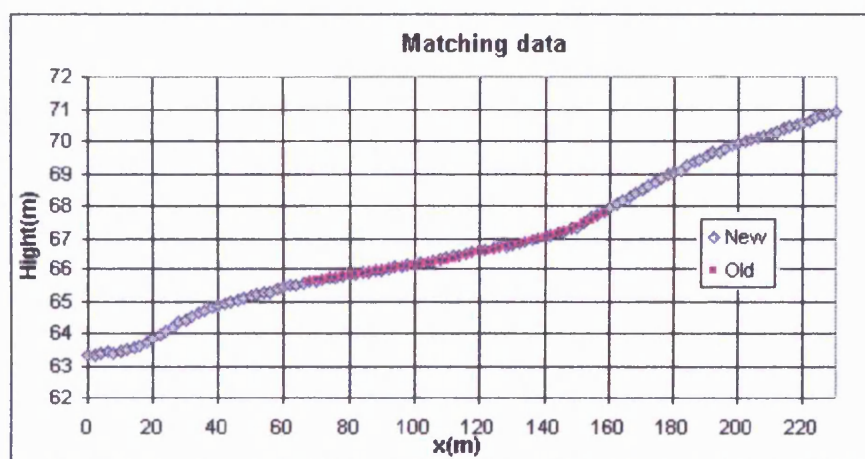


Figure 5.18 the elevation profile of the seismic line.
(Old = seismic reflection line; New = seismic refraction line)

5.5.2.3. Calculation of Velocities and Depths

Before calculation of velocities and depths, real data was modified. Several points have not been recorded due to poor quality, and these were filled using the principal interpolation. This will be useful in deciding the first point of the second boundary layer, which we will calculate. The next modification is to find the point at $x=230\text{m}$ that we have no data for due to overlap record at every station no.24 by $x=0\text{m}$ of following station. This has been done using the extrapolation principal. Furthermore, the data has been simplified to only 6 long shot records (116 stations for each of A, B, C, D, E, and F), for the source position of 0, 46, 92, 138, 184, and 230m. Figure 5.19 shows the modification of real data.

This produced very good t - x curves. The end time of each pair of records is nearly ± 2 ms. The curves indicate that the top boundary of the second layer is nearly horizontal and parallel to the surface elevation. This is indicated by the symmetrical curve of each pair of forward and reverse records. For example, the record of source position 92m and 138m (C and D) has a cross-point position at $x=116\text{m}$ and is symmetrical for the travel time below 80ms.

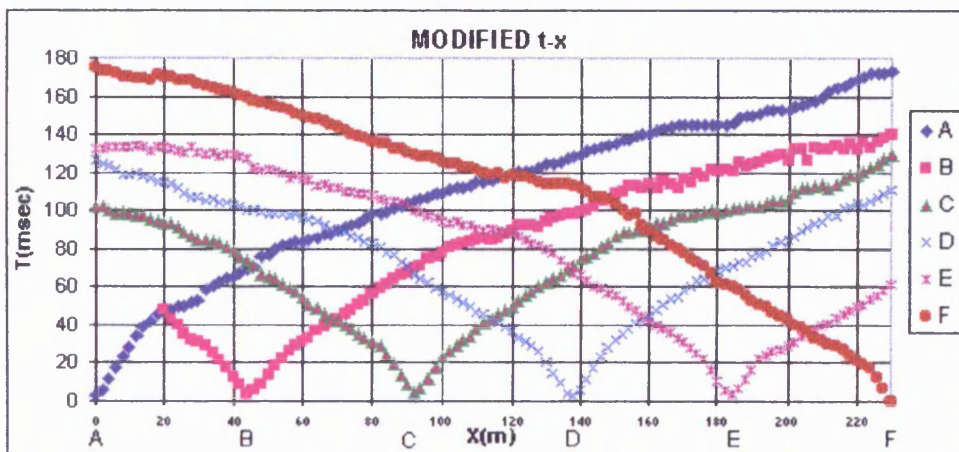


Figure 5.19 the T-x modification curves of real data

The bottom boundary of the second layer can be predicted as having a dip, and is not a horizontal boundary, as indicated by the unsymmetrical curve of source positions 0 and 230m, especially for the third layer at a range between (x, t) 50,80 and 144,110. However, the dip is estimated to be below 10° , as indicated by the T-minus curve that is normally linear and T-plus curve that is normally linear-horizontal (figure 5.20). Therefore, for this reason, the calculation method of plus-minus has been applied (Overmeeren, 1997).

Regarding velocity and depth calculation we will calculate three velocities (V_1 , V_2 , and V_3 as velocity of weathering layer, dry-sandstone, and wet-sandstone respectively) and two depths (h_1 and h_2 as depth of dry-sandstone top and bottom boundaries respectively). The velocity of V_1 and V_2 are calculated from travel-times of different shots on the line, which are averaged to simplify the correction of the arrival time (Musgrave, 1967). The average velocities obtained are 400 and 870m/s for V_1 and V_2 respectively. The velocity of wet-sandstone (V_3) is determined from the calculation of minus-time. Figure 5.20 shows the minus-time curve for the third layer from shot A-F, from which V_3 is obtained as 1990m/sec. This is smaller than expected at around 2060m/s (see appendix A) because the range of minus-time is too long. In this case we used 50m to 150m (see figure 5.19) for shot A – F, (the correct range is 56 to 148m). Therefore, we will obtain a smaller slope in the minus-time curve, and a bigger velocity value.

The depth of the dry-sandstone top layer (h_1) is derived from the calculation using the travel time of the refraction wave for the single refractor. The average of h_1 is obtained as 3.6 ± 0.5 m. The depth of the wet-sandstone top boundary (h_2) is calculated from the plus-time of the third layer, figure 5.20b. The average of the depth h_2 is 26.6 ± 1 m.

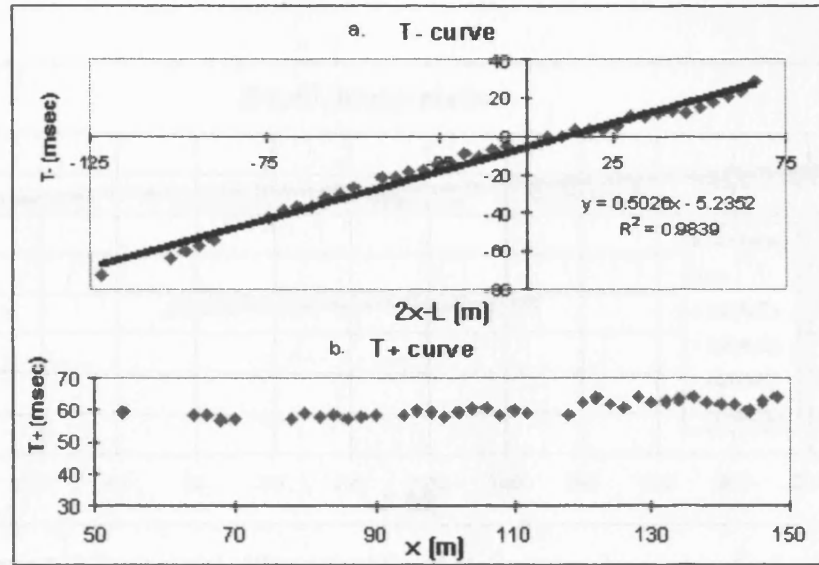


Figure 5.20 T-minus and T-plus from the calculation of the third layer from shot A-F

5.5.2.4. Interpretation

Figure 5.21 shows the result of the calculation using plus-minus method. The physical parameters of the calculation are very similar to the previous reflection survey. Plus-minus method produces an average velocity for three layers of V1, V2, and V3 of 400, 870, and 2090m/s respectively for the average Z1 (thickness) and Z2 of 3.6 ± 0.5 m and 26.6 ± 1 m. Meanwhile, the velocity structure from the reflection processing of 400, 830, and 2030m/s respectively are for V1, V2, and V3 and for Z1 and Z2 are 3.5 ± 0.5 m and 26.5 ± 1 m.

According to the Velocity-Porosity curve for dry-to-wet sandstone (see appendix A), the subsurface profile gives the porosities of 30-33%, figure 5.22. The 90m length of line from 58-148m is position-matched with the previous reflection line (figure 5.21). We can confidently say that both the refraction calculation and reflection process are satisfactory. Therefore, this refraction survey is strong evidence that the refractor is the water table.

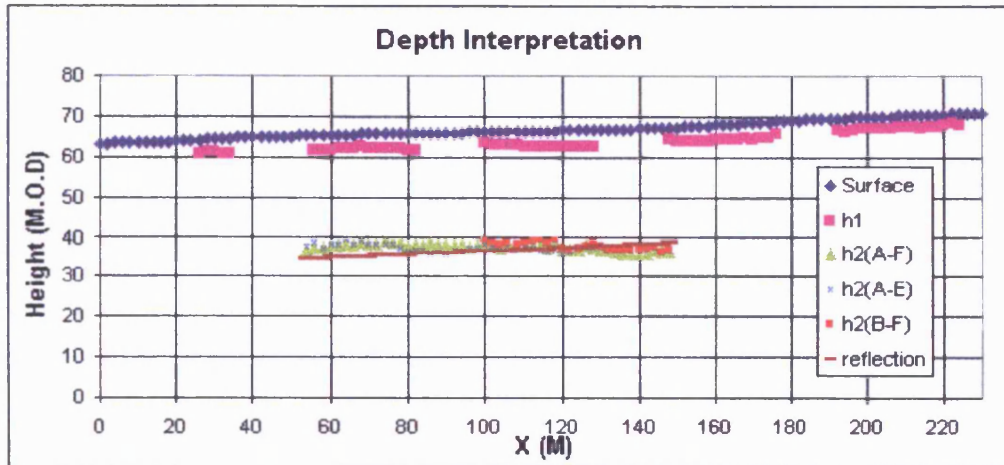


Figure 5.21 Depth model of the result of the calculation using plus-minus method
(h2(A-F): the result of A-F line, h2(A-E): the result of A-E line, h2(B-F): the result of B-F line)

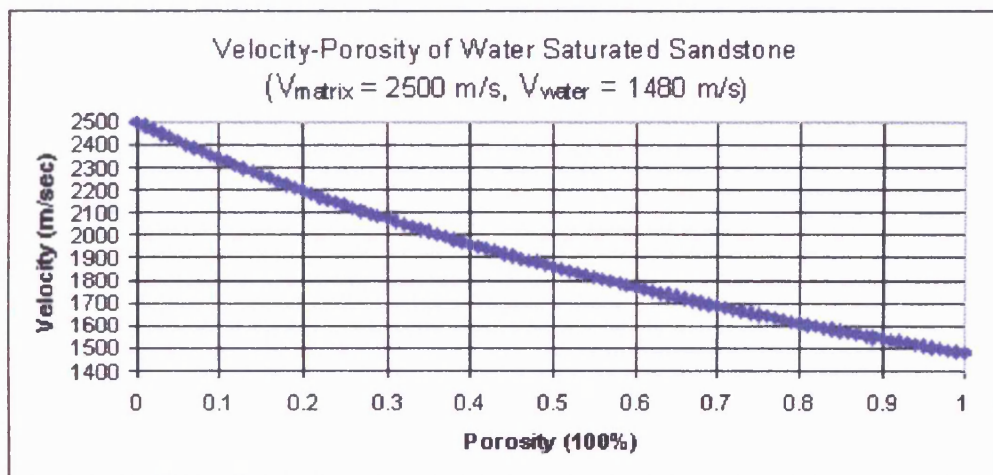


Figure 5.22 the curve of Velocity – Porosity of water saturated sandstone
(the calculation is in see appendix A)

5.5.2.5. Conclusion

As mentioned in the introduction, the objective is to locate the water table as a water saturated sandstone refractor was estimated to be around 30 m depth (from reflection survey). It is a satisfactory result that the calculation produced similar parameters to the reflection survey. The water table is at depth of $29-31 \pm 1\text{m}$ as the second boundary layer with sandstone porosities of 30-32%.

The evidence that the second boundary layer is the water table is the velocity structure that occurs as the Velocity Vs Porosity curve. Also, borehole information shows that the geological background of site location is near homogenous. Additionally, it shows the horizontal nature of the weathered sand layer and Sherwood Sandstone with a total thickness of about 120 m. This is underlain by impermeable Upper Permian (Mercian Mudstone). This is sufficient proof that the second boundary is a water table (having no geological structure at that depth). Additionally, further corroborating evidence for this, can be found in the resistivities survey

5.5.3. VES (Vertical Electrical Sounding)

The resistivity survey is performed using the Schlumberger configuration. The constant potential electrode is in the middle and the current electrode is a changing parameter. For the technical field, we used the middle point at 115m, which is in the same position as the refracted survey lay-out of 230m length. Using a potential electrode distance of 4 m, we recorded 13 measurements. Eleven of these being the current electrode distance, at 20m increments (starting at 20m and ending at 220m). The remaining 2 measurements consist of the first and last current electrode distance, of 10m and 230m respectively. Regarding the instrumentation, we used a Terrameter 300 to record the measurement, which reads 16 times then averages in every record.

Figure 5.23 shows the result of the survey and was interpreted using VES for Window Version 1.20. The interpreted curve is very close to the observed points with an error (standard deviation) of about 13%. This is shown in figure 5.23 that the interpretative curve (solid line) is very close to the data. Using 30 iterations the VES software produces the final numbers as interpretation including 2 layers; $h_1=2.9\text{m}$ and $h_2=26.6\text{m}$, and 3 resistivity values; $R_1=270\ \Omega\text{m}$, $R_2=324\ \Omega\text{m}$ and $R_3=75\ \Omega\text{m}$.

It is believed that this interpretation is reasonable and confirmed by the specific resistivity of saturated sandstone with porosity range of 30-35 % that has a resistivity range of 75-100 Ωm , figure 5.24. This figure is an implementation of Archie formula (Reynolds, 1997) for Mesozoic Sediments as confirmed by the Triassic Sherwood Sandstone.

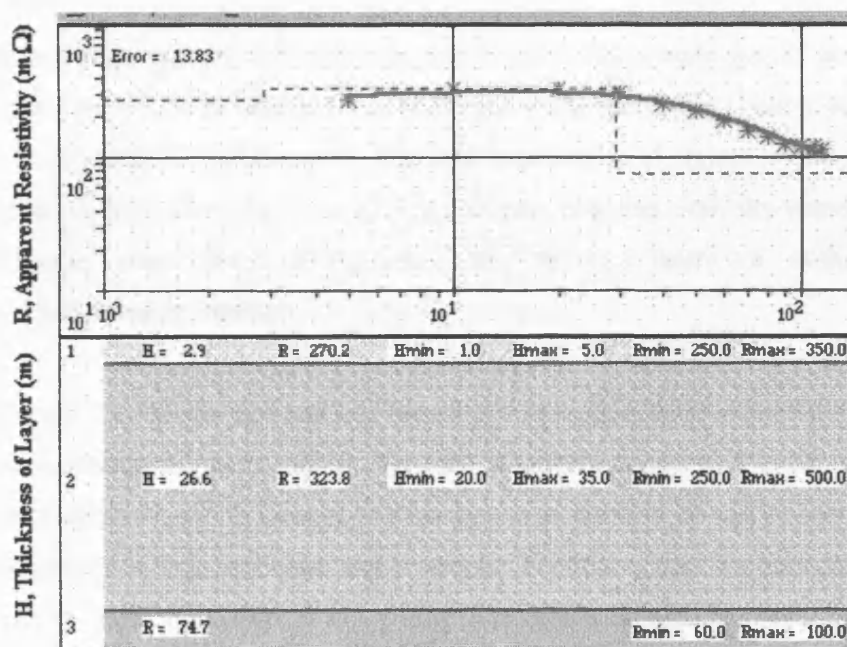


Figure 5.23 the result of VES survey and interpretation.
 (* is the measured points, solid line is interpretation curve,
 and dashed line is the interval apparent resistivity as the estimation value)

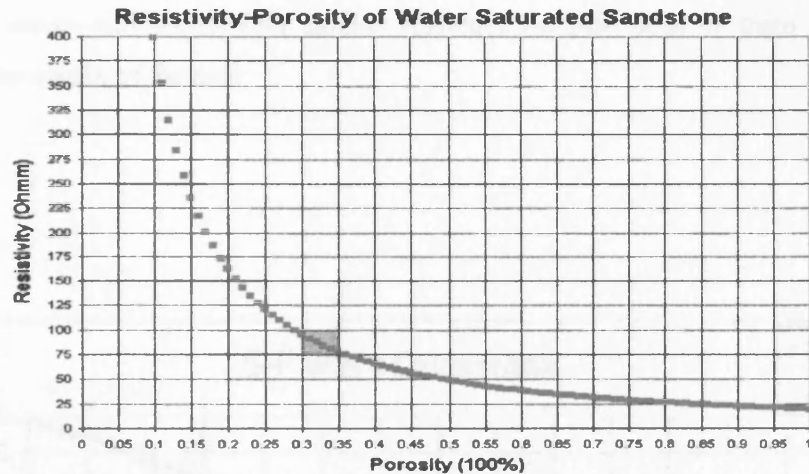


Figure 5.24 implementation of Archie formula (Reynold, 1997) for Triassic Sandstone

5.5.4. S-wave Refraction Survey

S waves cannot travel through a liquid medium. The simple way to do this is to make a one line test of S- and P-wave recording in the same line (Hasbrouck, 1991), which is to compare S- and P-waves. The ideal expectation of certain offsets is that the P-wave records have three layers as different slopes including; soil, dry sandstone, and wet sandstone, while, the S wave record only has two layers i.e. excluding wet sandstone due to liquid medium.

Figure 5.25 shows that test site record proves exactly those expected conditions. The only weakness of this result is the ratio S-wave velocity to P-wave velocity is around 80% while Sheriff & Geldart 1983 suggest that the ratio should be less than 70% when the ground is homogeneous and isotropic. Therefore, here we interpret that the subsurface or earth condition is not purely homogeneous and isotropic. A possible reason is that this is caused by unsaturated sandstone that is too dry. Both P and S waves travel through the matrix and pore space of the rock. When the pores are not very dry then P wave velocity will be bigger than S wave velocity. This is caused by S-waves that cannot travel through the liquid. On the other hand, when the pore is very dry then

P and S waves will travel with similar velocities because both of them only travel through the matrix of the rock.

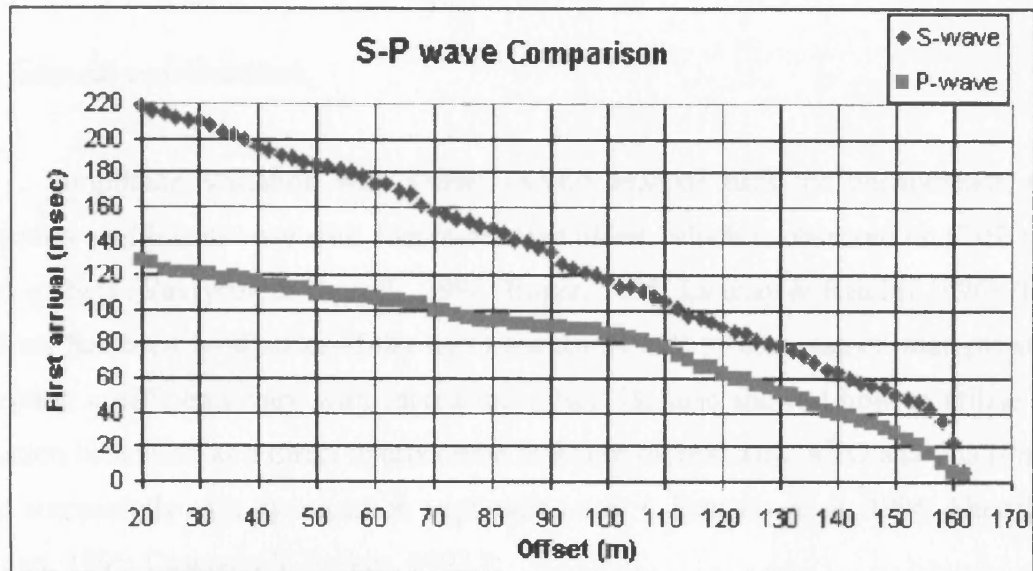


Figure 5.25 Test Site S- and P-wave refraction record (Picked raw data are in appendix E)

Chapter 6

AMPLITUDE VARIATION WITH OFFSET**6.1. General consideration**

Amplitude Variation with Offset (AVO) analysis uses the phenomenon that reflection coefficients vary with source-receiver offset, which is observed on CMP pre-stack gathers (Vavrycuk & Psencik, 1998; Ruger, 1998; Lindsay & Ratcliff, 1996). This analysis has been used successfully by Ostrander (1984) to demonstrate that gas sand reflection coefficients vary with increasing offset. He also showed how to utilise the variation behaviour as a direct hydrocarbon indicator on real data. AVO analysis is now used successfully as a hydrocarbon exploration tool (Santoso et al. 1996; Sheriff & Geldart, 1995; Castagna & Backus, 1993).

Castagna & Backus (1993) described AVO, as “a *seismic lithology*” tool, which provides an improved model of the reflection seismogram. The model allows a better estimation of both normal incidence reflection coefficients and background velocity. These properties might be directly related to lithology and fluid content. In the following five years he established the framework for AVO gradient and intercept interpretation. AVO interpretation has been facilitated by crossplotting the AVO (A) as coefficient reflection at normal incidence and gradient (B) as slope of offset dependency as pointed out by Castagna et al. 1998. In order to identify the existence of fluid in a reservoir, brine-saturated sandstone and shale that follow a well-defined, characteristic “background trend” in the A-B plane has been used. A classification of gas (fluid) saturated sandstone based on location in this plane has also been made.

AVO analysis has also been used to identify the reservoir fluid, such as gas, water and oil by plotting the value of the P-wave velocity against Poisson’s ratio (Santoso et al. 1996). Skidmore et al. (1997) concluded that AVO analysis helps seismic imaging in deepwater environments.

The author here presents an opportunity to use AVO analysis for identification of water saturated sandstone as a means of identifying the water table in shallow seismic exploration. The water table should theoretically produce a clear AVO response, which is very different from a lithological boundary. It is hoped to show that we can use this phenomenon to identify the water table. Firstly, data will be collected over the water-table and lithological reflection. Secondly, we will prove the amplitude errors introduced by instrumentation and processing. Finally, we will show the observed AVO anomalies, which cannot be due to data errors, and are compatible with the AVO model.

6.2. Theoretical Background

6.2.1. Reflection Coefficient

Referring to the simple form of Zoeppritz's equation (2.53) for normal incidence (assuming up to 15°), we need to expand further the general case where the angle of incidence exceeds 15° . Consequently, the equations for the coefficient of reflection and transmission (as a solution of the wave equations) become more complicated and include the term θ (the angle of incidence). Tooley et al (1965) shows the variation of amplitude with angle of incidence for several sets of parameters.

Figure 6.1. shows the P-wave reflection coefficient for various P-wave velocity ratios when, $\rho_2/\rho_1 = 1.0$, and $\sigma_1 = \sigma_2 = 0.25$. The critical angle varies as the variation of P-wave velocity ratio (α_2/α_1), and gives this figure its complex appearance. When there is no impedance contrast or the velocity ratio is unity, then the reflected energy is zero (no curve for this case). The reflected energy increases both as the ratio becomes larger and smaller than 1. The two peaks for $\alpha_2/\alpha_1 > 1$ occur at the critical angle for P- and S-waves, respectively (Sheriff and Geldart, 1995). In the special situation where one medium is a fluid and the other a solid, large amounts of S energy are generated in the solid medium at large angles of incidence by P-wave incident from either medium

(Tooley et al, 1965).

Aki and Richard (1980) derived the solutions to the equations for the reflected and transmitted P-wave, that is frequently used to find the amplitude variation with offset (AVO).

$$R_p \approx \frac{1}{2} \left[1 - 4 \left(\frac{\beta^2}{\alpha^2} \sin^2 \theta \right) \frac{\Delta \rho}{\rho} + \frac{1}{2} \sec^2 \theta \left(\frac{\Delta \alpha}{\alpha} \right) - 4 \left(\frac{\beta^2}{\alpha^2} \right) \sin^2 \theta \left(\frac{\Delta \beta}{\beta} \right) \right] \quad 6.1$$

$$T_p \approx 1 - \frac{1}{2} \left(\frac{\Delta \rho}{\rho} \right) + \left(\frac{1}{2} \sec^2 \theta - 1 \right) \left(\frac{\Delta \alpha}{\alpha} \right) \quad 6.2$$

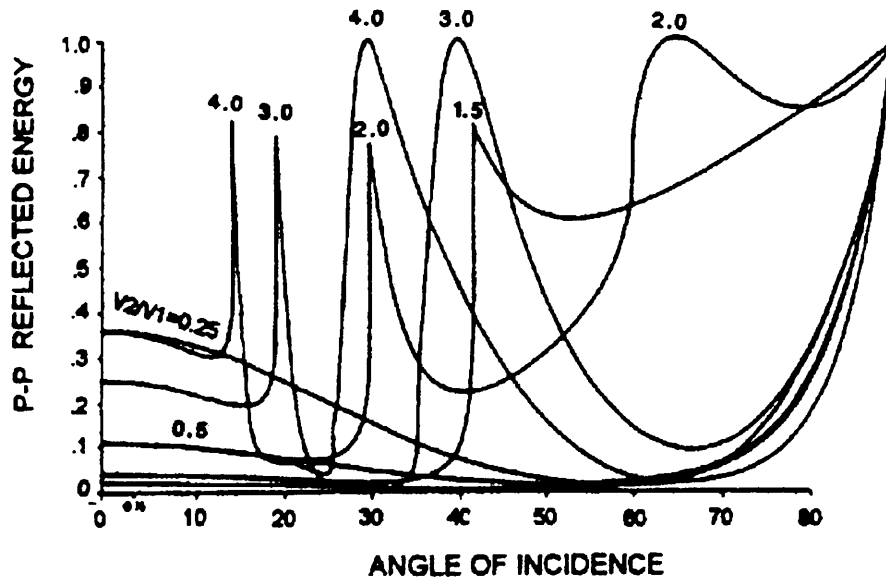


Figure 6.1 The effect on the reflected compression energy of varying the compression velocity ratio (V_2/V_1) (source Sheriff 1995)

Shuey (1985) made a simplification of these equations by changing β and $\Delta\beta$

with σ and $\Delta\sigma$ with;

$$\frac{\Delta\beta}{\beta} = \frac{\Delta\alpha}{\alpha} + 0.5\Delta\sigma\left(\frac{1}{1-\sigma} - \frac{2}{1-2\sigma}\right)$$

$$\beta^2 = \alpha^2 \frac{1-2\sigma}{2(1-\sigma)}$$

$$\Delta\alpha = \alpha_2 - \alpha_1 \text{ and } \alpha = (\alpha_2 + \alpha_1)/2$$

$$\Delta\beta = \beta_2 - \beta_1 \text{ and } \beta = (\beta_2 + \beta_1)/2$$

$$\Delta\rho = \rho_2 - \rho_1 \text{ and } \rho = (\rho_2 + \rho_1)/2$$

$$\Delta\sigma = \sigma_2 - \sigma_1 \text{ and } \sigma = (\sigma_2 + \sigma_1)/2$$

$$\theta = (\theta_2 + \theta_1)/2 \text{ with } \frac{\sin\theta_1}{\alpha_1} = \frac{\sin\theta_2}{\alpha_2}$$

then with further modification derived by him the relation:

$$R_p \approx R_0(1 + P \sin^2 \theta + Q(\tan^2 - \sin^2 \theta)) \quad 6.3$$

where ;

$$R_0 = \frac{1}{2} \left(\frac{\Delta\alpha}{\alpha} + \frac{\Delta\rho}{\rho} \right)$$

$$P = \left[Q - \frac{2(1+Q)(1-2\sigma)}{1-\sigma} \right] + \frac{\Delta\sigma}{R_0(1-\sigma)^2}$$

$$Q = \frac{\Delta\alpha/\alpha}{\frac{\Delta\alpha}{\alpha} + \frac{\Delta\rho}{\rho}} = \frac{1}{1 + \frac{\Delta\rho/\rho}{\Delta\alpha/\alpha}}$$

R_0 is the reflection coefficient for normal incident

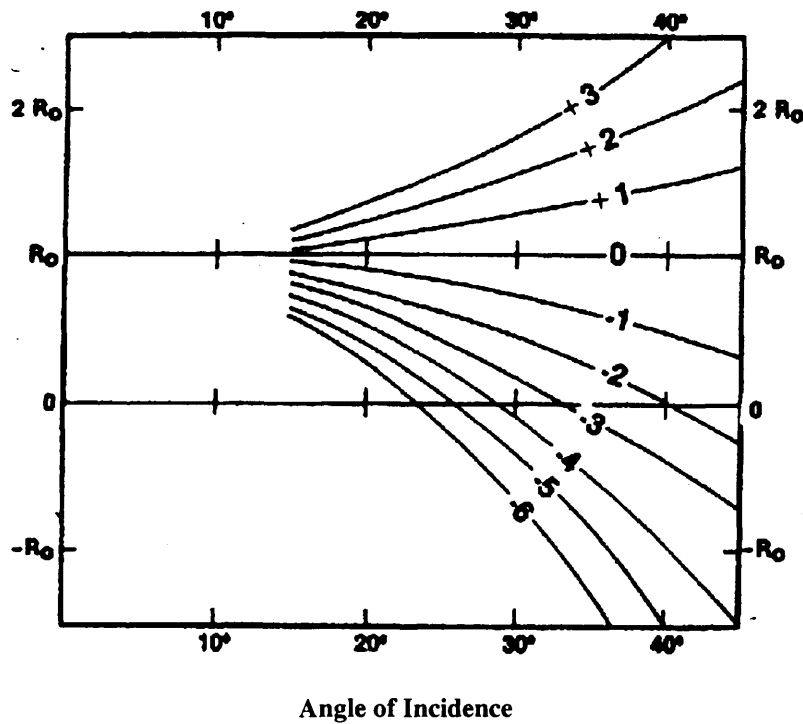


Figure 6.2 Parabolic approximation to $R(\theta)$ for different values of the dimensionless parameter P (source Shuey, 1985) with R_0 as Coefficient Reflection at normal incident

The simplification uses an assumption that Poisson's ratio is the elastic property most directly related to the angular dependence of reflection coefficient (Shuey, 1985). He also made a further modification to separate out the factor R_0 as the amplitude at normal incidence. It is easy to see that R_0 is an appropriate reference for $\theta \approx 0$. For intermediate angles ($0 < \theta < 30$ degree), the reflection amplitude is connected to the parameter P which is the sum of the two terms. The real component of that parameter is in the ratio $\Delta\sigma/R_0$. Figure 6.2 shows the variation of R_c that varies with Poisson's ratio (σ). It has specification of $P = -1$ for $\Delta\sigma = 0$, $P > -1$ for $\Delta\sigma > 0$, and then $P < -1$ for $\Delta\sigma < 0$. This is an approximation for the intermediate angle.

Lately, Hiltermann (unpublished and private communication reported by Sheriff, 1995) rewrote eq. 2.29 in the form:

$$R_p \approx R_0 \left[1 - 4 \left(\frac{\beta^2}{\alpha^2} \right) \sin^2 \theta \right] + \frac{\Delta \sigma}{(1 - \sigma)^2} \sin^2 \theta + R_0 \left(\frac{\Delta \alpha}{2\alpha} \right) \left[\tan^2 \theta - 4 \left(\frac{\beta^2}{\alpha^2} \right) \sin^2 \theta \right]$$

and he made a further approximation for intermediate angles that neglects the third term (dominated by velocity dependence). For a half space medium this is given by (Al-Ghamdi et al, 1998):

$$R_p \approx R_0 \cos^2 \theta + 2.25 \Delta \rho \sin^2 \theta \quad 6.4$$

Ostrander (1984) applied these results to practical cases, to study the variation of a P-wave reflection coefficient with angle of incidence. The reflection coefficient becomes more negative with increasing incident angle (figure 6.3). He concluded that for practical reflection cases there are three possible results:

1. For small changes in Poisson's ratio (figure 6.3a), the amplitude decreases with increasing incident angle, regardless of the polarity of the reflection coefficient.
2. The amplitude increases with incident angle for:
 - A positive reflection coefficient and an increase in Poisson's ratio (which is considered to be true for a gas/water contact or the base of a gas sand embedded in shale.
 - Or, a negative reflection coefficient and decrease in Poisson's ratio (which is considered to be true for the top of a gas sand embedded in shale)
3. The amplitude decreases with incident angle at first and then the waveform reverses polarity and the amplitude increases with opposite polarity, for either:
 - A positive reflection coefficient and a decrease in Poisson's ratio.
 - Or, a negative reflection coefficient and increase in Poisson's ratio.

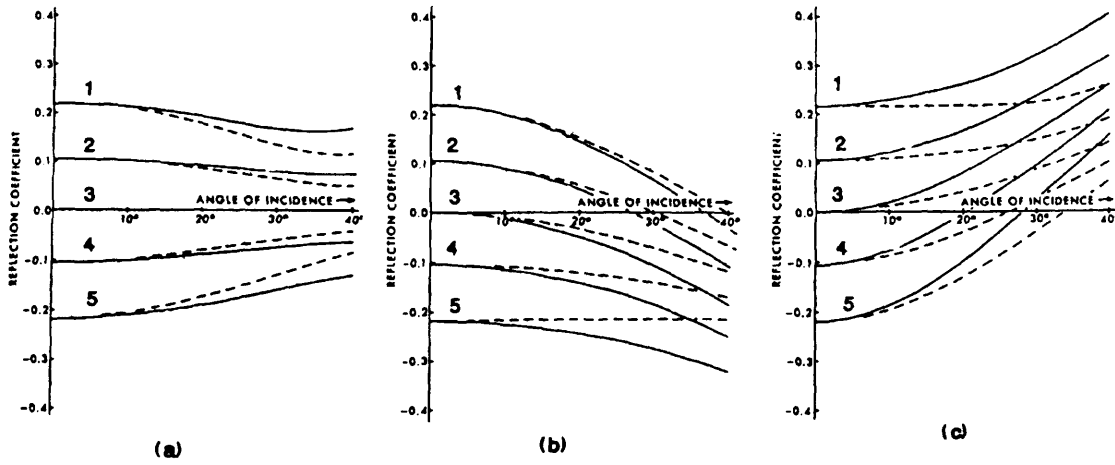


Figure 6.3 Variation of a P-wave reflection coefficient with angle of incidence For curves 1, $\alpha_2/\alpha_1 = \rho_2/\rho_1 = 1.25$; for 2, 1.11; for 3, 1.0; for 4, 0.9; and for 5, 0.8. (From Sheriff and Geldart, 1995) (a) No change in Poisson's ratio at the interface (solid curve, $\sigma_2 = \sigma_1 = 0.3$; dashed, $\sigma_2 = \sigma_1 = 0.2$). (b) Decreasing Poisson's ratio (solid, $\sigma_2 = 0.1$, $\sigma_1 = 0.4$; dashed, $\sigma_2 = 0.1$, $\sigma_1 = 0.3$). (c) Increasing Poisson's ratio (solid, $\sigma_2 = 0.4$, $\sigma_1 = 0.1$; dashed, $\sigma_2 = 0.2$, $\sigma_1 = 0.1$).

6.2.2. AVO Gradient and Intercept

Returning to a simplification of the Shuey formula (6.3) that only uses the first and second terms, the P-wave reflection coefficient as a function of angle of incidence $R_p(\theta)$ may be expressed (Castagna et al. 1998) as:

$$R_p(\theta) \approx A + B \sin^2 \theta \quad 6.5.$$

$$\text{where } A = \frac{1}{2} \left(\frac{\Delta \alpha}{\alpha} + \frac{\Delta \rho}{\rho} \right)$$

$$B = \frac{1}{2} \frac{\Delta \alpha}{\alpha} - \left(\frac{1 - 2\sigma}{2(1 - \sigma)} \right) \left(2 \frac{\Delta \beta}{\beta} + \frac{\Delta \rho}{\rho} \right)$$

where A is a normal incident P-wave reflection coefficient R_0 , called an AVO intercept, and an AVO gradient B (slope) as P in equation 6.3. The relationship between these A & B is dependent upon α , β , and ρ .

There is established framework for AVO gradient and intercept interpretation called “background trend”, relating to B/A (Rutherford & Williams, 1989, Castagna & Swan, 1997, and Castagna et al., 1998). They created four different classes of background trend (I, II, III and IV) that vary with different assumed parameters for brine and gas sand reflectors under Shale. Detail classification is summarised in table 6.1.

Table 6.1 The AVO behaviour of the gas and sand classes (after Castagna, 1997)

Class	Relative Impedance	A	B	Amplitude Vs. Offset
I	Higher than overlying unit	+	-	Decreases
II	About the same as the overlying unit	+	-	Increase or decrease; may change sign
III	Lower than overlying unit	-	-	Increase
IV	Lower than overlying unit	-	+	Decrease

In general, the background trend B/A becomes more positive with increasing background α/β . If we assume β is a constant value, then only with increasing α does the background trend B/A becomes more positive. The case of very high α/β , as would occur in a shallow reflection of very soft brine saturated sediment, the background trend B/A becomes positive (Castagna et al., 1998). In sand classification, Rutherford and William have classified that case, where B/A is negative and rises with increasing offset (Castagna & Swan, 1997; Al-Ghamdi et al, 1998; Castagna et al. 1998). This is graphically proved by Castagna (1997) from the physical model that shale over brine and gas sand, as class IV, is B/A negative and increases with increasing offset, where B is positive and A is negative.

Here, the research is applied to the case of a less well defined background relationship with a near surface water table (not in the background trend class criteria). In this case, we made an approach from class IV as brine sand. Then we proposed to expand a new class as ‘IV-plus’ for water saturated sandstone where B/A is positive and

increases with increasing offset where both B and A are positive. This approach is still using the term of class IV in this expanded classification, because the background trend of B is positive trend. The term 'plus' is for the extra case where water table is at near-surface. Note, keep in mind that it is only for water saturated sandstone that we have calculated in chapter two, the coefficient reflection (R_0) Vs porosity. In this AVO study we suppose that R_0 is positive and it will increase with increasing offset then B/A becomes positive.

6.2.3. AVO and Poisson's ratio

Referring to the Shuey formula (6.3), we can see that there are three variables that play important roles in the plane boundary of two isotropic media (Ostrander 1984, Santos et al, 1996; Castagne & Backus, 1993). They are P-wave velocities, densities, and Poisson ratios of the two media. Shuey (1985) chose Poisson's ratio to simplify his equation (6.3) by eliminating the P and S wave velocities. At the same time, Ostrander (1984) also used Poisson's ratio as changing parameter in specific case (sand gas) related AVO (figure 6.3).

Recently, Santos et al. (1996) used the empirical relation between P wave velocity and Poisson's ratio, provided by Wren (1984), to identify fluid saturation content in a reservoir (figure 6.4). He used a numeric analytical solution to estimate Poisson's ratio from the amplitude observation. We can write the relation between Poisson's ratio and the ratio of P-wave to S-wave velocity (α/β) as:

$$\sigma = \left(\left(\frac{\alpha}{\beta} \right)^2 - 2 \right) / \left(2 \left(\frac{\alpha}{\beta} \right)^2 - 2 \right) \quad 6.6$$

Figure 6.5 shows the relationship between σ and α/β . It is clearly seen that σ will decrease with decreasing α/β , and thus with decreasing α if β is constant. It also

easy to understand that every change in one of the physical parameters of a rock (e.g. lithology, porosity, pore fluid content, pressure) which affects α/β will also change σ (Santoso et al, 1995). In particular σ is very sensitive for α/β at range of 0.8 to 2.5.

Here, this technique will be used to identify reflections from either lithological or hydrological boundaries due to the fluid content of the rock (sandstone). The observed AVO curve will be used to estimate the AVO intercept and AVO gradient, from which Poisson's ratio will be calculated. Measured β from the refraction S-wave record will be used to confirm the calculated Poisson's ratio using σ versus α/β curve.

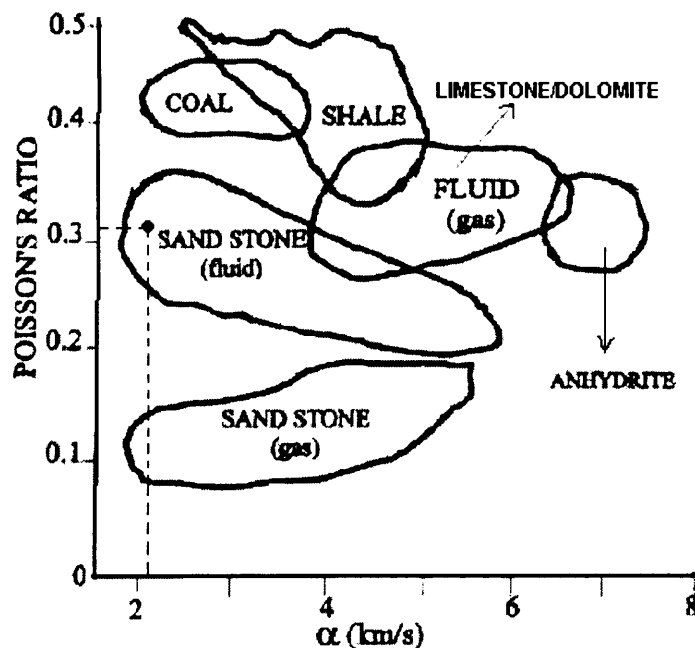


Figure 6.4 The relationship between P wave velocity and Poisson's ratio for some fluid saturated reservoir (after Santoso et al., 1996)

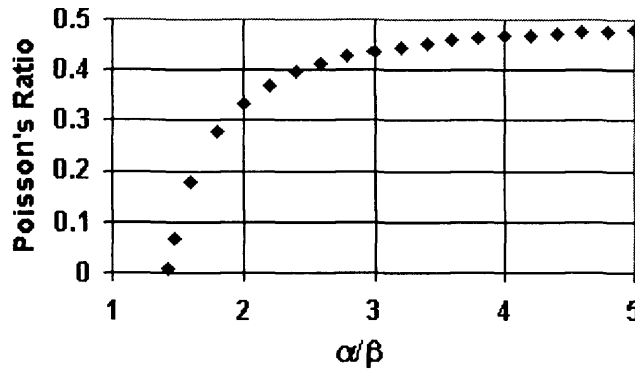


Figure 6.5 plot of Poisson's ratio versus α/β (The calculation is in the Appendix G)

6.2.4. Numerical Implementation

Clearly the mathematical AVO model is different from the physical model. The physical model (figure 6.1) shows the ideal natural condition. The model can treat the signal continuously from normal incidence to wide angle (90°) in the laboratory's experimentation, and the 1st and 2nd critical angle should appear. However, the mathematical model cannot cover the complete formulation as a complex form (Koefoed, 1962). Only the real part can be used in practice and this has to be approximated for three zones: normal incidence, intermediate angle, and wide angle (Shuey 1985, Ostrander, 1985). In this case study we only used intermediate angles 24° was the first critical angle and 35° was the widest data provided. Consequently, we cannot see any critical angle reflection.

The AVO curve for three different approximation formulae (Aki & Richards 1980; Shuey 1985; and Hilterman by Sheriff & Gekdart 1995) was calculated for a water table reflection in a pure sandstone with 30 % porosity. Figure 6.6 shows the calculated curves of different formulae for intermediate angles. Each formula has different specifications. Aki & Richards (1980) reported that their formula is only valid when; $\Delta\alpha/\alpha$, $\Delta\beta/\beta$, and $\Delta\rho/\rho$ are small and $\theta < 10^\circ$ if $\alpha_1 < \alpha_2$. Although the amplitudes are not the same for normal incidence, the lithology and water table curves have opposite trends (decrease and increase respectively curve of A&R-lit and A&R-wt).

This is because Aki & Richards formula does not separate the R_o factor while the other formulae do.

The implementation of the Shuey formula in this case study gives the value of the dimensionless parameter of $P = -0.1$ for lithology and $P = 0.9$ for water table. It can be seen in figure 6.2 that they have proven opposite trends the water table curve increases to nearly twice R_o at 40° , while the lithology decreases slightly. In other words, the curve of S-wt shows that the relative amplitude increases sharply as the water table. For the same effect the curve of S-lit decreases. This is important evidence that this formula can indicate water table anomalies. The middle term of the Shuey formula (eq.6.3) controls R_p , and the last term is always positive.

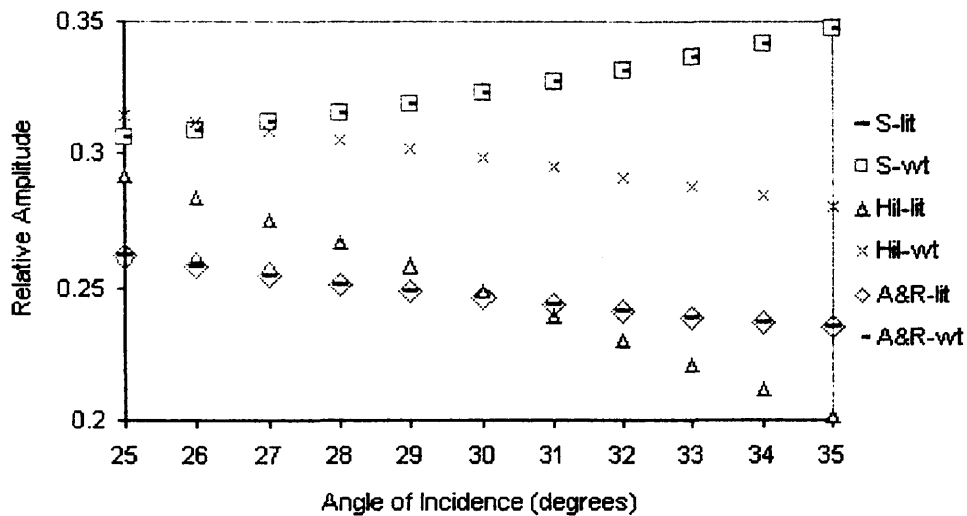


Figure 6.6 The curve of AVO for water table in typical sandstone of 30 % porosity using; 1. Hilterman formula (eq. 2.32) as Hlit and Hwt, 2. Shuey (2.31) as Slit and Swt, and 3. Aki & Richards (2.29) as ARlit and ARwt $R_c=0.2$, lithology velocity $\alpha_1 = 840$; $\beta_1 = 0.5 \alpha_1$; $\alpha_2 = 2050$; $\beta_2 = 0.5 \alpha_2$, Water table velocity $\alpha_1 = 840$; $\beta_1 = 0.5 \alpha_1$; $\alpha_2 = 2050$; $\beta_2 = 0.5 \alpha_1$

This implementation of the Shuey formula produces a value $R_o = 0.56$, very close to the coefficient reflection value from R_c versus porosity curve of water saturated sandstone that gives $R_o = 0.53$ for 33% porosity. There is a reasonable agreement

between the ideal calculation of the R_c versus porosity curve and the mathematical implementation of Shuey formula.

The Hiltermann (eq.2.32) approximation also can be used for this application. For the Lithology case, the H-lit curve is very close to the S-lit curve at most angles. However, the Water Table gives rather a different result, in that the H-wt curve is lower than the S-wt. This is because the Shuey approximation is more relevant for a case using intermediate angles. On the other hand, Hiltermann uses the approach of half space of velocity ($V_2 = 2V_1$). Both Shuey and Hiltermann approximations clearly prove that the lithological boundary and water table (as hydrological boundary) have opposite trends, decreasing and increasing respectively. This also indicates that $R_p = R_o$ for $\theta \approx 0$.

6.3. Factors affecting Seismic Amplitude

There are other factors affecting seismic amplitude, including offset dependence or independence, that must be removed or attenuated by processing (Adriansyah & McMichan, 1998; Gelinsky & Shapiro, 1997).

A. Factor without offset dependence (noise)

1. Random noise
2. Instrumentation
3. Source/receiver coupling (always checked before recording)
4. Mode conversions

B. Factor with offset dependence (noise)

1. Source/receiver directivity including ghosting and array response
2. Emergence angle
3. Coherent noise, multiples
4. Spherical spreading
5. Processing distortion, NMO errors and Stretch
6. Inelastic attenuation and anisotropy

7. Transmission coefficients and scattering above target.
8. Structural complexity
9. Near surface structure

6.3.1. Geometrical Spreading

Equation 2.52 indicates that geometrical spreading causes the intensity (I) and energy density (E) of spherical waves to decrease inversely as the square of the distance from the source. For a homogenous medium, wave amplitude will decay proportionately as $1/r$ and energy density as $1/r^2$, where r is the radius of the spherical wavefront. For layered earth, amplitude decay can be approximated by $1/(V^2(t).t)$ (Yilmaz, 1988), where t is the two-way travel time and $V(t)$ is the rms velocity of the primary reflection (no multiple) average over a survey area.

From that approximation, the gain function for geometric spreading compensation is defined by :

$$g(t) = (V(t)/V(0))^2 (t/t(0))$$

where $V(0)$ is the velocity value at a specified time $t(0)$. In practice, velocity usually increases with depth, which causes further divergence of the wavefront and a more rapid decay in amplitude with distance. Moreover, the frequency content of the initial source signal changes in a time-variant manner as it propagates. In general conditions, high frequency is more rapidly absorbed than low frequency. This is because of the intrinsic attenuation in rock. This attenuation can be corrected using wavefront-spreading correction provided in the flow of Offset Amplitude Recovery.

6.3.2. Near Surface Effects

The effects of the near surface on seismic amplitude are varied. A variation of source strength and receiver coupling can modify AVO. A special case will occur when the field has a wide variety of near surface impedance or surface conditions (Rutherford and Williams, 1989; Castagna & Backus, 1993). Other considerations are emergence angle dependent effects that include; source radiation pattern, geophone response, and array response. These affects can also be corrected by Offset Amplitude Recovery processing (Mojonero et al, 1999).

6.3.3. Equipment Tests

The instrumentation has been tested for a variety of geophone and seismograph channel sensitivities. Channel tests have been conducted using a signal generator to produce a signal which is recorded by the seismograph as a synthetic shot record. This data is loaded into a personal computer (PC) for transferring to the Promax processing system. The tests used a constant voltage with a constant amplitude for four different frequencies 10, 50, 100, and 400 Hz. This is assumed to cover the dominant frequency of the field study (100 Hz and 40 Hz for P and S waves respectively). For a 48-channel system, four records from different frequencies have been picked. Figure 6.7 shows the average value of the amplitude record from all difference frequencies gives 0.3 % error bars as 1 standard deviation.

Geophone tests were also conducted in Victoria Park, behind the Geology Department at Leicester University. Using 24-channel geophone cable, all 58 geophones were tested, divided into three gathers, where the first and second gathers were 24 channels and the last gather was only 10 channels. Each gather was fired using five different offset distances of 10, 20, 30, 40, and 50m. The total records considered of 15 shot records with 290 traces from 58 geophones for each offset position. After data was loaded to the PC and transferred to the promax, all traces were picked for the first

arrival signal. This picking was done without any special processing relating to AVO analysis. However, the study uses a normalisation factor to overcome the spherical spreading attenuation and in-constant source energy. The average value from each of the 58 geophones in one offset distance was used as a normal factor. Then each amplitude was divided by this factor with the result being about 1.0. Figure 6.8 shows the average of five different amplitudes from differences offset distance. This statistic gives a standard deviation of 0.097 with error bars of 9.7 %.

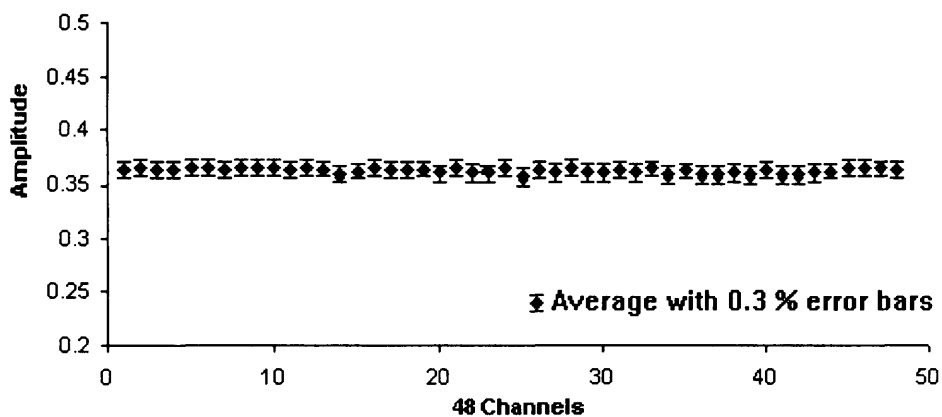


Figure 6.7 the result of Seismograph channels test for any consistence signal at four different frequency of 10, 50, 100, and 400 Hz.. Each data is averaged from 4 picked amplitudes and 0.3 % is average of each standard deviation.

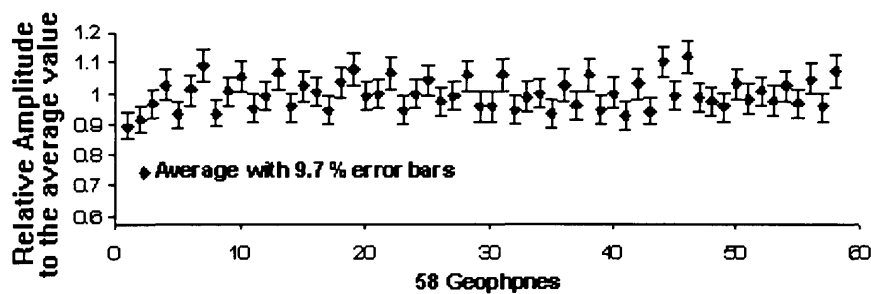


Figure 6.8 the result of Geophones test for first arrival signal at five different offset distance of 10, 20, 30, 40, and 50 m. Each data is averaged from 5 picked amplitudes and 9.7 % is the average of each standard deviation.

6.3.4. Processing System Test

The Promax processing system has been tested with reference to the sensitivity of amplitude in several processing operations such as filtering. In this case, synthetic traces have been generated for a reflector at 110 ms using 0.5 as a reflection coefficient. Figure 6.9 indicates the original trace and after F-k filter, Top mute, and Band pass filter. It is clear that the deviation of amplitude due to these processes is less than 2 %. The only sensitive filter parameter is the window of filter itself, for the f-k filter this is the range of velocity and frequency, and for bandpass filter this is the range of frequency only. This is in accordance with the purpose of the filters, whose output depends on the filter window. Therefore, the processing system without dependency of filter window only gives an amplitude deviation of less than 2 %.

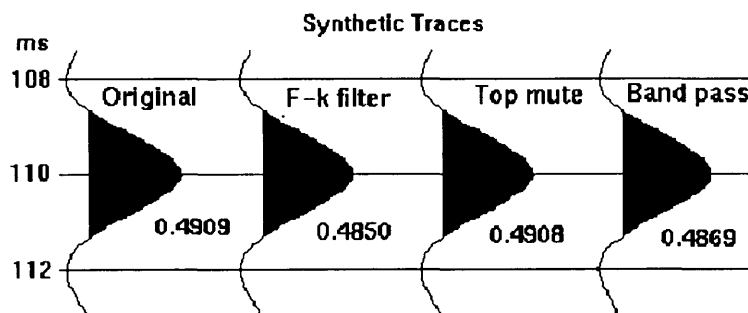


Figure 6-9 the result of Promax test, using synthetic traces the amplitude have been picked after three filtering. The deviation numbers are 1.20 %, 0.02 %, and 0.81 % for f-k filter, top mute, and bandpass filter respectively.

6.4. Seismic Processing for AVO Analysis

The main purpose of this processing for AVO analysis, is to prove that the amplitude of the signal from the water table reflection will increase with the Offset. This is based on the idea that a water saturated sandstone causes a sharp rise of P-wave velocity the while S-wave velocity is nearly constant. This analysis is commonly

utilised in oil exploration to identify reflectors that are hydrological boundaries (oil or water), before exploitation.

Common Mid Point (CMP) gather is the best technique to analyse (Ostrander 1984, Sheriff et al. 1995). The main consideration of CMP gather is to generate a single trace to CMP position for all the offset shooting. In this case we can assume that the medium is more homogeneous than when using shot record, which creates different reflection points for each offset. The main disadvantage of shot record is the assumption that all source energies are consistent. When all source energies are nearly constant, the shot record may be appropriate i.e. to use one source for all offsets.

However, this study will use the common offset gather. This will combine, (1) the stacking principle using CMP gather as based on the homogenous assumption, and (2) using a shot record with an assumption of constant source energy in order to optimise the amplitude recorded for AVO analysis.

6.4.1. The Principle Points of AVO Analysis

There are five principal points that can be considered :

1. A weak target reflector can be intentionally illuminated at the critical angle to obtain a stronger signal.
2. $\alpha_1 > \alpha_2$ (no critical angle) boundary can be identified by polarity of reflection.
3. A salt reflection can be identified by a characteristic increase of amplitude with offset caused by an increase in velocity, accompanied by a concomitant decrease in density across the interface. (Anstey, 1977).
4. Generally, water saturated rock can be identified by an AVO characteristic without the second critical angle due to $\alpha_1 > \beta_2$.
5. Because we cannot apply Knott's formula continuously at $0 < \theta < 90^\circ$ due to a complex solution term, we cannot see any critical angle on both Aki & Richard and Shuey implementations, which have conditional terms for normal incident, intermediate, and wide angles. In this case, use the intermediate angle $0 < \theta < 30^\circ$.

6.5. Application of the AVO to Edwinstowe Record

The initial aim of this application is to identify the reflection expected as the fluid (water) boundary in a saturated sandstone. This boundary will be indicated by an increase in the amplitude of the signal boundary with offset. The second aim is to see that the water table reflection can be proved by the background trend as a IV-plus classification, (a new proposed classification). The third is to show that plotting Poisson's ratio-Velocity diagram, provided by Wren (1984) to identify fluid saturated content, can also confirms the result after the calculation of Poisson's ratio of the rock.

6.5.1. Observed AVO of Edwinstowe Section

Part of Edwinstowe common offset gathers have been chosen for AVO analysis. Common offset gather for this case study is better than CMP gather. A possible reason is the homogeneous nature of the subsurface geology. There are some processes required to reduce factors that affect the seismic amplitude, especially factors with offset dependence. These processes are to make corrections that become major problems in the Edwinstowe record in relation to the AVO analysis, these include:

1. Spherical divergence correction
2. Emergence angle correction
3. Static correction
4. F-k Filter
5. Gain pre-NMO correction
6. NMO correction

These corrections mainly reduce; spherical spreading, emergence angle, near surface structure, and coherence noise. In order to measure the AVO on a sample by sample basis (as in this observation) it is important to correct the CMP gather for NMO correction (Spratt, 1987; Castagna & Backus, 1993; Mojonero, 1999). In this case, the velocity function for NMO correction has been changed using a trial and error method

for this special purpose. The velocity function becomes 1:0-900, 70-1070, 150-1500, 200-1800.

6.5.2. Amplitude Record

As mentioned in the section 6.4 the study will use common offset stacking to produce the gather for AVO analysis. Figure 6.10 shows a gather as a result of common offset stacked from 15 shot records (FFID 21-35 see appendix F) for offset 34 – 48 m. This analysis only uses the range of intermediate angle. In this case we take $> 28^\circ$ (after critical angle at 24°) converted from 34 m offset with a 30 m depth reflector as water table boundary. The maximum offset we have is 48m converted to 35° .

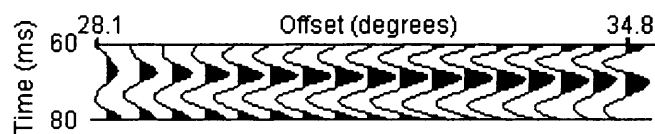


Figure 6.10 a part of Edwinstowe record as the result of 15 Common Offset stack
(Raw data enclosed in Appendix F)

Referring to the physical model (figure 6.1) as a representation of a complex formulation for normal or natural condition, figure 6.11 is representative of the Edwinstowe condition that has a compression velocity ratio V_2/V_1 of 2.5. The intermediate angle range is from 24° (as the first critical) to 50° . Observational data is only available for the range of angles from 24° to 35° . Within this more limited range, the reflected energy decreases sharply with increasing angle of incidence. The model curves in fig 6.11 are computed for boundaries with a lithological contrast, where the

partition of energy is mainly controlled by the properties of the matrix of the rock and not its fluid content.

Figure 6.12 illustrates the result of AVO record from figure 6.10 and mathematical model of both lithological and hydrological boundaries (lines). The stack-wt (figure 6.12) is the result of stacking using common offset by Promax, while the manu-wt plot is the result of manual averaging from 15 shot records (raw data, see appendix F). Both stack and manual plots have the expected trends i.e. increasing amplitude with increasing offset. These curves increase sharply from 28.1° until 32° . This increase is good evidence that the reflector is a hydrological boundary (water table), but if the curve decreased with increasing offset then it would be a lithological boundary (see figure 6.12, solid and dashed line are the mathematical model for water table and lithological boundary respectively).

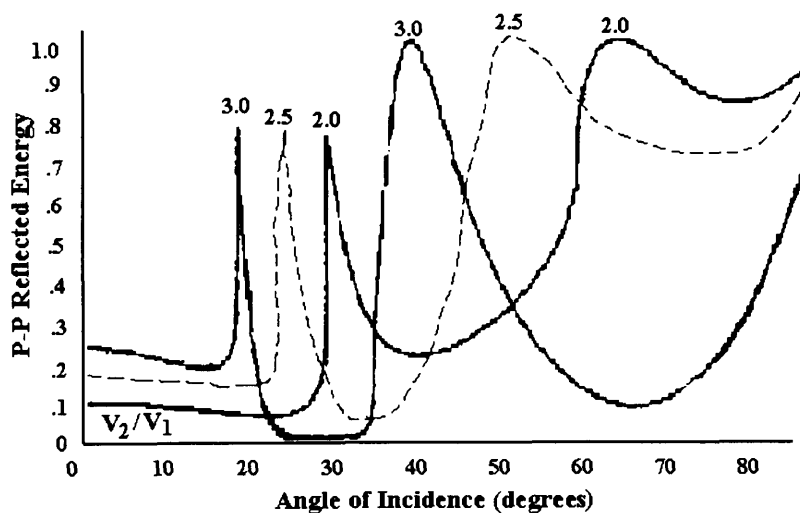


Figure 6.11 The effect on the reflected compression energy of the approximation of compression velocity ratio (V_2/V_1) of 2.5 (dashed line) for Edwinstowe (after figure 6.1)

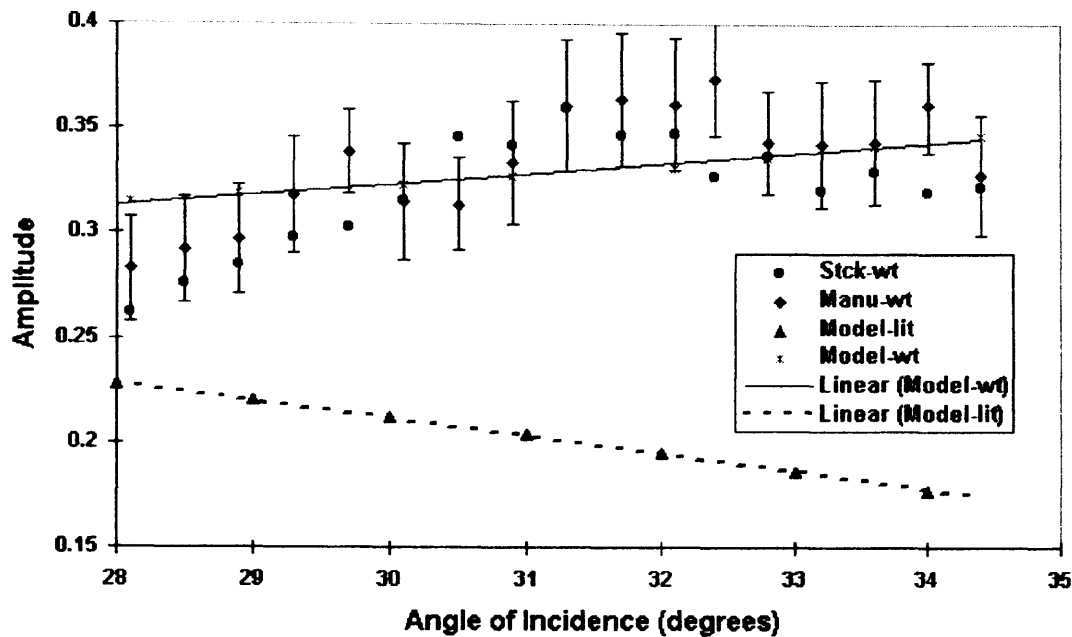


Figure 6.12 AVO analysis of Edwinstowe record as average (promax stacking as Stck-wt and manual averaging as Manu-wt) of 15 common offset gathers with average deviation of 17.5 % as error bars, comparing to the mathematical model using Shuey formula as lithological and water table boundary (Model-lit and Model-wt)

The error bars, from standard deviation of the averaging record (from manual picking) is 17.5 % with a maximum deviation of 29 %. Statistically, this distribution record is normal, as indicated by the maximum deviation among the data being greater than its standard deviation. The instrumentation test gives a result of the seismograph channel having an average deviation of 0.3%, and all geophone tests gives average deviations of 9.7%, with processing system tests indicating deviations of less than 2%. Thus it is reasonable to conclude that the variation of amplitude is statistically caused by variation in offset, not the instrumentation. To confirm this, it will be proved by AVO analysis of a known lithological boundary, which shows that the amplitude decreases with offset.

Regarding the AVO analysis for hydrocarbon exploration, the range of variation of amplitudes is only before the first critical angle. For example, Requeiro (1993) published the result of his investigation of a study area in Los Loundos. The target was

a shallow gas reservoir that has the following boundaries; Coal, Gas-Sandstone, and Water-Sandstone for a depth less than 1200 m. The sandstone porosity was 18 %. He presents AVO curves that represent the amplitude variations with distance, increasing in both directions from the centre of split-spread shooting.

Figure 6.13 shows that the range of amplitude record is up to 30° , which is below the first critical angle. This is different to the Edwinstowe AVO analysis, which applies to the intermediate angle i.e. after the first critical angle and before the second critical angle. Furthermore, the curves in figure 6.13 indicate that; coal curves sharply decrease, whereas gas-sand curves increase sharply. Water-sandstone curves show a steady increase. This is evidence that the water-sandstone boundary has specific characteristics, and this has been developed for detection of water table reflection in shallow depth target.

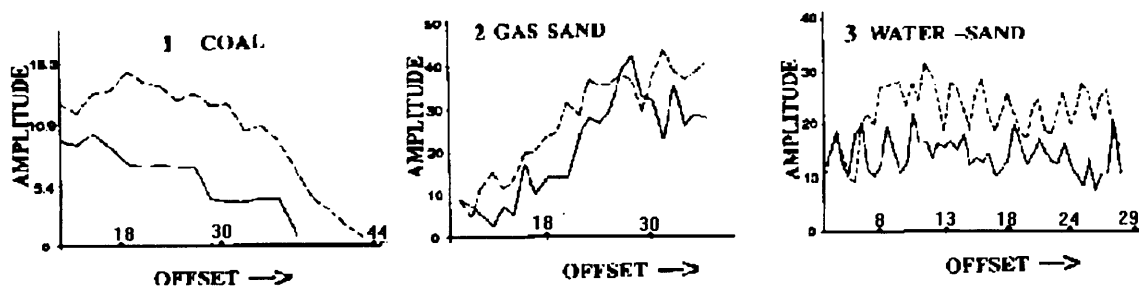


Figure 6.13 AVO curves showing trends for the difference reflector in the study area of Los Lanudos, The solid and dashes curves represent the amplitude variations as distance increases in both direction from the centre of split spread (after Regueire, 1993).

6.5.3. Background Trend

Referring to the AVO Gradient (B) and Intercept (A) at subsection 6.2.2 we will see that the Edwinstowe record has background trends or sand classification of 'IV-

plus' of water saturated sandstone. Figure 6-12 clearly indicates that both B (polynomial average) and A are positive. This indicates that the curve has a background trend or sand classification of 'IV-plus'. There are two important points in the curve, (1) a positive value of A around 0.5 is a very strong reflector, and (2) positive value of B with increasing amplitude with offset as AVO anomaly for the appropriate reflection boundary. This boundary is appropriate for the zero-offset of reflection coefficient value (0.5) and has been confirmed as water-sandstone with a large porosity (32 %) in this seismic interpretation.

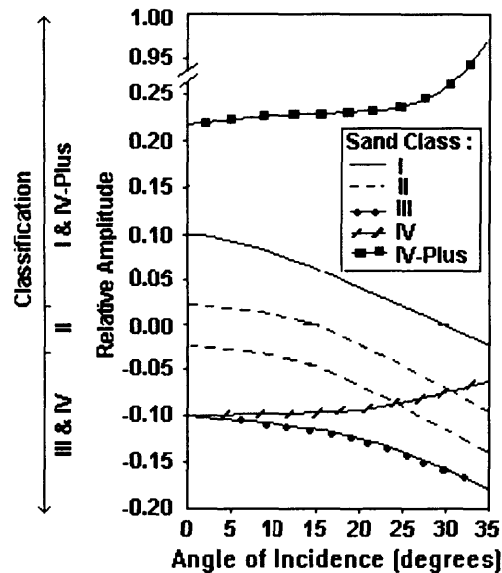


Figure 6.14 the background trend and classification of AVO analysis (after Castagna et al, 1998)

Figure 6.14 shows the background trend and classification of brine (fluid) and gas sand established by Rutherford & William (1989) for class I, II, and III. Class IV was proposed by Castagna & Swan (1997) and class IV-plus is proposed in this study for water saturated sandstone as a water table. According to the curve of reflection coefficient (RC) versus porosity in chapter 2, the reflection coefficient of sandstone with porosities of 10 to 40% (normal range of sandstone porosity) is between 0.22 and 0.57. In this case we put the minimum porosity of 10 % as 0.22 of reflection coefficient, see figure 6.12.

6.5.4. Poisson's Ratio

As mentioned in section 6.2.3, the calculation of Poisson's ratio can be used to identify the characteristics of a layer beneath a reflection boundary prior to fluid content. Equation 6.6 can only be used to calculate the Poisson's ratio of the first layer as a function of P-wave and S-wave velocity. S-wave velocity in m/s can be taken from S-wave test in figure 5.25 as 800m/s. By assuming that we have an ideal S and P-wave sandstone ratio of 0.65, then Poisson's ratio can be calculated as 0.1. This is an appropriate number for unsaturated sandstone, similar to that for a gas sandstone Wren curve (Santoso et al, 1996).

The main purpose of this sub-section will be to calculate Poisson's ratio for the second layer. We cannot use eq. 6.6 because we do not have information about S-wave velocity in this layer. We will apply the gradient and intercept AVO record to the Castagna et al (1998) formula (eq. 6.5) as a simplification of the Shuey (1985) formula (eq. 6.3). In contrast to Santosa et al. (1996) who made the estimation of Poisson's ratio using an analytical solution, here Poisson's ratio has been calculated using a numerical solution. From the implicit formula of gradient and intercept AVO (eq. 6.5) we have made the iteration in order to get the convergence condition. There are many convergence conditions dependence on the fixed parameter. The best convergence condition gives the following results:

Poisson's ratio of second layer: 0.34

Poisson's ratio of first layer: 0.1

P-wave velocity of second layer: 2050 m/sec

P-wave velocity of first layer: 1070 m/sec

S-wave velocity of second layer: 800 m/s

S-wave velocity of first layer: 800 m/s.

This result is confirmed by the Wren curve as the position of fluid sandstone (see figure 6.4) for the second layer of (α_2 , σ_2) as (2050, 0.34). In this case we can directly interpret this as a water saturated sandstone (the water table).

6.6. Application of the AVO Analysis to Croft Record

6.6.1. General Consideration

The main purpose of this section is to apply AVO analysis to a reflection from a lithological boundary. This boundary will prove that the amplitude decreases with increasing offset in the intermediate angle (between P and S critical angles). Previously, we have proven that the amplitude increased with increased offset for water table reflection as hydrological boundary.

This investigation needs data with appropriate post critical reflection, which we can analyze the trend of amplitude Vs offset for the same purpose as Edwinstowe survey. Croft records that had been taken one year before Edwinstowe are appropriate data for this purpose. This has a simple geophysical target, of a flat lying reflector. The records have a maximum offset of 136m, which is adequate when using AVO analysis for a depth of target of about 100 m. This gives us a chance to analyze ranges beyond the critical angle.

6.6.2. Geology of Croft Site

The field area is located near Croft Quarry (SP 523 956) South Leicester. The site is a small field west of Coventry road, beside the quarry entrance. Figure 6.15 shows the plan of the shot line of seismic survey. There is that there is a covering of 100-200m of bedded sediment (Mercian Mudstones underlying Sherwood Sandstone) unconformably overlying a granitic intrusion. Mudstones are flat lying, and no raypaths we considered to have entered the granite. This is geologically similar to the Edwinstowe location analysed for water table reflection. The primary interest of this site is to determine the depth of Mesozoic/Paleozoic cover over diorite rock east of Croft Quarry.

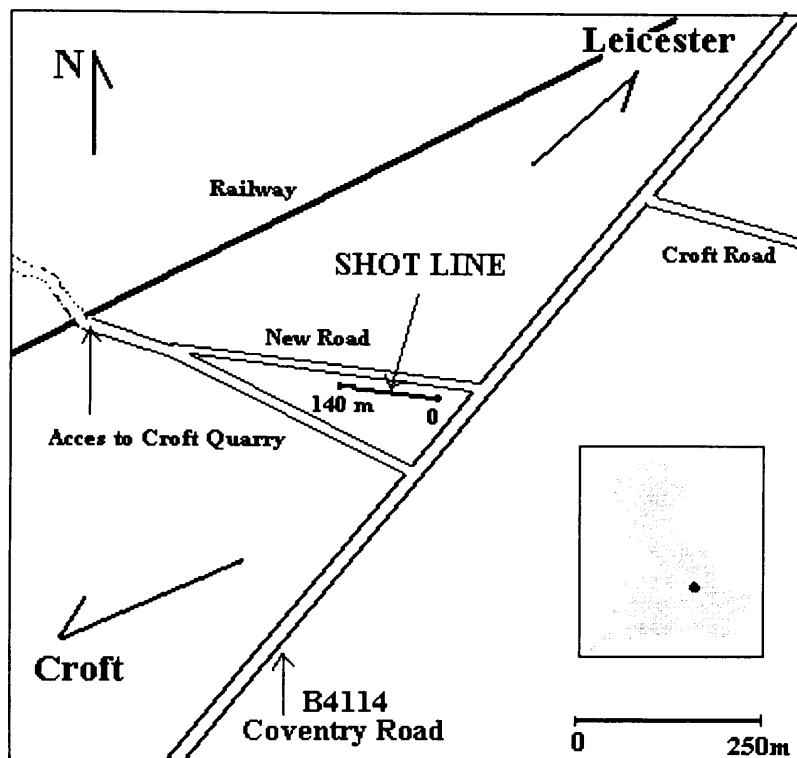


Figure 6 15 The plan location of shot line of seismic survey

6.6.3. Data Acquisition

Data was recorded along a line 140 m long orientated east to west on the west side of Coventry Road (see figure 6.16). The spread was shot rolling through fixed geophones also aligned east to west. Fifty-one shot point positions were taken shooting from 0 to 140 m, with fixed geophones at 2 m spacing from 0 to 94m. Thirty-one shots (0-60m) were recorded with 2 m shot spacings and 48 fold of cover. The last 20 shot points (60-140 m) were recorded with 4m shot spacings and 24 fold of cover. Both of them have 2-m geophone spacings. The source was 5 stacks of a sledgehammer at each shot position. The list of equipment and field recording parameters are given in table 6.2.

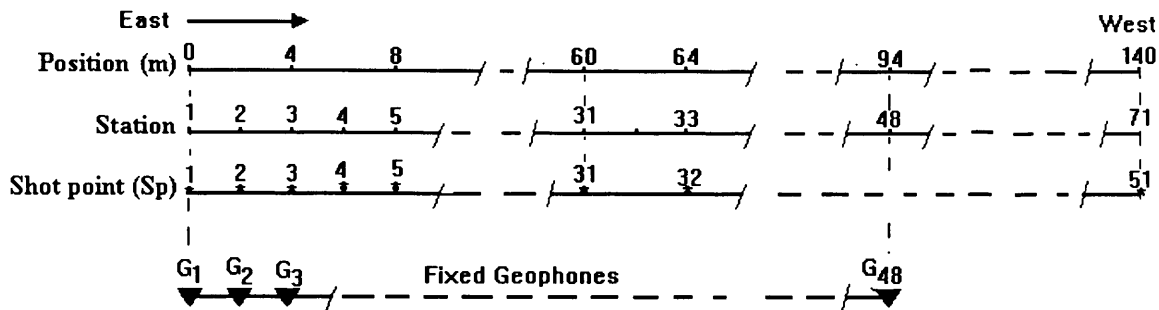


Figure 6.16 Survey lay-out of shallow seismic reflection at Croft

Table 6. 2 Technical description of the seismic data location, Equipment, and acquisition parameters.

SITE	
Site Name	: Croft Quarry
Geology Rock	: Dioritic rock overlain by sedimentary Mercian Mudstones
Depth Range	: 0-150 m
Date of Survey	: 12/11/1996
EQUIPMENT	
Seismograph	: Bison 9048
Geophone type	: Mark Products 100 Hz., And 125 mm spikes.
Source Type	: Hammer 5 stacks with plastic plate
Record length	: 500 ms.
Sample interval	: 0.5 ms.
Record Filters	: Locut: 100 Hz, Hicut: 1000Hz.
GEOMETRY	
Spread type	: Shot rolling through fixed spread
No. of channels	: 48
Length of line	: 140 m
Geophone spacing	: 2 m
Shot spacing	: 2 m or 4 m
No. of shots	: 51
Shot depth	: 0
Shot offset (X)	: 0 – 46 m
Shot offset (Y)	: 1m
Nominal Fold	: combination of 24 and 48
Length of Max Fold CDP	: 2 m with 40 fold
Nominal CMP spacing	: 1 m

6.6.4. Processing and Velocity Analysis

The processing sequence is similar to the processing for both the synthetic and the Edwinstowe record and is summarized in table 6.3. Figure 6.17 shows the final stack. The section shows that there are three simple layers with first and second boundary at 100 and 150m respectively, meaning the basement starts in the third layer. This also indicates that the zone below this boundary has no significant layering. This may be due to the very limited length of record.

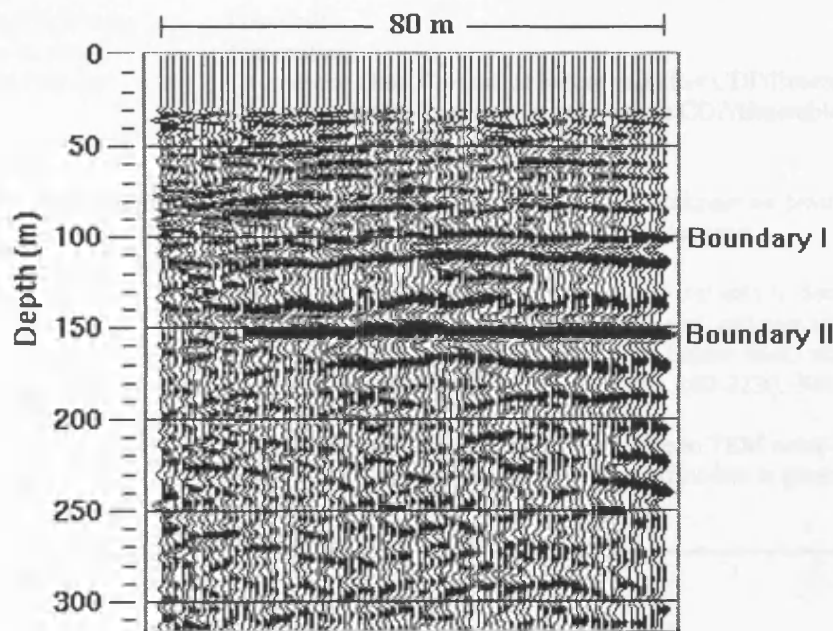


Figure 6.17 The final stack of Croft record

Velocity structures of those three simple layers have been derived based on refractor velocity and picking on semblance velocity analysis. The first layer as the top one is fixed by refractor velocity when doing the static correction that has 1900m/s with a thickness of 70m. This has been confirmed as a result of the stacking, and when the velocity at around 100m depth (first boundary) was changed to 2000m/s the results

becomes unstable and the boundary goes down sharply. However, if we use 1950m/s the result is constant (as for 1900m/s). This indicates that the first layer is nearly homogeneous with a velocity at 1900 to 1950m/s.

Table 6.3 Processing sequence for the data of Croft record

Geometry	: Input high and plan position of all source points and geophone position. Datum high used was the highest position that 98.4 at East end line.
Bining	: Matching pattern number using first live channel and station.
Field Static	: Elevation and Refraction static correction
Editing :	
- Trace kill/reverse	: Trace kill
- Trace muting	: Top muting
Gain and Polarity	: - Automatic Gain Control at before and after CDP/Ensemble stack - True Amplitude Recovery at only before CDP/Ensemble stack
Filter:	
-Band pass	: 50,60,300,400 Hz.
-Air Blast Attenuation	: 50, 100, 800, 1000 (frequency range to enhance air blast energy)
-F-k	: F-k analysis; -800, 800 m/s : 60, 400 Hz. (accept)
-NMO	: Velocity function from Velocity analysis (Velocity Analysis: Interactive analysis of Velocity to determine velocity function for moveout correction and estimate of velocity for depth conversion that have been done before NMO after f-k filter, Picked velocity : 0-1900, 120-1950, 160-2230, 300-3000)
No. of CMP's	: 118
Stack	: Mean stack, to exclude 12 shot record due to TEM noise in the same time measurement and first 30 m offset due to ground roll energy.
Output	: Variable area wiggle plot

The velocity structure of the second layer is derived from the semblance velocity (as Root mean square velocity, V_{rms}) of the second boundary in the velocity analysis. Using Dix formula (Equation 2.24), the interval velocity (V_{Int}) as the second layer velocity can be calculated. In this case, $V_{rms} = 2230$ m/s then the $V_{Int} = 4150$ m/s. The error barr of the V_{rms} (as semblance Velocity) observation is $< 7 \%$ (to derive this see section 2.3.3.1). As mentioned in the paragraph before the third layer is the basement, and there is no information about velocity except that it is normally higher than the above layer.

The prominent reflector at the first boundary at 100 m is interpreted as a contrast in velocity from 1900m/s to 4150m/s. The reflector is very difficult to interpret as the depth of Mesozoic/Paleozoic (Mercian Mudstone) cover over dioritic rock East of Croft Quarry, is between 100-200m thick. The velocity of the second layer is too high for dioritic rock. Regarding to geologists within the department and the consensus of opinion, if there is a high velocity in the Croft data then it is probably (1) Stockingford shales (Cambrian) or (2) Gypsum in the Mercian Mudstone (but not halite).

6.6.5. Amplitude Record of AVO

As described at 6.4, CMP gather is the best section for analysis (Ostander 1984, Sheriff et al. 1995). CMP gathers were also used in this observation. There are 22 CMP (numbers 50 to 72) gathers, which include the maximum offset that we have (136m).

The P-wave critical angle in this AVO analysis is calculated from the estimation of critical offset (can be seen in the pick of amplitude curve in figure 6.18) which is about 108 to 110 m and 100m of reflector depth. From this calculation, the critical angle is about 28° . This can be accepted by using the velocity contrast in the reflector, for 1900 to 4150 m/s that give the critical angle of 27° . The different of 1° is in the range of error bar of V_{rms} that $< 7\%$.

This proves that the sub-critical angle has an amplitude variation with offset by providing the curve with different ways of shooting that have similar curves. Figure 6.19 shows the curve of AveSR2&3 that has an amplitude variation with offset from East-to-West shot. The opposite direction (West-to-East shot) is illustrated by the curve of SR48 and SR49. All the curves have similar trends, going down from 66m, having a minimum of 80-90m, and then rising to the critical angle 110-116m. All of these curves have average standard deviations of 0.57 after reducing 76 % from 3.3 (also in standard deviation) by normalisation using the factor of the square root of the energy and by a random filter of 7 % wing smoothing data. This remaining error (standard deviation of

0.57) is reasonable due to the variation in geophones we have tested.

The AVO observation was carried out after the sequence of AVO processing similar to that of the Edwinstowe records. Figure 6.20 shows four AVO curves from both real data and mathematical models. Each involve lithological boundaries and water tables as hydrological boundaries from Croft and Edwinstowe records respectively. Each point has been normalized to the point at position of 28.2° . This will reduce the effect of the variation of the energy source, instrument gain, and the lateral variation of layer. Data lithology (Data-lit) has been taken from a supergather as a stack of 22 CMP gathers. Although individual CMP gathers do not give any trends the stack from the CMP gather still gives representative curves that have decreasing amplitude with increasing offset. This proves that the lithological boundary produces a negative AVO gradient.

The straight lines are the linear of four curves. We can see that both data and models of linear curves are joined at the point of 28.2° but in different places, (the model is exactly at 1 and the data is at about 1.07). This is because the linear of model point is from the data that has very a small variation, while the data point has a much wider variation. The slope of both Data-lit and Data-wt are also higher than both Model-lit and Model-wt because the real data has more complex parameters than the model. The most important consideration is that the application of AVO analysis gives sufficient evidence that amplitude increases with increasing offset for water table reflection as a hydrological boundary for both the model and real data. Conversely, the amplitude decreases with increasing offset for sub-bedding reflection as a lithological boundary for both model and real data.

6.6.6. Conclusion

The Croft data has specifications that are appropriate for application to the AVO analysis of lithological boundaries. Although several records have been excluded due to excessive noise, the data set still gives adequate CMP gathers for AVO observation. The

velocity structure of three simple layers is interpreted as geophysical interpretation, even if there is a high velocity (second layer) in the Croft data that is probably (1) Stockingford shales (Cambrian) or (2) Gypsum in the Mercian Mudstone (but not halite).

The AVO curve shown in figure 6.18 establishes that the offset is also for post critical angles, which the critical angle at about 110m offset. As mentioned before, (Edwinstowe data) the focus of our AVO analysis is the post critical angle. Therefore, figure 6.20 good evidence for both lithological and hydrological boundaries, and for mathematical models and observed data.

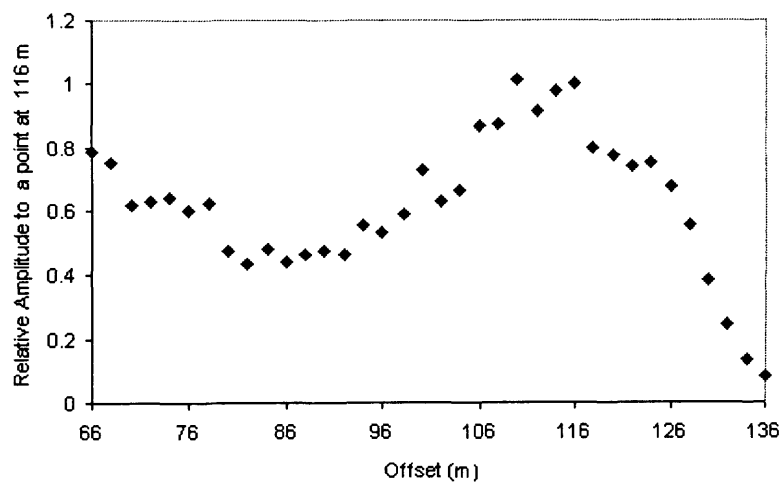


Figure 6.18 the AVO observation from the Croft record shows the P-wave critical angle at about 110 m

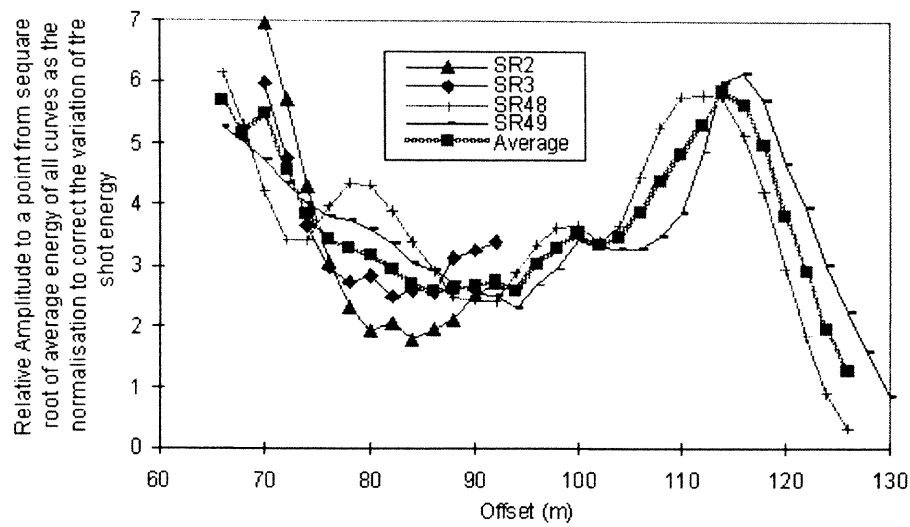


Figure 6.10 the curves of East-to-West shot (AveSR2&3) and the energy spectrum of most West-to East shot (Energy), both of them have deviation about 9 %

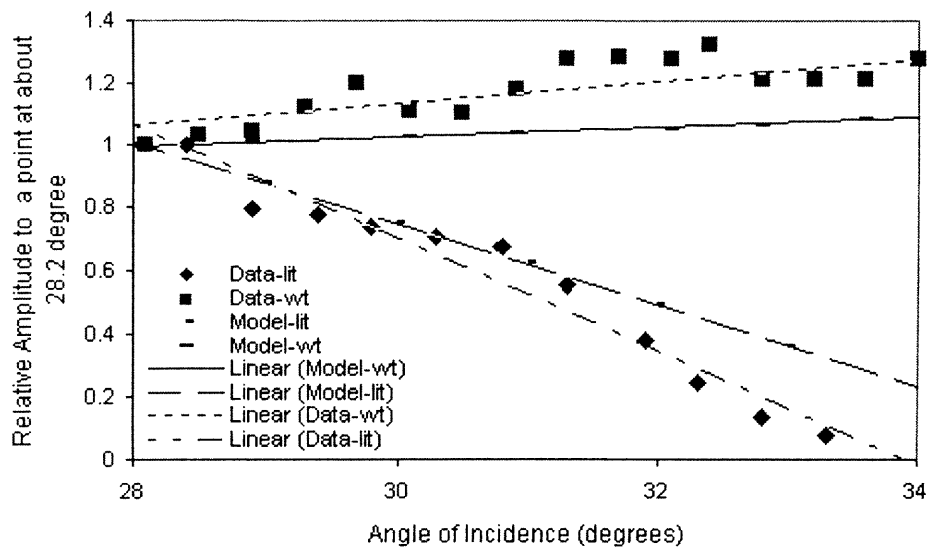


Figure 6.20 AVO analysis from mathematics model and observed data (Water table from Edwinstowe and lithology from Croft record)

Chapter 7

CONCLUSION AND RECOMMENDATION

7. 1. Shallow Seismic Reflection

The results of this study demonstrate the potential applications of Shallow Seismic Reflection Method in identifying the water table reflectors. This method has a good chance of locating the water table when using optimum field parameter and processing. The application of the method in real fields to identify water tables has been done successfully. This is supported by models, t-x curve and synthetic seismic traces. The main progress is to raise the resolution and S/N ratio. To increase the resolution for a flat-lying reflector (detecting water table), we can use a source that produces a high frequency. We can also reduce the distance of geophone interval. Meanwhile the optimum stack increases the S/N ratio.

Identification of the best field parameters is the key factor in getting optimum shot records. For example, getting the best window of record is very important. This is dependent upon the target and survey plan. Therefore, some of weaknesses in limiting parameters will be avoided such as deciding the geophone interval. If the geophone interval is too small, this will destroy the coherency of spatially aliased ground roll even if it improves coherency of true reflectors. The other field parameters that should be mentioned are maximum & minimum offset and source effort.

With the processing data, we should pay more attention to the limiting parameters in order to do process well. Three coherent noises in shallow record should be removed or attenuated; first arrival, ground roll, and airwave. Static correction and velocity analyses are extremely important. These processes that include interactive sequences should be done extra carefully.

1. Synthetic Seismic Model

A synthetic seismic model is ideal for seismic records, and the processing, and interpretation of certain objective of shallow target identification of water table reflectors. The model is very useful when designing the field parameter as a geometry input. This is a field configuration that includes type of spread, minimum and maximum offset, source and geophone spacing, and fold of cover as subsurface coverage. Promax version 7.2 has a facility to produce the traces as set of synthetic shot records.

The synthetic shot records are successfully produced and show that the model can ideally be examined to identify the water table. This is a sufficient even by one shot record (1 fold of cover). The contrast impedance of the boundary expresses the reflector. This is achieved by using $R_c = 0.15$ for the water table in pure sandstone with a porosity of 6 % as a minimum condition. This is so that the water table as a boundary can be clearly recognised, and a bigger porosity will give a clearer boundary. By producing the simulation of geometry input, then the assigned shot records will be ready to process as a normal field shot records.

Geometry input has been designed with exactly the same field record as at Edwinstowe. This made successful processing easy to compare with the field record. The processing is simpler than the real field record, but they do have similar sequences. The most important point is the static correction that to correct or to remove the additional signal on the reflector after added as the static error. The final stack shows that the model of water table in pure sandstone with 6% porosity can produce very strong reflectors at 30m with a maximum fold of cover 48 in 0.5m CMP spacing.

2. Edwinstowe Survey

Edwinstowe field and sub-surface conditions are ideal for shallow reflection recording and deal with the identification of water table reflection. The surface

topography is nearly flat with a slope of 1.3° . The geology is homogenous and pure sandstone lies below a weathering layer at around 3m depth with an estimated water table at 31 m depth (borehole record at approximately 700 m NNE on June 1981). It is believed that the processing of this survey produces the final stack, and the water table has subsequently been interpreted at 30 m depth. A one meter difference is not a too significant number if we see that the borehole record is 17 years old.

7.2. AVO Analysis

The AVO analysis performed on the Edwinstowe seismic data involved the investigation of amplitude variations with offset. The data analysed were selected for supergather as a stacked result from 15 common offset gathers that have been taken for AVO analysis. The AVO curve has an expected trend i.e. increasing amplitude with increasing offset for an intermediate angle. This rise is strong evidence that the reflector is a hydrological boundary i.e. water table. On the other hand, if the curve decreases with increasing offset then it is likely to be lithological boundary.

This analysis also has a very specific background trend, the AVO Gradient (B) and Intercept (A) of the AVO curve has a new background trend or sand classification of 'IV-plus'. This is for water saturated sandstone as just proposed by this study. There are two important points in this specification, (1) A positive value of around 0.5 is a very strong reflector and (2) a positive value of B is an increasing amplitude with an offset of the AVO anomaly for the water table as an appropriate reflection boundary.

The other side of AVO analysis is to apply the gradient and intercept AVO record to the Castagna (1998) formula, and then to calculate Poisson's ratio. This can be used to identify the characteristics of the layer beneath the reflection boundary prior to the fluid content that was provided by Wren (1985). This result is confirmed to the Wren curve as position of fluid sandstone (see figure 5.4) for the second layer of (α_2, σ_2) as (2050, 0.34). In this case we can directly interpret the water saturated sandstone, as

the water table reflector.

In order to compare to the lithological boundary, the AVO analysis has also been applied to the other field, Croft that expresses the appropriate lithological reflector at the similar intermediate angle. It is no doubt a result that the AVO curve for this purpose shows decreasing amplitude with increasing offset (figure 6.16).

7.3. Recommendations

There are three following recommendations that suggested from the results of this study:

1. It is suggested that this method be applied to a field that has more complex geological structures. For example, a layer that has a significant dip and a water table that is relatively flat lying. Therefore we can immediately distinguish them as hydrological and lithological boundaries.
2. The capability of the shallow seismic reflection method in identifying water tables opens opportunities in investigating the interface line between fresh and polluted water such as saline water in coastal areas. Therefore, it is suggested to further investigate the application of dealing with the identification of interface lines in coastal area using water table position combining with resistivity survey.
3. It is well known that the main constraint of shallow seismic reflection is the source generated noise (e.g. ground roll) hence it is recommended that further investigation is carried out in to this matter.

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Appendix A

The calculation of Porosity (%) Vs Coefficient Reflection (R_c) of water saturated rock for pure sandstone

V_m = Velocity of P-wave in matrix ρ_m = Density of matrix
 V_w = Velocity of P-wave in water ρ_w = Density of water
 V_g = Velocity of P-wave gas (air) ρ_g = Density of gas
 V_b = Average (bulk) Velocity ρ_b = Average density
 R_c = Coefficient reflection Z = Impedance

Pure Sandstone Properties

$V_m(m/s)$	$V_w(m/s)$	$V_g(m/s)$	$\rho_m(kg/m^3)$	$\rho_w(kg/m^3)$	$\rho_g(kg/m^3)$		
2500	1480	350	1850	1000	0		
Porosity	$V_b(m/s)$	First Medium $\rho_b(kg/m^3)$	$Z_1(V \times D)$	$V_b(m/s)$	Second Medium $\rho_b(kg/m^3)$	$Z_2(V \times D)$	R_c
0	2500	1850	4625000	2500	1850	4625000	0
0.01	2355.32	1831.5	4313762	2482.8882	1841.5	4572238.6	0.029088
0.02	2226.46	1813	4036578	2466.0091	1833	4520194.6	0.056519
0.03	2110.98	1794.5	3788148	2449.3579	1824.5	4468853.4	0.08244
0.04	2006.88	1776	3564220	2432.93	1816	4418200.9	0.106983
0.05	1912.57	1757.5	3361339	2416.7211	1807.5	4368223.4	0.130264
0.06	1826.72	1739	3176670	2400.7267	1799	4318907.3	0.152388
0.07	1748.25	1720.5	3007867	2384.9426	1790.5	4270239.8	0.173448
0.08	1676.25	1702	2852969	2369.3648	1782	4222208	0.193527
0.09	1609.94	1683.5	2710327	2353.9891	1773.5	4174799.6	0.212701
0.1	1548.67	1665	2578540	2338.8116	1765	4128002.5	0.231037
0.11	1491.9	1646.5	2456415	2323.8287	1756.5	4081805	0.248598
0.12	1439.14	1628	2342928	2309.0364	1748	4036195.7	0.265439
0.13	1389.99	1609.5	2237192	2294.4314	1739.5	3991163.3	0.281611
0.14	1344.09	1591	2138441	2280.0099	1731	3946697.1	0.297159
0.15	1301.12	1572.5	2046004	2265.7685	1722.5	3902786.3	0.312128
0.16	1260.81	1554	1959294	2251.704	1714	3859420.6	0.326554
0.17	1222.92	1535.5	1877795	2237.813	1705.5	3816590.1	0.340475
0.18	1187.25	1517	1801052	2224.0923	1697	3774284.7	0.353922
0.19	1153.59	1498.5	1728659	2210.5389	1688.5	3732494.9	0.366925
0.2	1121.79	1480	1660256	2197.1496	1680	3691211.4	0.379514
0.21	1091.7	1461.5	1595524	2183.9216	1671.5	3650425	0.391712
0.22	1063.18	1443	1534174	2170.8519	1663	3610126.7	0.403544
0.23	1036.12	1424.5	1475947	2157.9377	1654.5	3570307.9	0.415033
0.24	1010.39	1406	1420612	2145.1763	1646	3530960.1	0.426198
0.25	985.915	1387.5	1367958	2132.5648	1637.5	3492074.9	0.437058
0.26	962.596	1369	1317794	2120.1008	1629	3453644.3	0.447632

0.27	940.355	1350.5	1269949	2107.7817	1620.5	3415660.2	0.457936
0.28	919.118	1332	1224265	2095.6049	1612	3378115.1	0.467986
0.29	898.819	1313.5	1180598	2083.568	1603.5	3341001.2	0.477796
0.3	879.397	1295	1138819	2071.6685	1595	3304311.3	0.48738
0.31	860.797	1276.5	1098807	2059.9042	1586.5	3268038.1	0.49675
0.32	842.967	1258	1060453	2048.2728	1578	3232174.5	0.505919
0.33	825.861	1239.5	1023655	2036.772	1569.5	3196713.6	0.514898
0.34	809.436	1221	988321	2025.3996	1561	3161648.8	0.523697
0.35	793.651	1202.5	954365.1	2014.1535	1552.5	3126973.3	0.532327
0.36	778.47	1184	921708.2	2003.0316	1544	3092680.8	0.540798
0.37	763.859	1165.5	890277.2	1992.0319	1535.5	3058764.9	0.549117
0.38	749.786	1147	860004.3	1981.1523	1527	3025219.5	0.557295
0.39	736.222	1128.5	830826.7	1970.3909	1518.5	2992038.6	0.565338
0.4	723.14	1110	802686	1959.7458	1510	2959216.1	0.573255
0.41	710.516	1091.5	775527.8	1949.215	1501.5	2926746.4	0.581053
0.42	698.324	1073	749301.7	1938.7969	1493	2894623.8	0.588739
0.43	686.544	1054.5	723960.4	1928.4895	1484.5	2862842.7	0.59632
0.44	675.154	1036	699459.9	1918.2912	1476	2831397.8	0.603802
0.45	664.137	1017.5	675759	1908.2001	1467.5	2800283.7	0.611191
0.46	653.473	999	652819.3	1898.2147	1459	2769495.2	0.618493
0.47	643.146	980.5	630604.6	1888.3332	1450.5	2739027.3	0.625713
0.48	633.14	962	609081	1878.554	1442	2708874.9	0.632858
0.49	623.441	943.5	588217	1868.8756	1433.5	2679033.2	0.639931
0.5	614.035	925	567982.5	1859.2965	1425	2649497.5	0.64694
0.51	604.908	906.5	548349.5	1849.815	1416.5	2620263	0.653887
0.52	596.049	888	529291.6	1840.4298	1408	2591325.1	0.660778
0.53	587.445	869.5	510783.8	1831.1393	1399.5	2562679.4	0.667617
0.54	579.087	851	492802.8	1821.9421	1391	2534321.4	0.674409
0.55	570.962	832.5	475326.3	1812.8368	1382.5	2506246.9	0.681157
0.56	563.063	814	458333.3	1803.8222	1374	2478451.6	0.687867
0.57	555.379	795.5	441804.2	1794.8967	1365.5	2450931.4	0.694542
0.58	547.902	777	425720.1	1786.0591	1357	2423682.2	0.701186
0.59	540.624	758.5	410063.3	1777.3081	1348.5	2396700	0.707803
0.6	533.537	740	394817.1	1768.6424	1340	2369980.9	0.714397
0.61	526.633	721.5	379965.4	1760.0609	1331.5	2343521.1	0.720971
0.62	519.905	703	365493.2	1751.5622	1323	2317316.8	0.72753
0.63	513.347	684.5	351386	1743.1452	1314.5	2291364.4	0.734076
0.64	506.952	666	337630.4	1734.8087	1306	2265660.2	0.740613
0.65	500.715	647.5	324213.2	1726.5516	1297.5	2240200.7	0.747144
0.66	494.63	629	311122.1	1718.3727	1289	2214982.4	0.753674
0.67	488.69	610.5	298345.4	1710.2709	1280.5	2190001.8	0.760206
0.68	482.892	592	285872	1702.2451	1272	2165255.8	0.766743
0.69	477.229	573.5	273691	1694.2943	1263.5	2140740.9	0.773287
0.7	471.698	555	261792.5	1686.4175	1255	2116454	0.779844
0.71	466.294	536.5	250166.5	1678.6136	1246.5	2092391.8	0.786416
0.72	461.012	518	238804	1670.8815	1238	2068551.3	0.793006
0.73	455.848	499.5	227696	1663.2204	1229.5	2044929.4	0.799619
0.74	450.799	481	216834.1	1655.6291	1221	2021523.2	0.806256
0.75	445.86	462.5	206210.2	1648.1069	1212.5	1998329.6	0.812922
0.76	441.028	444	195816.5	1640.6527	1204	1975345.9	0.819621
0.77	436.3	425.5	185645.7	1633.2656	1195.5	1952569.1	0.826354
0.78	431.672	407	175690.7	1625.9448	1187	1929996.5	0.833127
0.79	427.142	388.5	165944.6	1618.6893	1178.5	1907625.3	0.839943
0.8	422.705	370	156401	1611.4983	1170	1885453	0.846805

0.81	418.36	351.5	147053.6	1604.3708	1161.5	1863476.7	0.853717
0.82	414.103	333	137896.4	1597.3062	1153	1841694	0.860682
0.83	409.932	314.5	128923.6	1590.3034	1144.5	1820102.3	0.867705
0.84	405.844	296	120129.9	1583.3619	1136	1798699.1	0.874788
0.85	401.837	277.5	111509.8	1576.4806	1127.5	1777481.9	0.881937
0.86	397.908	259	103058.2	1569.6589	1119	1756448.3	0.889155
0.87	394.055	240.5	94770.32	1562.896	1110.5	1735596	0.896447
0.88	390.277	222	86641.39	1556.1911	1102	1714922.6	0.903815
0.89	386.569	203.5	78666.89	1549.5435	1093.5	1694425.8	0.911266
0.9	382.932	185	70842.45	1542.9525	1085	1674103.4	0.918803
0.91	379.363	166.5	63163.88	1536.4172	1076.5	1653953.2	0.92643
0.92	375.859	148	55627.15	1529.9371	1068	1633972.9	0.934153
0.93	372.42	129.5	48228.35	1523.5115	1059.5	1614160.4	0.941977
0.94	369.043	111	40963.73	1517.1396	1051	1594513.7	0.949906
0.95	365.726	92.5	33829.68	1510.8207	1042.5	1575030.6	0.957946
0.96	362.469	74	26822.7	1504.5543	1034	1555709.2	0.966102
0.97	359.269	55.5	19939.44	1498.3397	1025.5	1536547.3	0.974379
0.98	356.125	37	13176.64	1492.1762	1017	1517543.2	0.982784
0.99	353.036	18.5	6531.168	1486.0631	1008.5	1498694.7	0.991322
1	350	0	0	1480	1000	1480000	1

The equations of physical properties of rock for the calculation are given ;

$$\rho_b = \rho_m - \phi(\rho) (\rho_m - \rho_f)$$

and

$$V_b = V_f \cdot V_m / (\phi(v) (V_m - V_f) + V_f)$$

Where ρ_f and V_f are density and velocity for fluid that can be water and gas. (Sheriff, 1983, p-4)

Appendix B

Processing using Promax version 7.2

B.1. Pre-processing

The first thing we have to do is to set the field parameters as called geometry input. This is a flow that contains 2D Land Geometry Spreadsheets. The flow should be executed in the first time when we start to do processing in promax system for a line. It is used to enter all required field geometry such as receivers, sources, and possibly pattern of spreadsheet. Then we do binning the data and finalising the database. After we execute this step, as one flow, the menu bar will appear. Then we start doing to enter the field geometry from that menu. As remaining for us, we should do extra careful and critically think when setting and entering field parameters in this step. One mistake we have done will make some difficulties in the next step.

Firstly, we fill the set-up button in the menu bar. The most importance in this part is to decide the method of Assign Midpoint within the geometry Set-up window. In this case we usually choose matching pattern numbers using first live channel and station. This means that we should complete the Receivers, Sources, and Patterns spreadsheets then the traces spreadsheet will be filled in automatically. In the next rows we should specify the receiver and source station spacing, the first and last live channel station, the bearing of line, the type of source, and the unit of length in m or km.

Secondly, we fill the Receivers spreadsheet button. This means to open the SRF Ordered Parameter File (OPF) spreadsheet for entering, importing, or editing receiver information. There is very good idea to define every meter in the line then to decide the position (x,y,z) of every receiver station that is required for Assign midpoints step in Bin.

Next button is Sources spreadsheet, which opens the SIN OPF spreadsheet for entering, importing, or editing (shot) information. The number of column is depending

on the assign midpoint method that we have chosen in the set-up. For commonly method we used, matching pattern number using first live channel and station, is has 19 columns. It includes Mark Block, Source, Station, X, Y, Z, FFID, Offset, Skid, Uphole, Hole Depth, Calc Fold*, Static, Pattern, Pat Num Chn, 1st Live Sta, 1st live Sta, Gap Chn, and Gap Size. Here, it does not wont to be described all of the columns in this button. However, we should careful and remember that the Station number here is the position where we defined in Receiver button, and this is should be matched with the Source number.

Finally, we should fill Patterns spreadsheets button. This button appears due to matching pattern method. The main purpose is to match the live channel to the station position number in the line. It has relation with the Gap Chn and Gap Size in the Source spreadsheet. There are two manners to define these: Static and Dynamic Gap definition. The static gap definition uses a gap by the jump in the receiver station numbers within the pattern definition entered the pattern spreadsheet. Whereas, the dynamic gap definition needs adding a gap to an existing pattern definition via the Gap Chn and Gap Size columns in the source spreadsheet. After we complete this spreadsheet, the bin button should be touched for execution of all spreadsheet as binning and finalising for the process of geometry input.

The next step after geometry input is Inline Geometry Header Load. This is a step to fix the identification information and tabulation of parameter that precede data as on magnetic tape (Sheriff, 1983). This automatically loads geometry information from the database to trace header in a processing flow. Trace headers include identification information and tabulation of trace parameters, which precede data as on magnetic tape. It uses two or three trace headers to match the corresponding database parameters. Then the geometry information is loaded in to the trace headers. It builds up the database with the geometry spreadsheet. This process is assumed that the survey data contain sufficient information to uniquely tie the database to trace headers of field data. It then would be the standard matching in channel numbers between the trace headers and geometry database. The significant parameter is primary and secondary headers to match the database. In this case FFID (Field File ID number) is chosen. It means to match

FFID from the trace header to the FFID of database.

Then we need First Break Picking for certain purpose obviously the refraction static correction. It is the first recorded signal attributable to energy generated by the seismic source. First Break Picking automatically picks first breaks for refraction static analysis. On the other hand, first breaks on reflection records are normally used for determining a near surface static model. There are nine difference picks are made, and then the median of these nine picks is used as the first break time. By using Database/header Transfer, it loads the picks from the database into the trace headers.

The last pre-processing is editing. This includes trace editing that follows format verification. The data are rearranged. Field data are usually time-sequential that the first sample for each channel is recorded before the second channel for any channel. Editing may involve detecting dead or exceptionally noisy traces. Bad data may be zeroed out or replaced with interpolated value. Anomalously high amplitudes, which are probably noise, may be reduced to zero or to the level of the surrounding data.

Trace Kill/reverse is commonly used for killing or reversing traces. The list of traces for editing may be retrieved from database entries created by earlier interactive screen selection or we can use the editor to specify trace number for editing. These step cans also be use to find traces in which all samples are zero as dead or reverse polarity. Scanning for dead traces immediately upon input are strong suggested.

B.2. Refraction Static Correction

This correction is commonly used to correct the elevation and weathering variation. Refraction data have to be corrected for elevation and weathering variation, as with reflection data. Refraction static correction is applied to seismic data to compensate for the effect of that variation in the elevation, weathering thickness, weathering velocity, or referent to datum. Refraction static correction, which is base on first-break refraction arrival time, provides a means of dealing with such long-wavelength

variation. The objective is to determine the refraction arrival times, which would have been observed, if all measurements had been made on a (usually) flat plane with no weathering or low-velocity material present.

In practice, this correction consists only one step i.e. refraction static. It provides for first break pick editing on linearly moved out shot records, layer selection and editing, refractor velocity analysis and editing, refractor delay-time computation and editing, quality control displays of various solution, depth model display and editing, shot and receiver static display, and database output. There are three static solution methods can be used in processing; Diminishing residual matrices (DRM), Generalised reciprocal method (GRM), and Delay time. Commonly, in this process we did manually to pick the first break on the screen using interactive display by mouse, then to continue processing till saving the output to the database. Then, in the next flow where we apply the refraction static correction, we choose the Delay Time Method.

B.3. Filtering

Here will be described the only filters which being applied in this research as the relevance tool.

B.2.1.Trace Muting.

Trace muting performs several different types of muting depend on input data. Generally, there are three types of muting; top muting, surgical muting, and bottom muting. In the simple way, top muting is for removing the signal of first arrival and before, surgical muting for airwave, and bottom muting for ground roll. In this application, it only uses top muting for zeroing the first arrival. First breaks and the refraction wavefronts that follow then are usually so strong that they have to be excluded from the stack to avoid degrading the quality of shallow reflection (Sheriff and Geldart, 1995). This energy should be removed by muting, which involves arbitrary

assigning value of zero to trace during the mute interval. Most of muting process in this study is done by picking in Trace Display, interactive process.

B.2.2. Band-pass filtering

Bandpass filter applies a frequency filter(s) to each input trace. Theoretically, the filter algorithm operates in the frequency domain. We may specify one or more sets of bandpass filter frequencies, and set of notch of filter parameters. Filters are four-frequency Ormsby or Butterworth, and may be zero phase or minimum phase. Normally we use minimum phase.

B.2.3. Air Blast Attenuation

This automatically seeks out anomalous energy on a trace-by-trace basis by given a pilot velocity, a relative noise amplitude threshold, and approximate energy envelope width. Theoretically, strong energy from sources such as shot-generated air blast may contaminate seismic data with very high amplitude and broad band noise. Air blast attenuation attempts to face these problems by providing us with a relatively automated air blast attenuation tool.

B.2.4. f-k filtering

Before using this f-k filtering, it is strongly recommended to apply f-k analysis in order to design the filter window. Then, the actual picking of the polygon on screen is done within f-k analysis. It is also recommended that gain corrections be applied before applying this filtering. If the data have highly variable amplitude, the highest amplitudes will dominate the f-k spectrum. Filtering will cause artifacts of these high amplitudes to be spread to other regions of the data in T-X. The f-k filter is a frequency-wave number filter applied to the data in the frequency-wave number domain.

B.4. NMO Correction

Dynamic correction is Normal Move Out (NMO). It is to replace the position of geophone at the shot point because of offset. This is applied from a space variant velocity field. The velocity parameters are have been decided from velocity analysis. A sample by sample velocity is built at each of the locations where we have defined time-velocity pair. For any intermediate location, the velocity is linearly interpolated on a sample-by-sample basis from the location on either side. For any point before the first velocity location, or beyond the last location, the first or last time velocity function is used.

The objective of velocity analysis is to provide comprehensive interactive velocity analysis, velocity quality control, and velocity field modification capabilities. Interactive screen will produce the velocity of significant reflection. Then this felicity will be used as velocity parameter for NMO and CDP stack. There are several methods can be used to analysis the velocity, velocity analysis. For example, picking velocity analysis method, velocity panel method, conventional velocity analysis method, etc.

The common method that has been used is the picking analysis method to the velocity. This involves a considerable number of calculations and hence is fairly expensive to execute; therefore, too few analyses are often run. Where only a limited number of velocity analyses is to be run, their location should be selected judiciously, based on the best available geologic information. So that analysis is not wasted in noisy area and so that change in geology is adequately sample.

The picking analysis method uses mouse to pick the appropriate semblance velocity. Picking involves selecting the time-velocity value to be used in subsequent processing. We often have in mind only achieving a good stack, and stacking can often tolerate appreciable velocity error. Velocity interpretation is time-consuming and hence expensive and has significant potential for error, special case when we know little about the local geology and hence do not factor this into the interpretation.

B.5. Automatic Gain Control (AGC)

AGC automatically varies the gain applied to trace samples as a function of sample amplitude within an AGC time window (Promax manual, 1997). The AGC operator length defines the length of the AGC window used for gain computation. The AGC program moves the window down the trace sample-by-sample and calculates the scale factor at each location. The scale factor is equal to the inverse of the mean, median, or RMS amplitude in the window. The scalar is applied to the sample at the beginning, centre, or end of the AGC window.

At the start and end of the trace, where there less data in the window than the operator length requested, the window will be made as long as possible. Therefore, the window will grow at the start of the trace until it reaches the full operator length, and will remain constant until it reaches the end of the data, where it will shrink to a progressively smaller value (Promax manual, 1997).

B.6. True Amplitude Recovery

This applies a single time-variant gain function to traces to compensate for loss of amplitude due to wavefront spreading and inelastic attenuation. The process offers a number of true amplitude recovery schemes which can be used separately, or in combination. The data may be corrected for amplitude loss. Options include correction for spherical divergence and inelastic attenuation, a dB/sec curve or power-of- time curve. At least one of these options must be selected if we use this process in our flow.

B.7. Stacking

This is a vertically stack to increase the signal to noise ratio. Stacking velocities were obtained at key points along the profiles using the semblance spectrum method when doing velocity analysis. The determined velocities varied due to the structure

along the profiles, and the variation in physical properties of the layer. These velocities were also used in NMO correction of the data.

The final stack is normally output of the final processing datum. In some cases a stack can be output at the floating datum. There is an option for turning off the application of the final static shift. The stack process expects to find the elevation static, including final datum static in the trace header (Promax manual, 1997).

Appendix C

T-X Seismic Curve Model

Layer	Thicknes s	Vp	dt	Vrms	X-Interval	Vr-GR	V -AW
1	3	400	15		1	170	350
2	28	840	60	820.674			
3	89	2050	90	1752.301			

Station	x(m)	t1(ms)	t2	t3	t4	t5(1&2lyr)	t6
1	0	0	0	0	-15		-75.5477
2	1	-2.5	-5.88235	-2.85714	-15.2069		-75.5575
3	2	-5	-11.7647	-5.71429	-15.8114		-75.587
4	3	-7.5	-17.6471	-8.57143	-16.7705		-75.636
5	4	-10	-23.5294	-11.4286	-18.0278		-75.7047
6	5	-12.5	-29.4118	-14.2857	-19.5256		-75.7929
7	6	-15	-35.2941	-17.1429	-21.2132		-75.9006
8	7	-17.5	-41.1765	-20	-23.0489		-76.0276
9	8	-20	-47.0588	-22.8571	-25		-76.174
10	9	-22.5	-52.9412	-25.7143	-27.0416		-76.3395
11	10	-25	-58.8235	-28.5714	-29.1548	-25.0949	-76.524
12	11		-64.7059	-31.4286	-31.3249	-26.2854	-76.7275
13	12		-70.5882	-34.2857	-33.541	-27.4758	-76.9497
14	13		-76.4706	-37.1429	-35.7946	-28.6663	-77.1905
15	14		-82.3529	-40	-38.0789	-29.8568	-77.4497
16	15		-88.2353	-42.8571	-40.3887	-31.0473	-77.7272
17	16		-94.1176	-45.7143	-42.72	-32.2378	-78.0227
18	17		-100	-48.5714	-45.0694	-33.4282	-78.3361
19	18		-105.882	-51.4286	-47.4342	-34.6187	-78.6671
20	19		-111.765	-54.2857	-49.8121	-35.8092	-79.0155
21	20		-117.647	-57.1429	-52.2015	-36.9997	-79.3811
22	21		-123.529	-60	-54.6008	-38.1901	-79.7636
23	22		-129.412	-62.8571	-57.0088	-39.3806	-80.1628
24	23		-135.294	-65.7143	-59.4243	-40.5711	-80.5785
25	24		-141.176	-68.5714	-61.8466	-41.7616	-81.0103
26	25		-147.059	-71.4286	-64.2748	-42.952	-81.4581
27	26		-152.941	-74.2857	-66.7083	-44.1425	-81.9216
28	27		-158.824	-77.1429	-69.1466	-45.333	-82.4005
29	28		-164.706	-80	-71.5891	-46.5235	-82.8946
30	29		-170.588	-82.8571	-74.0355	-47.7139	-83.4035
31	30		-176.471	-85.7143	-76.4853	-48.9044	-83.927
32	31		-182.353	-88.5714	-78.9383	-50.0949	-84.4648
33	32		-188.235	-91.4286	-81.3941	-51.2854	-85.0168
34	33		-194.118	-94.2857	-83.8525	-52.4758	-85.5825
35	34		-200	-97.1429	-86.3134	-53.6663	-86.1617
36	35		-205.882	-100	-88.7764	-54.8568	-86.7542
37	36		-211.765	-102.857	-91.2414	-56.0473	-87.3596
38	37		-217.647	-105.714	-93.7083	-57.2378	-87.9778
39	38		-223.529	-108.571	-96.1769	-58.4282	-88.6084
40	39		-229.412	-111.429	-98.6471	-59.6187	-89.2512
41	40		-235.294	-114.286	-101.119	-60.8092	-89.9059

42	41	-241.176	-117.143	-103.592	-61.9997	-90.5723
43	42	-247.059	-120	-106.066	-63.1901	-91.2501
44	43	-252.941	-122.857	-108.541	-64.3806	-91.939
45	44	-258.824	-125.714	-111.018	-65.5711	-92.6389
46	45	-264.706	-128.571	-113.496	-66.7616	-93.3494
47	46	-270.588	-131.429	-115.974	-67.952	-94.0703
48	47	-276.471	-134.286	-118.454	-69.1425	-94.8014

t1 : Direct wave $t1 = x / V0$	t4 : first boundary reflector $t4 = (x^2 + 4 \cdot h1^2)^{0.5} / V0$	t5 : refractor (2 layer model) $t5 = 2 \cdot h1 \cdot ((1/V0)^2 - (1/V1)^2)^{0.5} + x/V1$
t2 : Ground roll, Vr = 250 m/s $t2 = x / Vgr(\text{Gound roll Velocity})$	t6 : second boundary reflector (water table) $t6 = ((x / Vrms)^2 + (2 \cdot (h1+h2)/Vrms)^2)^{0.5}$	
t3 : Air Wave, V = 330 m/s	t3 = x / Vaw (Air wave Velocity)	

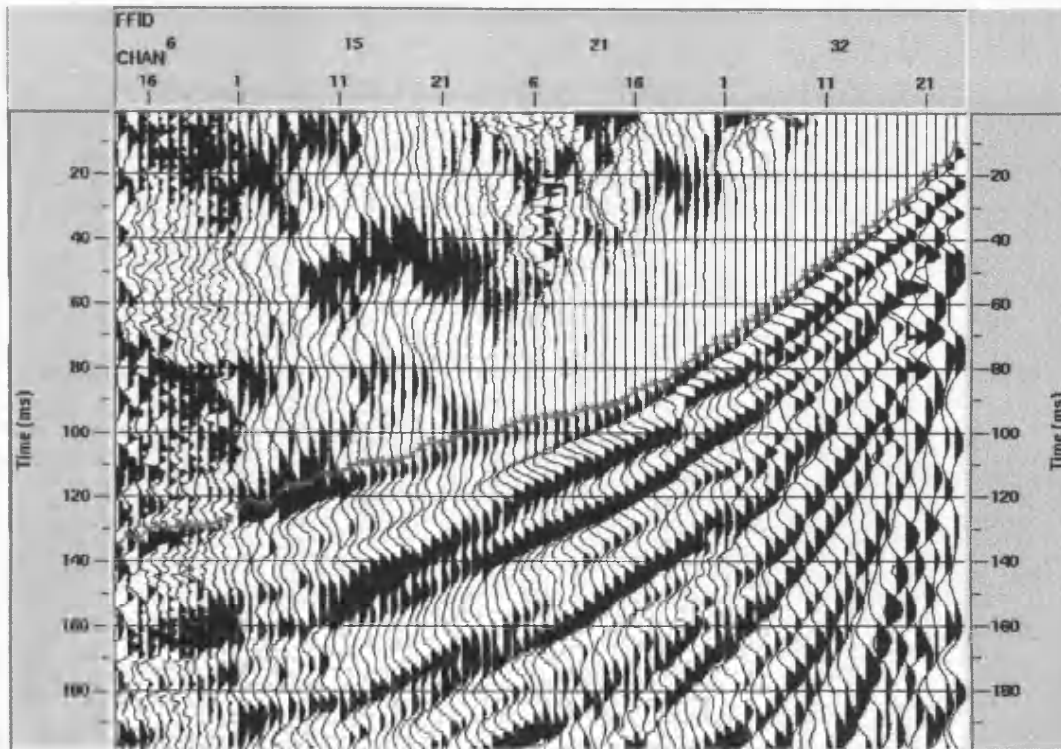
Appendix D

Normalisation of Croft Amplitude Record

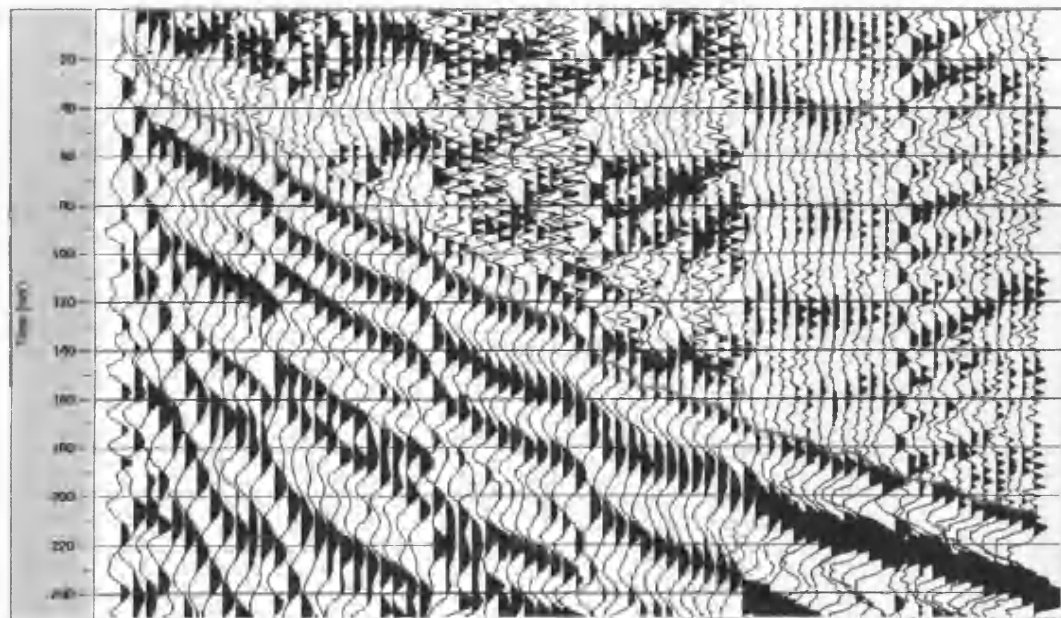
Filtered Normalised Amplituds					Normalised Amplituds					Raw Energy				Raw Amplitude							
SR2	SR3	SR48	SR49	Average	offset	Sr2	Sr3	Sr48	Sr49	average	StDev	SR2	SR3	SR48	SR49	SR2	SR3	SR48	SR49	StDev	
		6.139751	5.259369	5.69956	66			6.139751	5.259369	5.69956	0.622524			72.0801	193.21			8.49	13.9	3.825448	
		5.300868	5.032346	5.166607	68			5.568443	5.145857	5.35715	0.298813			59.29	184.96			7.7	13.6	4.17193	
6.96	5.985877	4.22816	4.717036	5.472768	70	6.96	5.985877	4.194412	4.691811	5.458025	1.254108	48.4416	46.7856	33.64	153.76	6.96	6.84	5.8	12.4	2.979217	
5.71	4.743195	3.432668	4.363889	4.562438	72	5.66	4.900718	2.921625	4.313439	4.448945	1.157846	32.0356	31.36	16.3216	129.96	5.66	5.6	4.04	11.4	3.23804	
4.313333	3.649285	3.432668	3.998129	3.848354	74	4.51	3.34299	3.181967	4.086416	3.780343	0.625942	20.3401	14.5924	19.36	116.64	4.51	3.82	4.4	10.8	3.292278	
3.056667	2.955016	3.977459	3.821556	3.452674	76	2.77	2.704146	4.194412	3.594533	3.315773	0.712196	7.6729	9.5481	33.64	90.25	2.77	3.09	5.8	9.5	3.118472	
2.32	2.721649	4.363152	3.745881	3.287671	78	1.89	2.817913	4.555999	3.783719	3.261908	1.158479	3.5721	10.3684	39.69	100	1.89	3.22	6.3	10	3.607098	
1.916667	2.84125	4.322172	3.594533	3.168655	80	2.3	2.642887	4.339047	3.859393	3.285332	0.970133	5.29	9.1204	36	104.04	2.3	3.02	6	10.2	3.590395	
2.056667	2.473696	3.893089	3.36751	2.947774	82	1.56	3.062949	4.071472	3.140486	2.958727	1.038996	2.4336	12.25	31.6969	68.89	1.56	3.5	5.63	8.3	2.893411	
1.8	2.584545	3.389277	3.039587	2.703352	84	2.31	1.715251	3.268748	3.102649	2.599162	0.722671	5.3361	3.8416	20.4304	67.24	2.31	1.96	4.52	8.2	2.868395	
1.956667	2.575794	2.924035	2.900851	2.589337	86	1.53	2.975436	2.827612	2.875626	2.552168	0.684222	2.3409	11.56	15.2881	57.76	1.53	3.4	3.91	7.6	2.541692	
2.106667	3.132959	2.482899	2.73689	2.614854	88	2.03	3.036695	2.675745	2.724277	2.616679	0.42256	4.1209	12.0409	13.69	51.84	2.03	3.47	3.7	7.2	2.194827	
2.503333	3.270062	2.420224	2.535091	2.682178	90	2.76	3.386746	1.945339	2.610766	2.675713	0.591703	7.6176	14.9769	7.2361	47.61	2.76	3.87	2.69	6.9	1.972182	
2.72	3.386746	2.400939	2.514912	2.755649	92	2.72	3.386746	2.639587	2.270231	2.754141	0.46499	7.3984	14.9769	13.3225	36	2.72	3.87	3.65	6	1.38605	
		2.885466	2.295456	2.590461	94			2.617891	2.663738	2.640815	0.032418			13.1044	49.5616			3.62	7.04	2.418305	
		3.338655	2.680134	3.009395	96			3.39892	1.952399	2.675659	1.022845			22.09	26.6256			4.7	5.16	0.325269	
		3.627925	2.951301	3.289613	98			3.999155	3.424265	3.71171	0.406508			30.5809	81.9025			5.53	9.05	2.489016	
		3.668905	3.407869	3.538387	100			3.485701	3.477237	3.481469	0.005984			23.2324	84.4561			4.82	9.19	3.090057	
		3.382046	3.312015	3.34703	102			3.521859	3.322105	3.421982	0.141248			23.7169	77.0884			4.87	8.78	2.764788	
		3.659263	3.298141	3.478702	104			3.138577	3.136703	3.13764	0.001325			18.8356	68.7241			4.34	8.29	2.793072	
		4.459576	3.298141	3.878858	106			4.317351	3.435616	3.876484	0.623481			35.6409	82.4464			5.97	9.08	2.199102	
		5.262299	3.477237	4.369768	108			5.922798	3.322105	4.622452	1.838968			67.0761	77.0884			8.19	8.78	0.417193	
		5.780574	3.885879	4.833227	110			5.546748	3.673991	4.610369	1.324239			58.8289	94.2841			7.67	9.71	1.442498	
		5.795038	4.862078	5.328558	112			5.872176	4.661541	5.266859	0.856048			65.9344	151.782			8.12	12.32	2.969848	
		5.758879	6.012329	5.885604	114			5.966189	6.250703	6.108446	0.201182			68.0625	272.91			8.25	16.52	5.847773	
		5.134538	6.14602	5.640279	116			5.438272	7.124742	6.281507	1.192515			56.5504	354.569			7.52	18.83	7.997378	
		4.208875	5.710893	4.959884	118			3.999155	5.062616	4.530885	0.75198			30.5809	179.024			5.53	13.38	5.550788	
		2.938499	4.677937	3.808218	120			3.189199	4.94532	4.06726	1.241765			19.4481	170.825			4.41	13.07	6.123545	
		1.851327	3.950202	2.900764	122			1.627142	4.025877	2.82651	1.696161			5.0625	113.21			2.25	10.64	5.932626	
		0.896736	3.02193	1.959333	124			0.737638	2.87941	1.808524	1.514461			1.0404	57.9121			1.02	7.61	4.659834	
		0.325428	2.256358	1.290893	126			0.325428	2.160503	1.242966	1.297594	%reduced		0.2025	32.6041			0.45	5.71	3.719382	
		1.582856			128			1.729159			0.802384	75.7138			20.8849				4.57		
		0.858904			130			0.858904			3.303868				5.1529				2.27		
											Average-	12.21665	15.95177	23.35963	85.3325	34.215			Average	3.303868	
											Afactor-	1	1.14269	1.382792	2.6429						

Appendix E

Picked Raw data for refraction P and S-wave



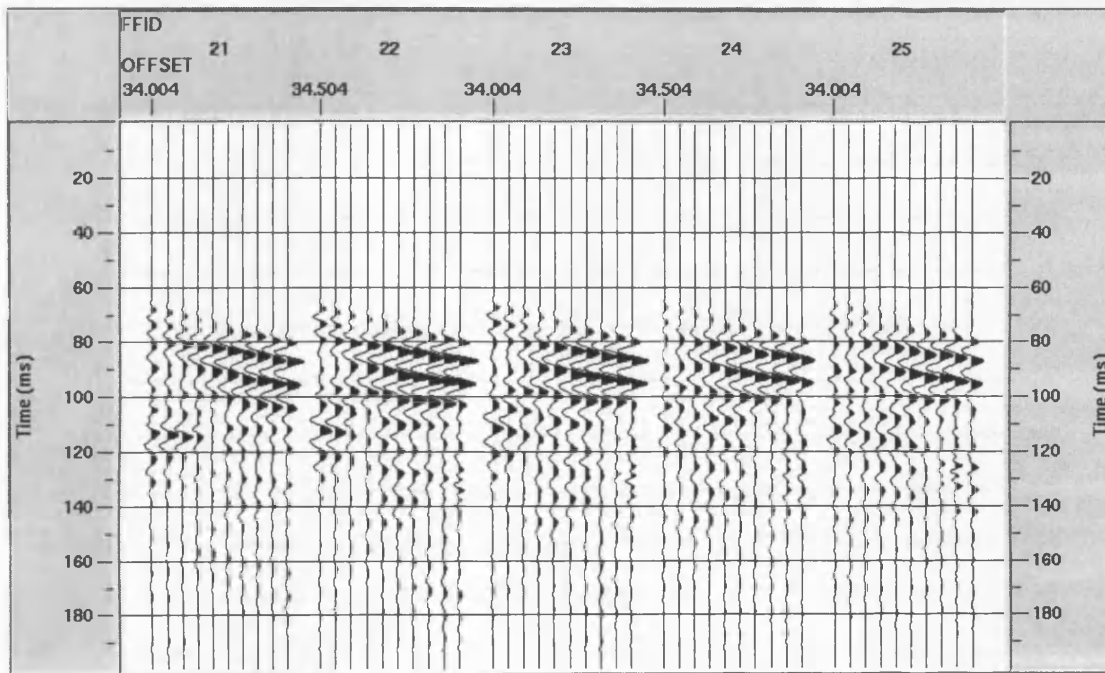
a). P-wave refraction record



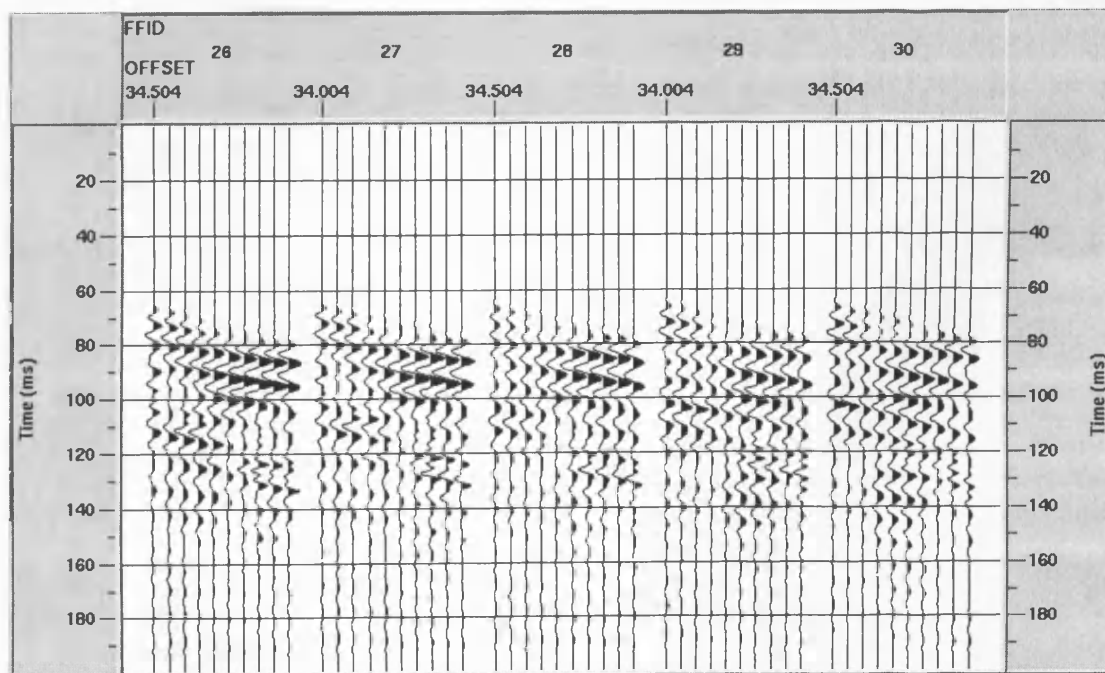
b). S-wave refraction record

Appendix F

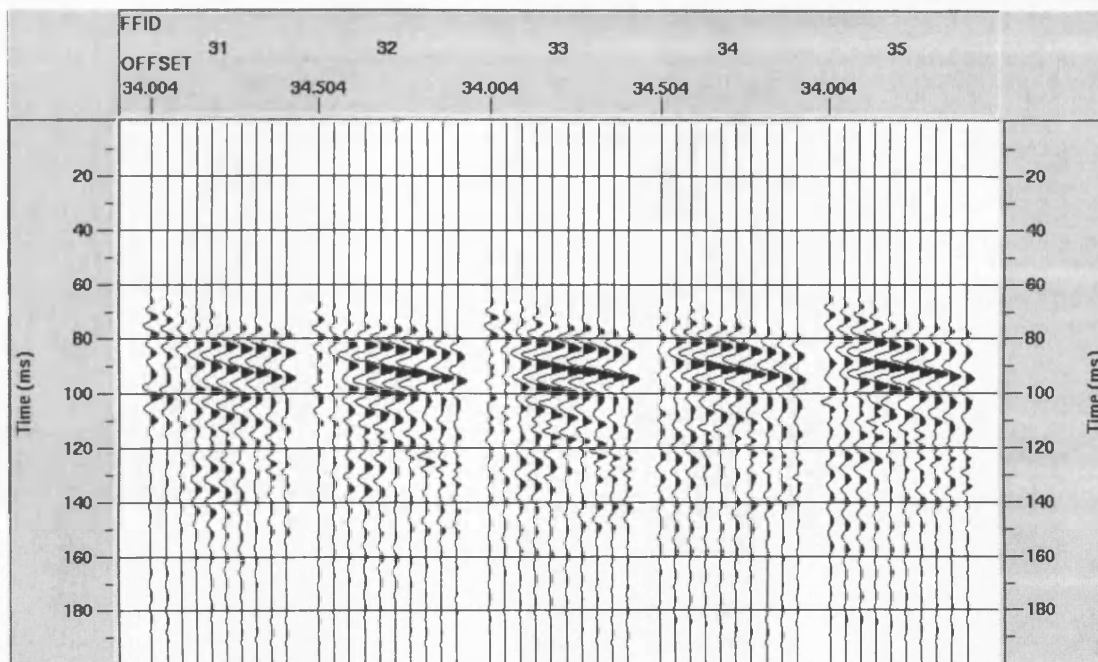
Raw data before stacking for AVO analysis for 15 shot record with offset between 40 to 48 m; (a) FFID 21-25, (b) FFID 26-30, (c) FFID 31-35



(a) FFID 21-25



(b) FFID 26-30



(c) FFID 31-35

Appendix G

The Calculation of Poisson's Ratio (σ) Vs (α/β) as Ratio of P-wave velocity to S-wave Velocity using formulae:

$$\alpha / \beta = \sqrt{\frac{1 - \sigma}{\frac{1}{2} - \sigma}}$$

α/β	σ
1.42	0.008068
1.47	0.0693
1.6	0.179487
1.8	0.276786
2	0.333333
2.2	0.369792
2.4	0.394958
2.6	0.413194
2.8	0.426901
3	0.4375
3.2	0.445887
3.4	0.452652
3.6	0.458194
3.8	0.462798
4	0.466667
4.2	0.469952
4.4	0.472767
4.6	0.475198
4.8	0.477314
5	0.479167