2-D SEISMIC REFLECTION DATA INTERPRETATION OF LINE 985-QPR-04

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CERTIFICATE OF APPROVAL

This dissertation by **Iftikhar Ahmad** is accepted in its present form by the Department of Earth Sciences, Quaid-i-Azam University Islamabad as satisfying the requirement for the award of degree of **M.Sc Geophysics.**

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In the name of Allah,
The most Beneficent,
The most Merciful.

I found Whose Favours always with me,
Only the mistakes have been mine.

Dedications

My whole thesis work is dedicated to my PARENTS ,Brothers, Cousins and especially to my sister

Abstract

A seismic reflection line 985-QPR-04 of Qadirpur area (Sindh Province) was provided by the department of Earth Sciences, Quaid-e-Azam University for interpretation. The given line is oriented in SE direction. The section comprised of a total of shot points from 200-926, from which SP-200 SP-450 were used in the interpretation. The section is a final stacked migrated section. The root mean square velocity, Dix interval velocity and Dix average velocity at different time were given. Thirteen were present in the allotted shot points, i.e. SP-200, SP-225, SP-250, SP-275, SP-300, SP-325 SP-350 SP-375, SP-400, SP-425, SP-450.

Velocity versus time graph was prepared. For seismic section, arrival times (two way) of each marked reflector was determined. Using these arrival times and velocity information, the depth of each reflector was calculated using the formula $s= (v*t)/2$. finally a depth section was constructed. This depth section provides a reliable picture of reflectors present in the subsurface.

In total seven prominent reflectors were marked and then interpreted. The three normal faults were marked. Due to extensional regime normal faults, Horst and Graben structure are found in the depth section.

Acknowledgement

All praise to Almighty Allah, the most merciful and the creator of the universe who gave me the opportunity to come to this stage in my life. I am nothing He blessed me with knowledge and enables me to complete my work. Acknowledgement. All respects to Holy Prophet Muhammad (P.B.U.H), who is the last messenger, whose life is a perfect model for the whole humanity.

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1.1 Introduction to the Seismic Line:

A seismic survey was carried out in **QADIR PUR** area by **O.G.D.C.L** in 1998.The party number is SP-5. The data acquisition and processing were made by selecting appropriate field and processing parameters. This dissertation pertains to the interpretation of 60 fold stacked and migrated time structure map of **QADIR PUR** area. The line number is **985-QPR-04.** Although one lines is not sufficient to map the structure, but i made an attempt to map it.

1.2 Survey Parameters:

Source

Energy Source	Dynamite
Charge pattern	1 Hole
Average shot depth	21 meters
Average charge size	3 Kilograms
S.P.interval	50 meters

Instruments

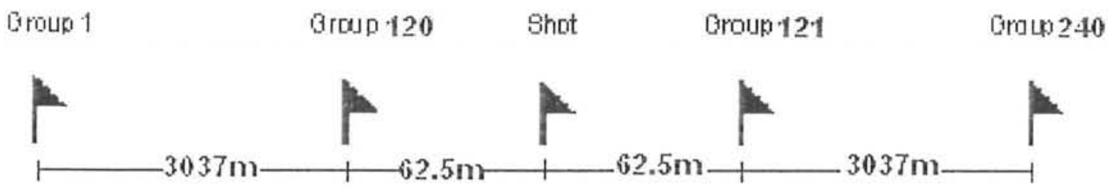
System	I/0 System one
Format	DMUX SEG D
Notch Filter	Out
Aliasing Filter	In
Field sampling interval	2 ms
Record length	5 seconds
No. of data traces	240

Cable

Spread	3037-62.5-x-62.5-3037
Group interval	25 m

Type of Geophones	L-10A
Geophones/Group Base	24.30m
Geophone Code	0312 Linear

SPREAD DIAGRAM



1.3 Processing of the Seismic Line

This line was processed by OGDCL in 1999. The processing sequence of the seismic line is given in the following table.

1. **Demultiplex**
2. **Preprocessor**

    Geometry definition

    Bad trace edit/ Reverse

3. **Geometrical Spreading Compensation**
4. **Exponential Gain**
5. **FK Dipfilter**
6. **Band Pass Filter**

Low cut	High cut
(HZ)	(HZ)
6-8	65-70

7. **Surface Consistent Deconvolution**

Operator Length	240msec
Prediction Distance	8msec

No of Autocorrelation Windows 2

Prewhitening 0.1%

Trace Balance

8. CMP sort

9. Velocity Analysis (CVA)

10. Normal Move out (NMO) Correction

11. Residual Statics Correction

Max. Shift 20msec

12. Velocity Analysis (CVA)

13. Stack

14. Finite Difference Migration

15. Multichannel Coherency Filter (MCCF)

16. Time Variant Filter

Time Zone	Low Cut	Slope	High Cut	Slope
Msec	HZ	db/oct	HZ	Db/oct
1800	10	16	60	36
2000	8	16	40	36

18. Random Noise Attenuation

19. RMS Gain

Amplitude	2000
Starting Gate	256
End Gate	512

1.4 Display parameters of the seismic line

Display parameters are the information related to the seismic section such as scale, time per inch/sec on it etc. Display parameters of the line are given below.

Scales

1. **Horizontal Scale**	80 Traces = 1km
	40 Traces per inch
2. **Vertical Scale**	10 centimeter per second
3. **Display Amplitude**	10 db
4. **Polarity**	Normal
5. **Datum Velocity**	1700 m/sec
6. **Datum Plane**	Mean Sea Level
7. **Fold**	6000%
8. **Processing sampling interval**	2 msec

1.5 Objective

The main objectives of the dissertation are;

- To determine the times for each CDP at some constant interval on the basis of variation of these velocities in each velocity panel
- To determine the average velocities at certain constant interval of time from each CDP data
- To determine the mean of all average velocities determined for each constant time from different methods
- To prepare the mean line graph of mean average velocity vs. selected constant time
- To analysis the stratigraphic structure using the time section and the Dix Iso-Velocity contour map
- To calculate the depth of each interface using the provided average velocity information
- Attempt to differentiate the lithology in terms of clastic and non-clastic material
- To study the effects of the velocity variation along the profile and its distribution i n the subsurface.

2.1 History of the Earth:

The planet **EARTH** came into being about 4.6 billion years ago, upto the JURASIC age there was only one land mass on the earth which was called **Pangea**, this land mass started breaking about 200 million years ago and was divided into two parts. the northern part was called **Laurasia** while southern part was called **Gondwana Land**. This breakage was initiated by two rifts, one in the northern part, between North America and Africa. It gave birth to the **North Atlantic Ocean**. Second rift was in the southern part, between South America and Africa which gave birth to the **South Atlantic Ocean**. Now due the rift that was produced between South America and Africa, a Y shaped crack was produced in the southern part of the Gondwana Land due to which India was separated from the Gondwana Land, it was done about 130 million years ago.(Tarbuck et al, 2000).

Now there was one transform fault, **Chagos Mauritius Fault** and other was along the **Ninty East Ridge**, also a ridge was present south of the **Srilanka**, this ridge was spreading in such a way that it pushed India north ward, India started its north ward journey. Between 130 and 80 million years, it moved at 3 to 5 centimeter/year, then, between 80 and 55, its speed was increased to average speed of 16cm/year, during this time it varied from 15 to 25cm/year. Due to this movement, intra oceanic subduction was occurred, the evidence of this subduction is present in the form island arcs which include **Kohistan, Laddakh, Nuristan and Kandhar Island arcs**. (Kazmi & Jan, 1997).

At this stage, India's movement was slowed down by its collision with Kohistan and Laddakh island arcs, it was done about 55 to 50 million years ago.The most important feature of this collision is the upheaval of the **Great Himalayas**.

The Himalayas are divided in the following sequence.

Laddakh / Tibbet Block

Indus – Tsangpo Suture Zone

Tethyan Himalayas

Main Mantle Thrust

High Himalayas

Main Central Thrust (MCT)

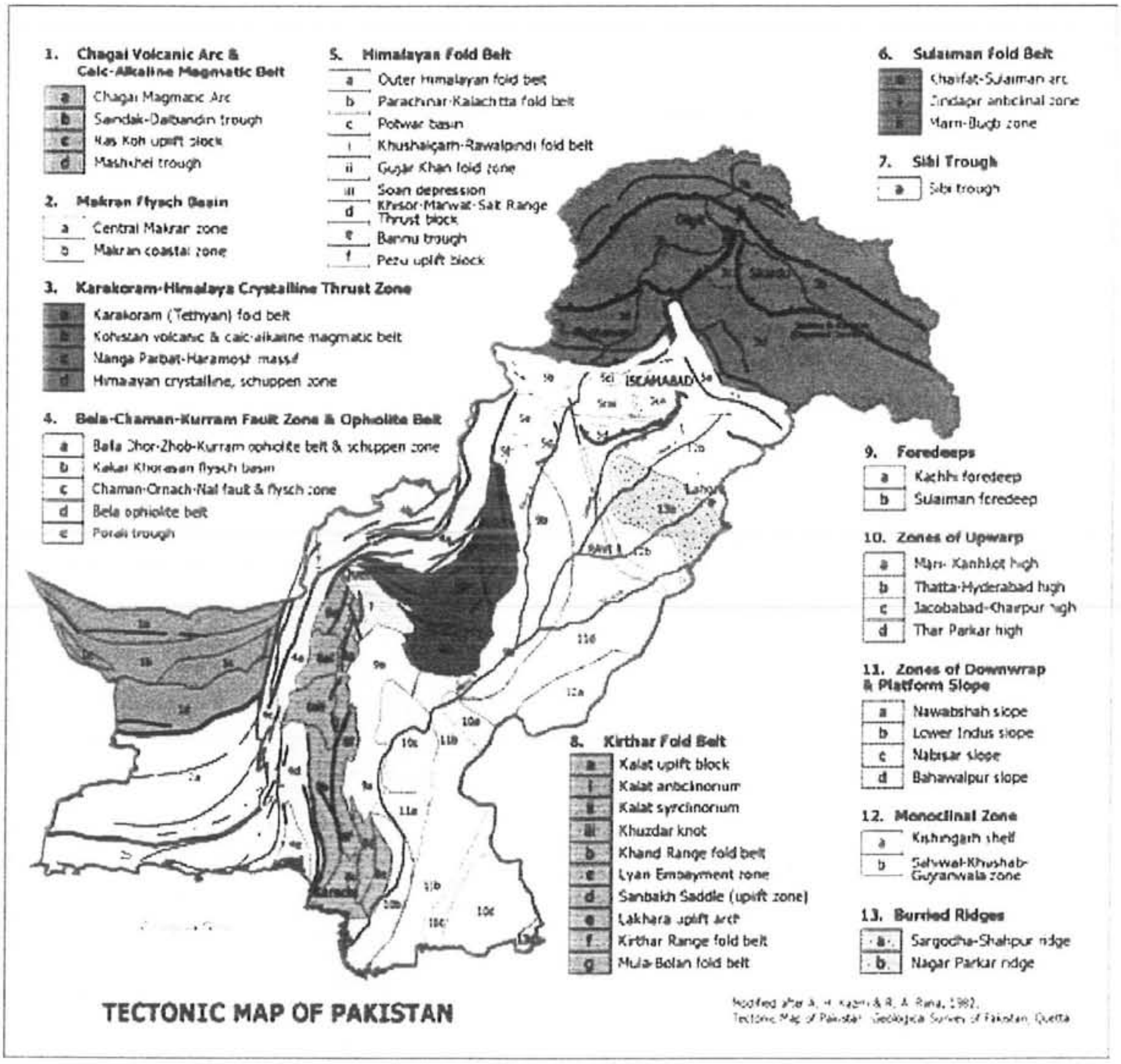
Lesser Himalayas

Main Boundary Thrust (MBT)

Sub Himalayas

Himalayan Frontal Fault (HFF)

Himalayan Fore Deep



Map#1Tectonic Map of Pakistan

2.2 BASINS OF PAKISTAN

Sediments get developed at certain place, they may be deposited at the place of their origin or may get transported to other place by transporting agents, and the sediments deposited at the place of origin are called **Molasses,** what happens with the transported sediments? Its answer is that they are often deposited at a place which is characterized by regional subsidence and where they could get preserved for longer periods of time, such a place is called a **Basin,** in a basin the receptacle or container which is the substratum of the basin is called **Basement**, the accumulated sediments in a basin are called **Sedimentary Cover**, the gradual settling of sediments in a basin is called **Subsidence**, the point of maximum sedimentary accumulation in a basin is called **Depocenter** (Duval.1999)

In terms of the genesis and different geological histories, Pakistan consist upon two major sedimentary basins, the Indus Basin and the Balochistan Basin, these basins were developed during different geological episodes

There is another smaller basin which has its own geological history of development, this basin was developed due to the collision between India and Eurasia and is classified as the Median basin. It is the **Kakar Khorasan** Basin which is also known as **Pishin Basin**.

Indus Basin is divided into the following classes,

1. Upper Indus Basin

   It further has following partition

   (a) Kohat Sub Basin

   (b) Potwar Sub Basin

2. Lower Indus Basin

   It further has following partition

   (a) Central Indus Basin

   (b) Southern Indus basin

2.1 The Upper Indus Basin

Boundaries of The Upper Indus Basin

Directions	Structural Boundaries	Geographic Boundaries
East	MBT	India
West	Kurram Fault	Kurram Cherat Ranges
North	MBT	Margala Hills, Kala Chitta Ranges
South	Sargodha High	Punjab Plain

2.2.2 Lower Indus Basin

Boundaries Of The Lower Indus Basin

Directions	Boundaries
East	Indian Shield
West	Kirthar And Sulaiman Ranges
North	Sargodha High
South	Off Shore

The Lower Indus Basin is further divided into two classes,

(a) Central Indus Basin

(b) Southern Indus Basin

The Central and Southern Indus Basin are separated by Jacobabad and Mari Khandkot highs, these are collectively termed as Sukkur Rift, these highs have been active since Jurasic times and at least up to Paleocene

2.2.2.1 Central Indus Basin

The Centeral Indus Basin having following divisions

(a) Punjab Plateform

(b) Sulaiman Depression

(c) Sulaiman Fold Belt

The Boundaries of Central Indus Basin

Directions	Boundaries
East	Indian Shield
West	Marginal Zone Of The Indian Plate, Sulaiman Range
North	Sargodha High,
South	Jacobabad KhairPur High, Mari Khandkot High

2.2.2.2 SOUTHERN INDUS BASIN

This basin is located just south of the Sukkur Rift which is a divide between the Central and the Southern Indus basin. The oldest rocks encountered in the area are of Triasic age. The Central and the Southern Indus basins were undivided until middle cretaceous when Khair Pur-Jacobabad High became a prominenet feature, this is indicated by homogeneous lithologies of Chiltan Limestone which is of Jurasic age and Sembar Formation which is of lower cretaceous age and lie across the high. This part of the Indus Basin comprises the following four main units.

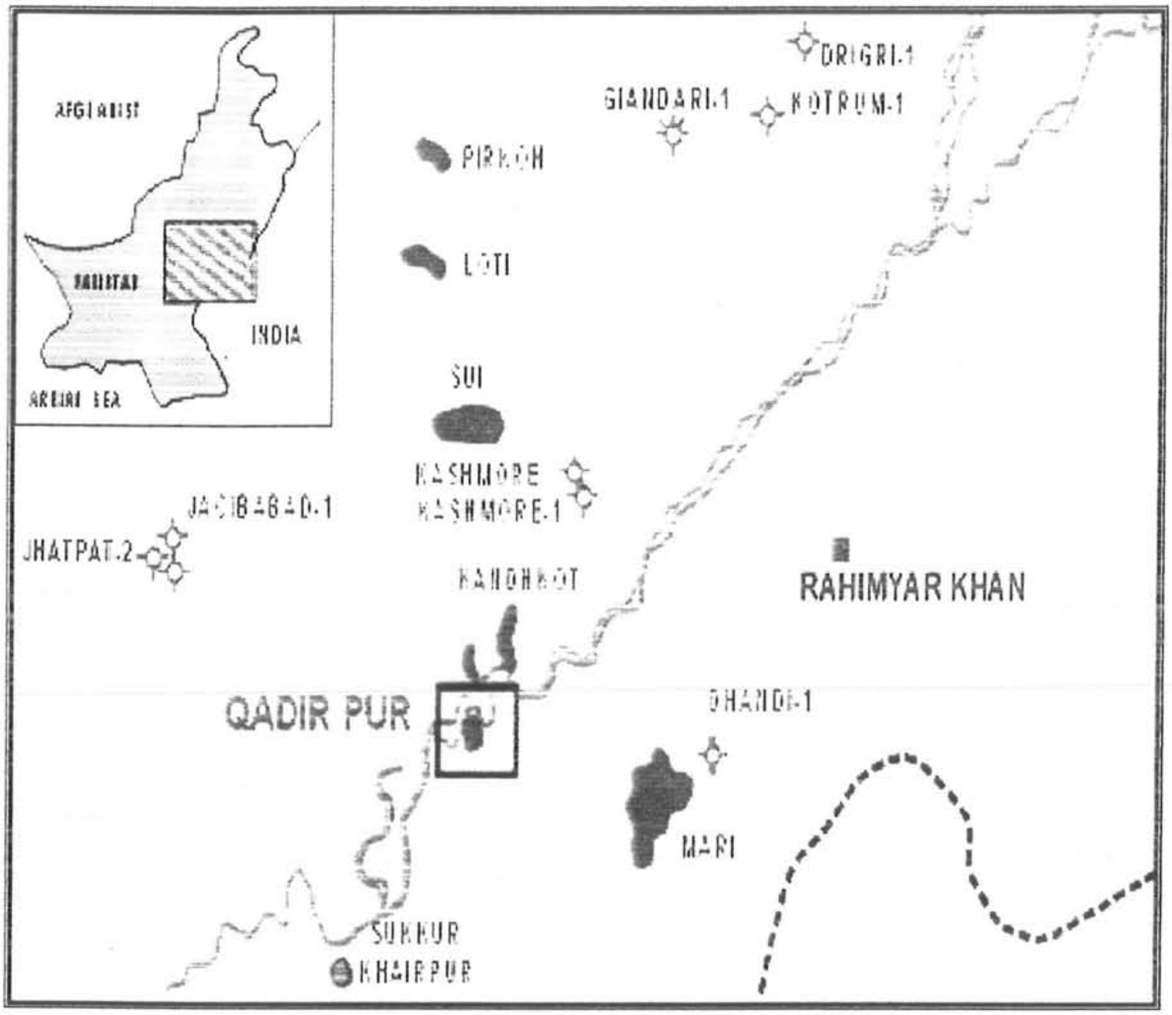
Southern Indus Basin (a) Thar Plateform
(b) Karachi Trough
(c) Kirthar Foredeep
(d) Kirthar Fold belt
(e) Offshore Indus

The Boundaries of the Southern Indus Basin

Directions	Boundaries
East	Indian Shield
West	Marginal Zone of The Indian Plate ,Kirthar Range
North	Jacobabad KhairPur High, Mari Khandkot High
South	Off shore, Murray Ridge- Oven Fracture Plate Boundary

2.3 Introduction to Qadirpur Area:

Qadirpur area administratively lies in Ghotki and Jacobabad of districts Sindh Province. Geologically the Qadirpur is situated with the Mari Kandhkot high, central Indus basin of Pakistan.



LEGENDS

WELL

GAS

GAS SHOW

DRY

FIELD

GAS

Map #2 Location Of Qadirpur

The intended 3D seismic data acquisition and processing programme of Qadirpur joint venture an area of 364 sq km. Previously about 420 lines km of 2D seismic survey has been carried out by OGDCL in 1990 , 1992 and 1998.

In 1990 in Qadirpur, gas was discovered in Early Eocene Sui Main Lst, Sui upper Lst and Middle Eocene Habib Rahi Lst. So for 25 wells have been drilled for development of field. Most of the wells in Qadirpur Gas field area confined to Sui main lst. while Qadirpur-1 and QadirpurX-2 were drilled to Pab/Ranikot (Cretaceous /Pleocene) Formations

2.3.1 Structure:

On surface the Qadirpur structure is covered by Alluvium of floodplain area of Indus river.It is a NW-SE trending anticline comparatively broad in its Southern half.

2.3.2 Prospect:

Source Rocks:

Potential source rocks are Sembar Shles, Shles of Mughalkot Formation, Ranikot Formation and Sirki are also considered for their source potential.

Reservoir Rocks:

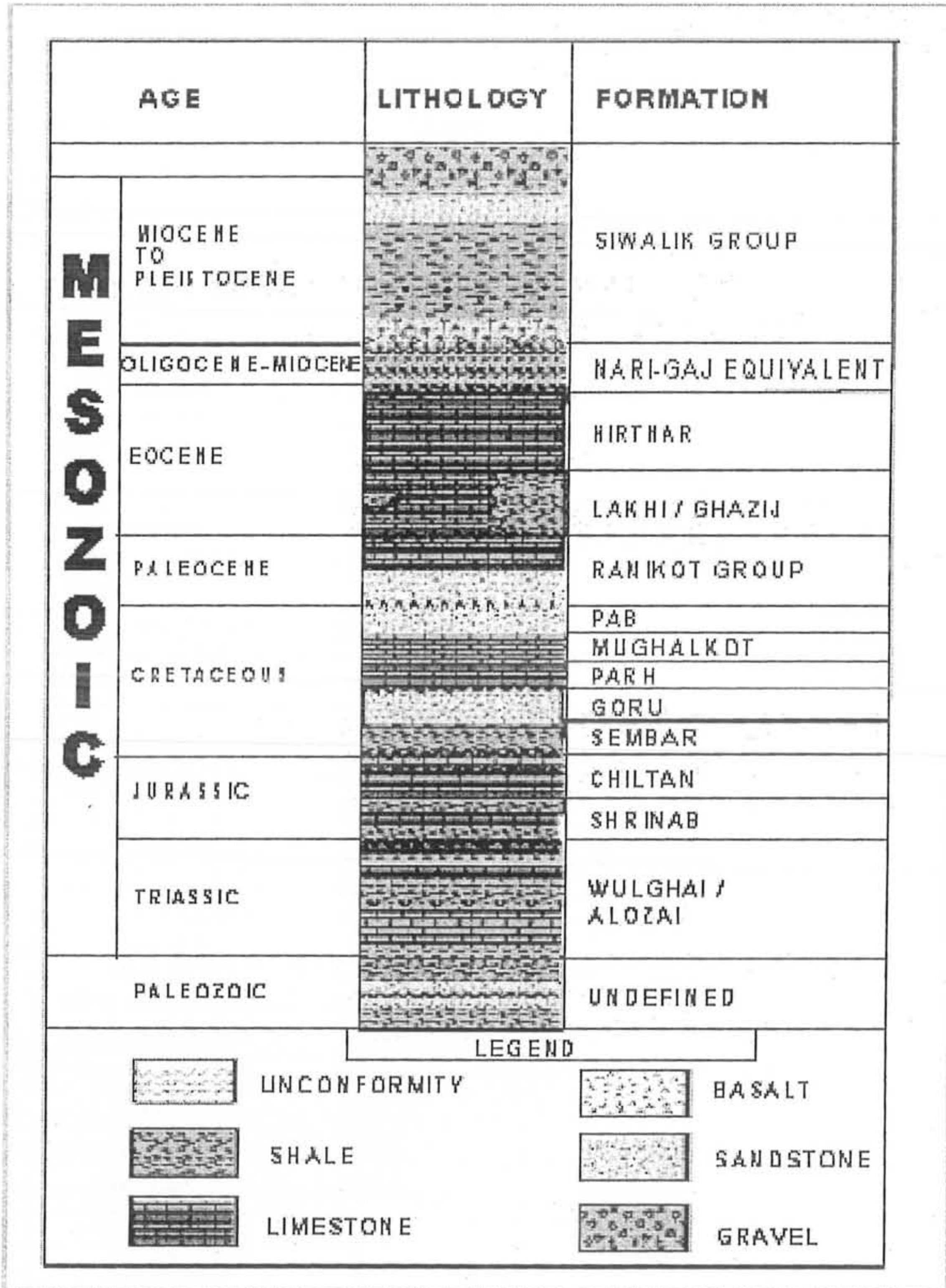
Sui Main Limestone and Sui Upper Limestone are the main producer where as limestone of Habib Rahi is considered as secondary reservoir.

Cap Rock:

The Ghazij Shales act as cap for Sui Main Limestone and Sui Upper Limestone and also Sirki Shales over Habib Rahi Limestone act as a cap rock.

Detailed stratigraphy of the area is given as below:

Stratigraphic Column of Central Indus Basin.



(www.slb.com)

2.4 Stratigraphy of the Area:

I. **Alluvium:**

Age: Recent

Lithology: Sandstone, Siltstone and Clay deposit by the River Indus.

Sandstone: offwhite, light gray, fine to medium grain, subangular to subrounded friable.

Siltstone: Earthy light grey, medium hard.

Clay: Khaki, earthy, soft and stickly.

Contact: Contact with underlying Siwaliks is unconfirmable.

II. **Siwaliks:**

Age: Pliocene

Lithology: Sandstone with interclation of clay.

Sandstone: Offwhite, Yellowishwhite, friable to medium hard, fine to medium grained, subangular to subrounded, moderately sorted, slightly calcareous.

Clay: khaki,soft.

Contact: Contact with underlyingNari Formation is unconformable.

III. **Nari Formation:**

Age: Oligocene

Lithology: Sandstone with streaks of clay.

Sandstone: Off white ,transparent, quartoze, medium to coarse grain, subangular to subrounded, loose, moderately sorted, non-calcarious.

Clay: Brown, soft.

Contacts: contract with underlying Kirther Fornmation is unconfirmable.

IV. **Kirther Formation:**

Kirther Formation is divided into four members:

a) Drazinda Member:

b) Pirkoh Limestone Member

c) Sirki Member

d) Habib Rahi Limestone Member

a) **Drazinda Member:**

Age: Middle Eocene

Lithology: Shale with interclation of marl

Shale: Greenish grey,soft,laminated,moderately indurated.

Marl: greenish grey, off white,soft.

Contact: Upper contact is unconformable with Nari formation and lower contact with Pirkoh Limestone is conformable.

b) **Pirkoh Limestone Member:**

Age: Middle Eocene

Lithology: Limestone with thin marl

Limestone: Dirty white, white, creamy,mediamgrained,hard and fossiliferrous.

Marl: Grey, soft to mediam hard,grading to limestone.

Contact: Upper contact with Darzinda Member and lower contact with Sirki Shale Member is conformable.

c) **Sirki Member:**

Age: Middle Eocene

Lithology: Shale with thin bands of limestone.

Shale: Bluesh green, soft, slightly calcarious.

Limestone: Offwhite, hard, fosilliferous.

Contacts: Upper contact with Pirkoh Limstone and lower contact with Habib Rahi Limestone is confirmable.

d) **Habib Rahi Limestone Member:**

Age: Middle Eocene

Lithology: Limestone with thin bands of Marl.

Limestone: Offwhite, white, creamy, fossilliferous.

Marl: Greenish grey, soft, sticky.

V. **Ghazij Formation:**

Age: Lower Eocene.

Lithology: Shale with thin beds of limestone.

<u>Shale:</u> Greenish grey, pyretic, calcareous with occasional fossils.

<u>Limestone:</u> White, medium hard having thickness 8 to 10m.

Contact: Upper contact with Habib Rahi Limestone and lower contact with Sui Upper Limestone is conformable.

VI. **Sui Main Limestone:**

Age: Lower Eocene.

It has following three units.

a. Sui Upper Limestone.

b. Sui Shale.

c. Sui Main Limestone.

a. **Sui Upper Limestone:**

Age: Lower Eocene.

Lithology: Limestone 100%.

Limestone: White, off-white, medium hard, fossileferous.

Contact: Upper contact with Ghazij Shale and lower with Sui Shale member is conformable.

b. **Sui Shale:**

Age: Lower Eocene.

Lithology: Shale with thin bands of limestone.

<u>Shale:</u> Greenish grey, pyretic, calcareous with occasional fossils.

<u>Limestone:</u> White, medium hard having thickness 8 to 10m.

Contact: Upper contact with Sui Upper Limestone and lower with Sui Main Limestone is conformable.

c. **Sui Main Limestone:**

Age: Lower Eocene.

Lithology: Limestone with traces of shale.

<u>Limestone:</u> Off-white, creamy, Medium to hard, calcitic veins, marly and highly fossilferous.

<u>Shale:</u> Light greenish grey, light grey, laminated fossileferous.

Contact: Upper with Sui Shale and lower with Dunghan Formation is conformable.

VII. **Dunghan Formation:**

Age: Upper Paleocene.

Lithology: Off-white, creamy grey, medium to coarse grained limestone.

Contact: Upper contact with Sui Main Limestone and lower contact with Ranikot Formation is conformable.

VIII. **Ranikot Formation:**

Age: Upper to Middle Paleocene.

Lithology: Nodular limestone with sandstone and argarrigileous shale.

Contact: Upper contact with Dunghan Formation and lower contact with Pab Sandstone is disconformable.

IX. **Pab Sandstone:**

Age: Lower Cretaceous.

Lithology: White to brown colour sandstone with subordinate shale.

Contact: Upper contact with Ranikot Formation is disconformable and lower contact with Fort Munro Formation is conformable.

X. **Fort Munro Formation:**

Age: Lower Cretaceous.

Lithology: Limestone, marl and shale.

Contact: Upper contact with Pab Sandstone is conformable and lower contact with Parh Limestone is conformable.

XI. **Parh Limestone:**

Age: Lower Cretaceous.

Lithology: Light grey micritic limestone and marl.

Contact: Upper contact with Fort Munro Formation is conformable and lower contact with Goru Formation is conformable.

XII. **Goru Formation:**

Age: Middle to Upper Cretaceous.

Lithology: Sand. light grey shale and some gluconitic beds of sandstone.

It has following two units.

a. Upper Goru

b. Lower Goru

- Upper shale
- Middle sand
- Lower shale
- Basal sand
- Talhar shale
- Massive sand

Contact: Upper contact with Parh Limestone and lower Sembar Formation is conformable.

XIII. **Sembar Formation:**

Age: Upper cretaceous.

Lithology: Green to greenish grey shale + thin bands of sandstone and siltstone. Shale is highly fossileferous.

Contact: Upper contact with Goru Formation is conformable and lower contact with Chiltan Limestone is disconformable.

XIV. **Chiltan Limestone:**

Age: Middle to Late Jurassic.

Lithology: Dark grey to black, oolitic to pisolitic limestone.

Contact: Upper contact with Sembar Formation is disconformable and lower contact with Shrinab Formation is conformable.

3.1 Seismic Reflection Method

Seismic method is one of the most commonly used geophysical techniques for investigating the subsurface (Parasnis, 1997). It has advantage over other methods due to greater penetration, higher resolution and accuracy. It is somewhat similar to earthquake seismology with the difference being the energy source. In earthquake seismology source of energy is natural but in exploration seismic source is artificial. Usually dynamite (or vibroseis) is used and the waves generated are recorded by detectors laid along the ground. Variations in reflection times from place to place on the surface usually indicate structural and stratigrapihc features. Depths to reflecting interfaces can be determined from the seismic waves velocity and the corresponding velocity (Telford, 2004).

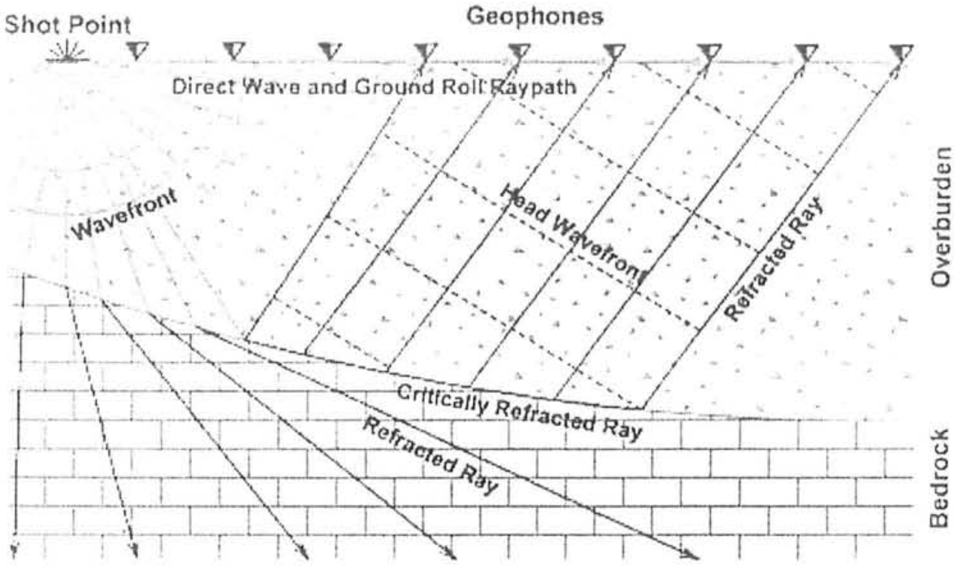
3.2 Basic principal of seismic:

The basic principle of seismic method is the generation, detection of seismic waves. The sound waves rush down and down until they meet an interface layer have acoustic mpedance contrast. A replica at the downward traveling sound wave reflects back towards the surface from the boundary between the two layers.

The original pulse continues its downward journey, gradually becoming weaker, sending reflection back to the surface, every time it encounters an interface. These are usually so many reflections that once they start arriving, they often overlap to form an continues stream of sound. The last reflection to arrive are normally very weak, often 100 thousandth of the strength of the early echoes, and so the geophones that detect them must be very sensitive. The travel times of reflected arrivals from subsurface interfaces are measured between media of different acoustic impedances (Badley, 1985).

3.2.1 Refraction method:

Seismic refraction involves measuring the travel time of the component of seismic energy, which travels down to the top of reflector (or other distinct acoustic impedance contrast), is refracted along the top of reflectors, and returns to the surface as a head wave along a wave front similar to the bow wake of a ship (Figure 4 shows the seismic reflection geometry.) The waves which return from the top of interface are refracted waves, and for geophones at a distance from the shot point, always represent the first arrival of seismic energy (Telford, 2004).



Seismic Refraction Geometry (modified from Robinson & Coruh, 1988)

Seismic refraction is generally applicable only where the seismic velocities of layers increase with depth. Therefore, where higher velocity (e.g. clay) layers may overlie lower velocity (e.g. sand or gravel) layers, seismic refraction may yield incorrect results. In addition, since seismic refraction requires geophone arrays with lengths of approximately 4 to 5 times the depth to the density contrast of interest , seismic refraction is commonly limited to mapping layers only where they occur at depths less than 100 feet (Dobrin, 1988).

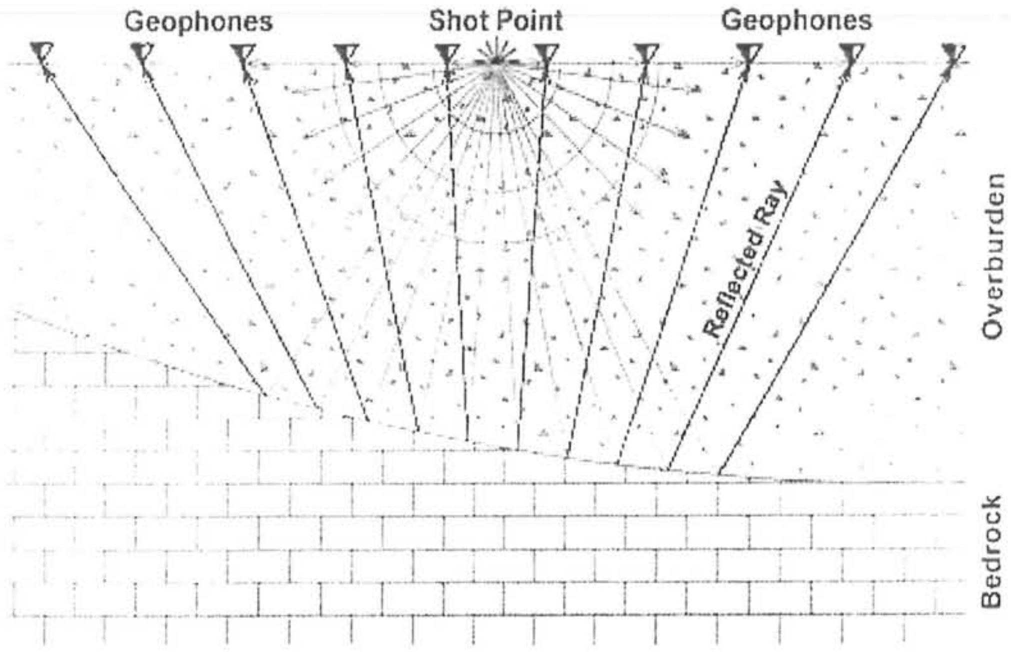
Greater depths are possible, but the required array lengths may exceed site dimensions, and the shot energy required to transmit seismic arrivals for the required distances may necessitate the use of very large explosive charges.

In addition, the lateral resolution of seismic refraction data degrades with increasing array length since the path that a seismic first arrival travels may migrate laterally off of the trace of the desired seismic profile (Yilmaz, 2001). Recent advances in inversion of seismic refraction data have made it possible to image relatively small, non-stratigraphic targets such as foundation elements, and to perform refraction profiling in the presence of localized low velocity zones such as incipient sinkholes (Telford, 2004).

3.2.2 Reflection method:

Reflection method is the most widely used geophysical technique. The structure of subsurface formations is mapped by measuring the times required for a seismic wave (or pulse), generated in the earth by a near surface explosion, mechanical impact, or vibration, to return to the surface after reflections from interfaces between formations having different physical properties. The reflections are recorded by detecting instruments responsive to ground motion (Dobrin, 1988).

Seismic reflection uses field equipment similar to seismic refraction, but field and data processing procedures are employed to maximize the energy reflected along near vertical ray paths by subsurface density contrasts. Figure 5 shows the seismic reflection geometry. Reflected seismic energy is never a first arrival, and therefore must be identified in a generally complex set of overlapping seismic arrivals - generally by collecting and filtering multi-fold or highly redundant data from numerous shot points per geophone placement. Therefore, the field and processing time for a given lineal footage of seismic reflection survey are much greater than for seismic refraction (Kearey, 2002).



Seismic Reflection Geometry (modified from Robinson & Coruh, 1988)

However, seismic reflection can be performed in the presence of low velocity zones or velocity inversions, generally has lateral resolution vastly superior to seismic refraction, and can delineate very deep density contrasts with much less shot energy and shorter line lengths than would be required for a comparable refraction survey depth.

The main limitations to seismic reflection are its higher cost than refraction (for sites where either technique could be applied), and its practical limitation to depths generally greater than approximately 50 feet. At depths less than approximately 50 feet, reflections from subsurface density contrasts arrive at geophones at nearly the same time as the much higher amplitude ground roll (surface waves) and air blast (i.e. the sound of the shot). Reflections from greater depths arrive at geophones after the ground roll and air blast has passed, making these deeper targets easier to detect and delineate (Kearey, 2002). Seismic reflection is particularly suited to marine applications (e.g. lakes, rivers, oceans, etc.) where the inability of water to transmit shear waves makes collection of high quality reflection data possible even at very shallow depths that would be impractical to impossible on land.

3.2.3 Comparison of Refraction and Reflection:

The differences between seismic refraction and reflection are summarized in the given table

	Refraction	**Reflection**
Typical Targets	*Near-horizontal density contrasts at depths less than ~100 feet*	*Horizontal to dipping density contrasts, and laterally restricted targets such as cavities or tunnels at depths greater than ~50 feet*
Required Site Conditions	*Accessible dimensions greater than ~5x the depth of interest; unpaved greatly preferred*	*None*
Vertical Resolution	*10 to 20 percent of depth*	*5 to 10 percent of depth*
Lateral Resolution	*~1/2 the geophone spacing*	*~1/2 the geophone spacing*
Effective Practical Survey Depth	*1/5 to 1/4 the maximum shot-geophone separation*	*>50 feet*

Compression of the seismic reflection & refraction method

3.3 Properties of Seismic Waves :

3.3.1 Theory of Elasticity:

The seismic method utilizes the propagation of waves through the earth, since this propagation depends upon the elastic properties of the rock so it is nessary to know the elastic properties of the subsurface material.

The size of a solid body can be changed by applying forces to the external surface of the body. These external forces are opposed by internal forces, which resist the changes in size and shape. As a result, the body tends to return to its original condition when the

external forces are removed. Similarly, a fluid resist changes in size (volume) but not changes in shape. This property of resisting changes in size or shape and of returning to the undeformed condition when the external forces are removed is called elasticity.

3.3.2 Stress:

"The force (F) applied per unit area (A) of the body"

Its unit in SI system is Pascal and one Pascal is equal to one Newton per square meter.
Mathematically:

$$\textbf{Stress} = \textbf{F/A}$$

3.3.3 Strain:

Strain can be defined "As the change in size and shape of the body when external forces are applied on that body"

These changes are called Strain. It has four types.

- Longitudinal Strain
- Transverse Strain
- Shear Strain
- Dilation

3.3.4 Hooke's Law:

According to this law, "Stress is the directly proportional to strain provided the elastic limit of the body is not exceeded. This limiting value depends upon the nature of rock body".

Mathematically

$$\textbf{Stress} \ \alpha \ \textbf{Strain}$$

3.3.4 Elastic Modules:

The linear relationship between stress and strain in the elastic filed is specified for any mater ial by its various elastic modules, each of which expresses the ratio of a particular type of stress to the strain and provides a measure of rigidity. There are certain types of elastic modules as given below;

- Bulk modulus
- Shear modulus
- Young's modulus
- Poisson's ratio

3.3.5 Bulk Modulus (K):

It is the ratio of stress to the volumetric strain and is given by the relation

$$K = \frac{\text{volume stress}}{\text{volume strain}} = \frac{P(\text{pressure})}{\Delta V/V}$$

Mathematically it can be represented as,

$$k = \frac{1}{\kappa} = \frac{P}{\Delta V / V}$$

Where k is the compressibility coefficient

3.3.6 Shear Modulus (μ) :

The shear modulus is defined as

> "The ratio of shearing stress "τ" to the resulted shear strain "tan θ". It is also called as rigidity modulus. It is denoted by "μ". For liquids and gases, shear modulus (μ) is zero

$$\mu = \frac{\text{shear stress}}{\text{shear strain}}$$
$$= \frac{F/A}{\tan\phi}$$
$$= \frac{F/A}{\Delta L/L}$$

Mathematically it can be represented as (where τ is the shear stress);

$$\mu = \frac{\tau}{\tan\Theta}$$

3.3.7 Young's Modulus (E):

It is defined as the

"The ratio between longitudinal stress and longitudinal strain. It is also called stretch modulus". It is denoted by "E"

$$E = \frac{\text{longitudinal stress}}{\text{longitudinal strain}}$$
$$= \frac{F/A}{\Delta L/L}$$

Mathematically Young's Modulus can be represented by the shear (μ) and bulk module (k).

$$E = \frac{9k\mu}{3k + \mu}$$

3.3.8 Poisson's Ratio (σ):

It is used to show that the change in diameter (d) is proportional to the change in length (l). Poisson's ratio varies from 0 to ½ and has the value ½ for fluids.

$$\sigma = \frac{\text{transverse strain}}{\text{longitudinal strain}}$$
$$= \frac{\Delta D/D}{-\Delta L/L}$$

Mathematically it is represented as Bulk modulus (k) and Shear modulus (μ);

$$\sigma = \frac{3k - 2\mu}{2(3k + \mu)}$$

Relationship between Elastic Module

The all four module can be interrelated in the following way (Dobrin, 1988);

$$K = E / 3(1-2\sigma)$$

$$\mu = E / 2(1+\sigma)$$

3.4 Seismic Waves:

Wave is a progressive disturbance propagated from point to point in a medium or space without progress or advance by the points themselves. Seismic waves are generally referred to as elastic waves because they propagated like that in an elastic band when it is stretched.

The theory of elasticity reveals that the energy propagated through the earth in the different form of seismic waves.

Seismic waves are parcels of elastic strain energy that propagate outwards from a seismic source such as an earthquake or an explosion (Kearey, 2002).

3.4.1 Laws governing seismic waves:

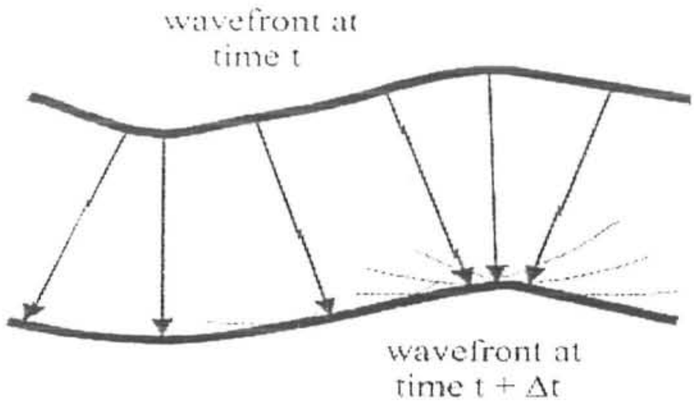
There are three fundamental laws that govern the seismic wave propagation.

- Huygen's principle
- Fermat's principle
- Snell's law

3.4.2 Huygens's Principal:

According to this principal,

> "Every point on a wave front is a source of new wave that travels away from it in all directions" Figure 6 shows the generation of wave fronts by succeeding waves



Huygens's Principle (modified from Robinson & Coruh, 1988)

3.4.3 Fermat's Principal:

It states that

> "Elastic waves travel between two points along the paths requiring the least time"

3.4.4 Snell's Law:

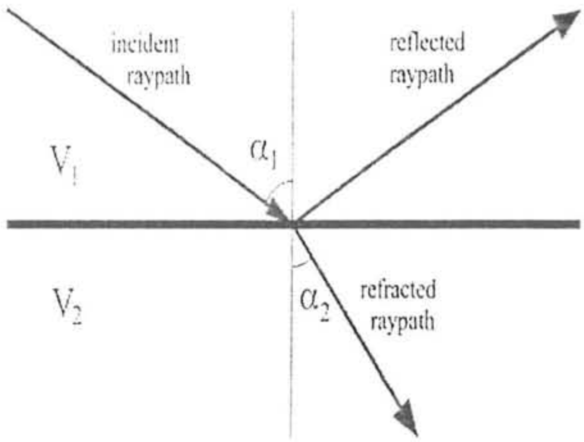
According to this law

> "Direction of refracted of reflected waves traveling away from a boundary depends upon the direction of the incident waves and the speed of the waves"

Mathematically,

$$\frac{\sin(\alpha_1)}{V_1} = \frac{\sin(\alpha_2)}{V_2}.$$

Where V_1 and V_2 are velocities in the upper and lower layers, α_1 is the angle of the incident ray-path with respect to the vertical, and α_2 is the angle of transmission of the refracted ray-path with respect to the vertical. Figure 7 shows the reflection of incident waves form a surface.



Refraction and reflection of an incident wave (modified from Robinson & Coruh, 1988)

3.4.5 Reflection- and transmission-coefficients:

To derive the reflection and transmission coefficients for elastic waves, the boundary conditions at the interface are needed and are described by the Zoeppritz-Equations. These reflections coefficients depend on

- Difference in density
- Difference in velocity
- Angle of incident of the wave

The Reflection- and Transmission coefficient give the ratio between the incident amplitude A_0 and the reflected (A_R) and transmitted (A_T) amplitude, respectively. In the special case of a incident wave perpendicular at an interface for a P-wave, a simple expressions for the reflection and transmission coefficient is obtained.

3.4.6 Reflection coefficient:

These coefficients compare the amplitude of incident wave and reflected wave. Value of reflection coefficient varies from -1 to +1(Khan, 1988). For R=0, there will be no reflection, wave will be transmitted. It can be mathematically represented as;

$$R = \frac{A_R}{A_0} = \frac{v_2\rho_2 - v_1\rho_1}{v_2\rho_2 + v_1\rho_1} = \frac{Z_2 - Z_1}{Z_2 + Z_1}$$

3.4.6 Transmission coefficient :

Transmission coefficients are those which compare the amplitude of incident wave and refracted wave. Value of transmission coefficient varies from 0 to2 (Khan, 1988).

If Ai is the amplitude of incident wave and At is the amplitude of transmitted wave, then transmission coefficient T is given as follow;

$$T = \frac{A_T}{A_0} = \frac{2v_1\rho_1}{v_2\rho_2 + v_1\rho_1} = \frac{2Z_1}{Z_2 - Z_1}$$

The product $Z = v\,\rho$ is known as the acoustic impedance.

Reflection and transmission in terms of energy:

Sometimes the coefficients which describe the energy and not the amplitudes are introduced as Reflection- and Transmission coefficients

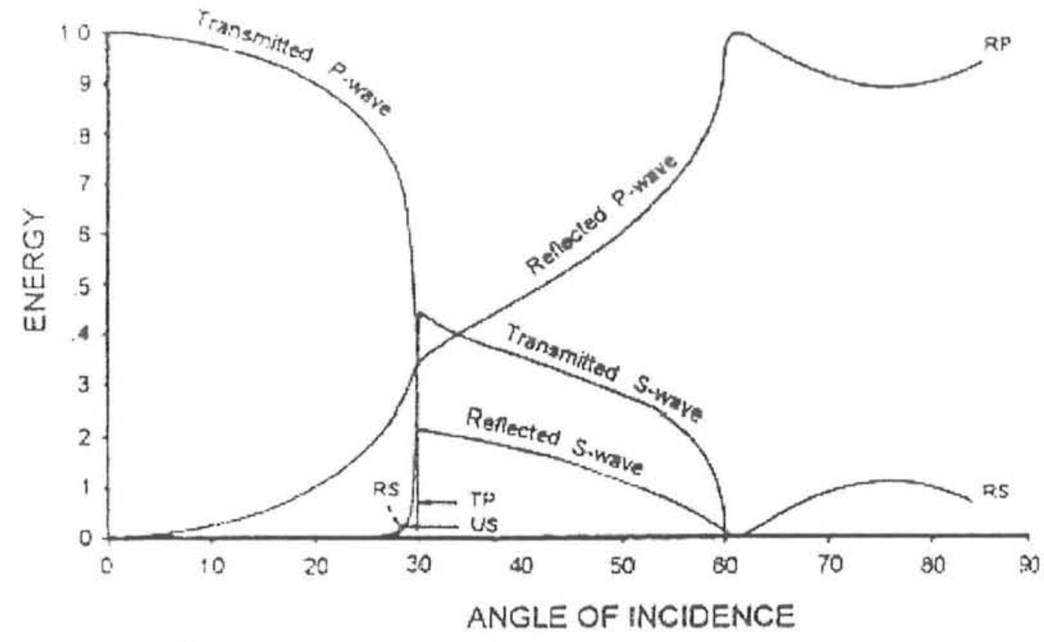
Reflection coefficient in terms of energy

$$E_R = \frac{(Z_2 - Z_1)^2}{(Z_2 - Z_1)^2}$$

Transmission coefficient in terms of energy

$$E_T = \frac{4Z_1Z_2}{(Z_2 + Z_1)^2}$$

- Obviously the total amount of energy is the same before and after the reflection and transmission. so that $E_R + E_T = 1$
- In a general case these coefficients are depending on the angle of incidence, also conversions between P- and S-waves occur at an interface (Figure 8 shows the angle dependent reflection and transmission co-efficents for P and S-waves.)



The angle dependent reflection and transmission co-efficents for P and S-waves (Rehman, 1989)

3.5 Types of seismic waves:

Seismic waves are messengers that convey information about the earth's interior. Basically these waves test the extent to which earth materials can be stretched or squeezed some what as we can squeeze a sponge. They cause the particles of materials to vibrate, which means that passing seismic waves temporarily deforms these particles can be described by its properties of elasticity. These physical properties can be used to distinguish different materials. They influence the speeds of seismic waves through those materials (Robinson & Coruh, 1988).

These waves are generated by Earth's material as a result of an earthquake or an explosion. Seismic waves are of two types; the body waves and surface waves. When a stress is suddenly applied to an elastic body or when stress is suddenly released the corresponding displacement is propagated outward as an elastic wave. Different types of propagation give rise to different waves. So seismic waves can be divided into two parts,

- Body waves

- Surface waves

3.5.1 Body Waves:

Theses are these waves which can travel though the earth interior and provide vital information about the structure of the earth. The body waves can be further divided into the following:

- P- waves (Primary waves)
- S- waves (Secondary waves)

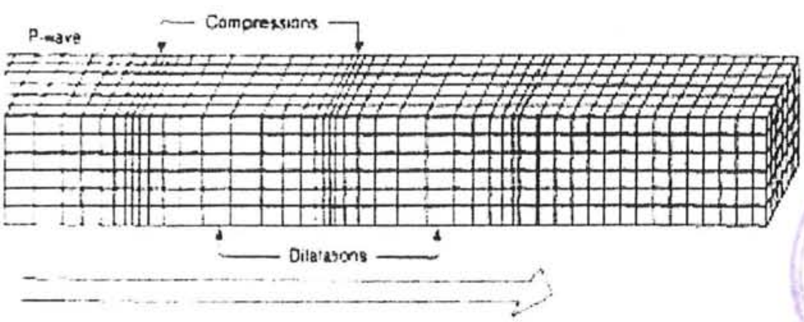
3.5.1.1 P –waves (Primary Waves):

The particular kinds of waves of most interest to seismologists are the compressional or P-waves also called as compressional waves, longitudinal waves, primary waves, pressure waves, and dilatation waves (see Figure 9). In this case the vibrating particles move back and forth in the same direction as the direction of propagation of waves. P-waves can pass through any kind of material - solid liquid or gas. The P-waves velocity depends upon density and elastic constants (Dobrin, 1976).

The seismic velocity of a medium is a function of its elasticity and can be expressed in terms of its elastic constants. For a homogeneous, isotropic medium, the seismic P-wave velocity Vp is given by;

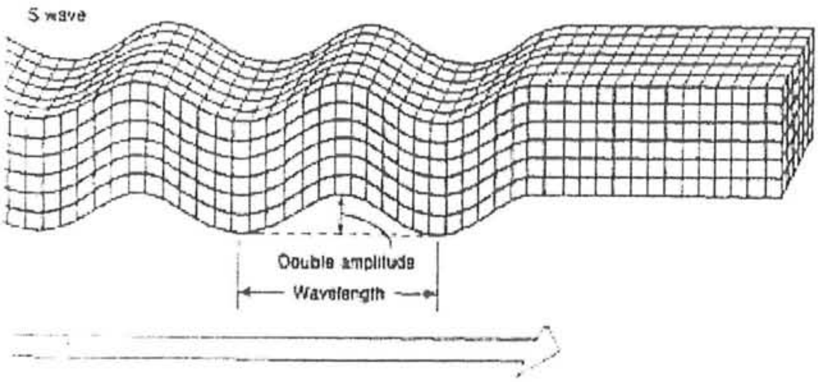
$$Vp = \sqrt{\frac{(4/3)\mu + k}{\rho}}.$$

Where μ is the shear modulus, k is the bulk modulus and ρ is the density of the medium.



3.5.1.2Swaves(Secondary waves):

In shear waves, the particles vibrate in a direction perpendicular to the direction of propagation of waves (see Figure 10).



The propagation of S-waves in an elastic medium (modified from Robinson & Coruh, 1988)

They are also called as Shear waves, transverse waves, and converted waves. For ideal gases and liquid μ=0. S-waves cannot pass through fluids (Dobrin, 1976). The velocity of S-waves is given by (using the same notation as of V_p);

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

Characteristics of Body Waves

- These travel with low speed through layers close to the earth's surface, as well in weathered layers. (Robinson & Coruh, C, 1988)
- Frequency of body waves in exploration vary from 15Hz to 100 Hz. (Parasnis, 1997)

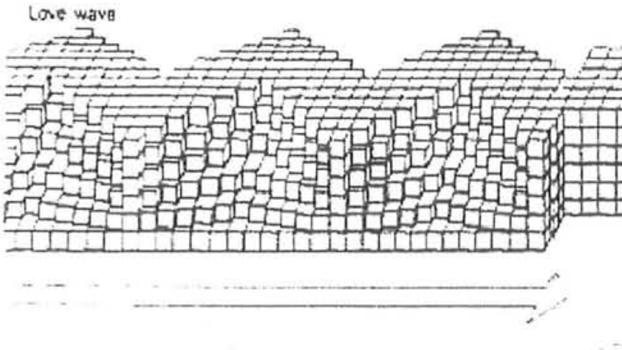
3.5.2 Surface Waves:

A part from body waves more complicated patterns of vibration are observed as well. These kinds of vibrations can be measured only at locations close to the surface. Such vibrations must result from waves that follow paths close to the earth's surface ,hence known as surface waves. In a bounded elastic solid, surface waves can propagate along the boundary of the solid. Frequency of surface waves is less than 15Hz (Parasnis, 1997). Surface waves are also of two types;

- Raleigh waves
- Love waves

3.5.2.1 Love Waves:

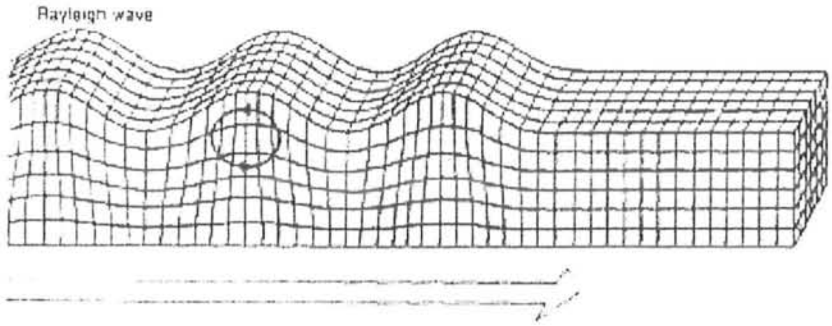
A type of surface waves having a horizontal motion i.e. transverse to the direction of propagation (Kearey, 2002). The velocity of these waves depends on the density and modulus of rigidity and not depends upon the bulk modulus (k). The Figure 11 shows the propagation of Love-waves in an elastic medium.



The propagation of Love waves in an elastic medium (modified from Robinson & Coruh, 1988)

3.5.2.2 Ragleigh Waves:

Type of surface waves having a retrograde, elliptical motion at the free surface of a solid and it is always vertical plane. Raleigh waves are principal component of ground roll (Kearey, 2002). The Figure 12 shows the propagation of Ragleigh waves in an elastic medium.

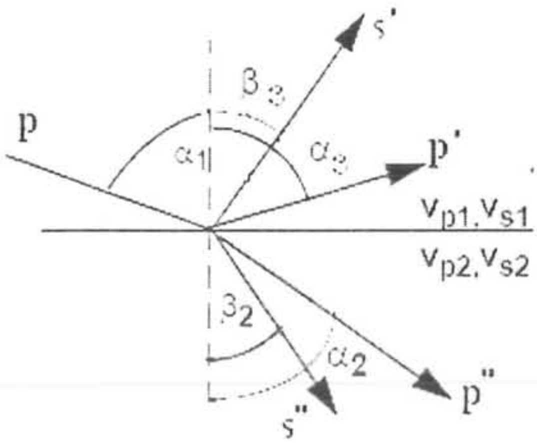


The propagation of the propagation of Ragleigh waves in an elastic medium (modified from Robinson & Coruh, 1988)

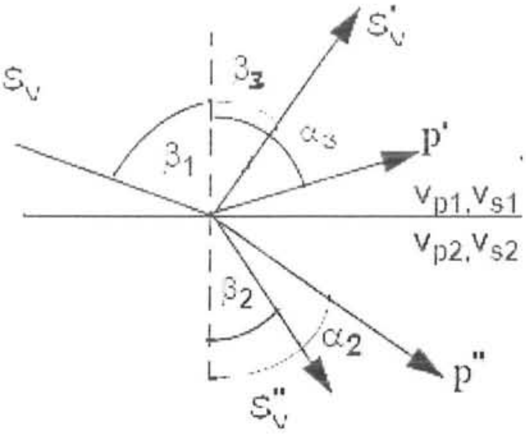
3.6 Wave Conversion:

When a wave reaches the boundary between two substances having velocities, it divides up into waves that reflect from the boundary or refract across the boundary. So an incident wave is converted into reflected and refracted waves. An incident wave can be P-wave, S_V- wave or S_H-wave (Robinson & Coruh, 1988).

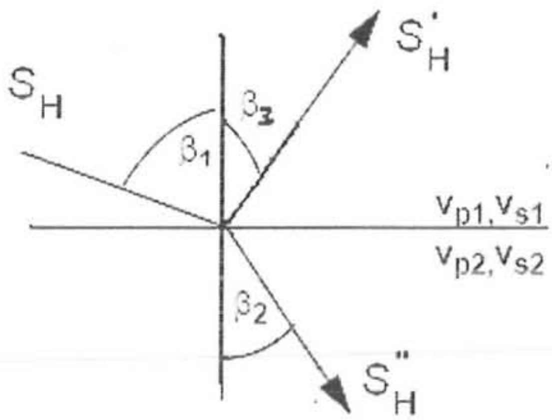
- When incident wave is P-wave then it is reflected and refracted as P-wave and S-wave (see Figure 13 a)
- When incident wave is S_V-wave then it is reflected and refracted as P-wave and S_V-wave .
- When incident wave is S_H-wave then it is reflected and refracted as S_H-wave



Wave conversion of P-wave into various waves



Wave conversion of S_V-wave into various waves (modified from Robinson & Coruh, 1988)



Wave conversion of S_H-wave into various waves (modified from Robinson & Coruh, 1988)

Where

- P is the incident wave
- P' is the reflected wave
- P" is the refracted wave
- S is the incident wave
- S 'is the reflected wave
- S " is the refracted wave
- S is the incident wave
- S ' is the reflected wave
- S "is the refracted wave
- α1 incident angle of P-wave
- α2 reflected angle of P-wave
- α3 refracted angle of P-wave
- β1 incident angle of S-wave
- β2 reflected angle of S-wave
- β3 refracted angle of S-wave

4.1. Intrduction

The purpose of seismic data acquisition is to record the ground motion caused by a known source in a known location. First step in seismic data acquisition is to generate a seismic pulse with a suitable source. Second is to detect and record the seismic wave propagating through ground with a suitable receiver (geophone/seismometer), digital or analogue form (Kearey et al, 2002).

The record of ground motion with time is called as seismogram. It is the basic information to be used in seismic data interpretation. During the seismic data acquisition certain operations are to be applied, such as conversion of the ground motions into electrical signals, amplification of these signals and filtering of the signals as well.

4.2. Acquisition Setup

It includes

- The spread configuration.
- Shooting types
- Shooting parameters
- Recording parameters.

The Spread Configuration

For acquisition of data and as well as to have quality of data high certain field operations are adopted. So the first step in this practice is the choice of spread type. The spread is defined as the lay out on the surface of, of the detectors, which give recorded output for each source. Spread is made up of equal inter-receiver distance and a defined offset. There is certain number of spreads called as basic spreads and shown in Figure 4.1

These spreads are

- End Spread.
- In-Line Offset Spread.
- Split Spread/Centre Shooting.
- Cross Spread.
- L Spread.

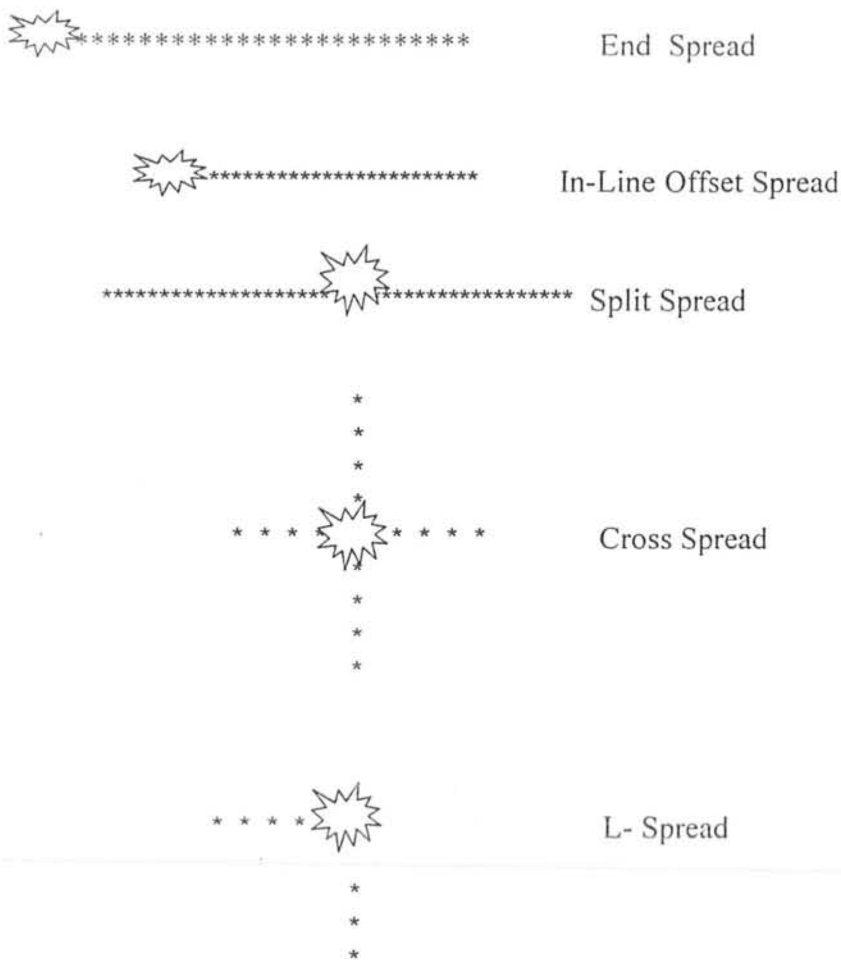


Figure 4.1: The basic spreads used in seismic acquisition

Along with these basic spreads, another technique called as Fan shooting can also be used. In this technique ,geophones are arranged in an arc, fanning out in different directions from the source (Robinson &Coruh,1988).

Fan shooting is used in Transmission method. Transmission method differs from normal refraction method in the sense that it does not involve critical incidence of waves over the interface. In Transmission method, source and detector are on opposite sides of the investigated interface. Other techniques used in transmission method in velocity logging(well shooting, continuous velocity logging(CVL) and uphole survey).

Fan shooting is used for determination of the dimensions of velocity anomalous structures (Al-Sadi,1980).

Shooting Types

There are different types of shootings used in the field. These types are:

- Symmetric shooting (The number of channels on sides of source is same.)
- Asymmetric shooting(The number of channels on sides of source is not same.)
- End shooting. (The source is at one end of the spread)
- Roll in/Roll out shooting. (In roll along method receivers are added in the spread while shooting in the source along the spread. Roll-out shooting is one in which the receivers are removed from the spread while shooting out along the spread.)

Shooting Parameters

Shooting parameters include

- Source size.
- No. of holes.
- Hole depth.
- Shot at or between the pickets.

Recording Parameters

The recording parameters include

- Group Interval
- Group Base
- Number of Channels
- Number of Geophones in a Group
- Geophone Array(Linear or Weighted)
- Sample Rate
- Record Length
- Coverage (Folds)
- Zero Offset and Common Offsets

Group Interval

Group is defined to be as combination of a certain no of geophones. Group interval is the interval between the mid points of two consecutive groups.

Group Base

It is the total length of that collectively feed a channel. Group base can act for the suppression of noise.

Number of Channels

Total number of recording units used in the survey.

Number of Geophones in a Group

Total number of geophones in a group

Geophone Array

It can be linear, weighted or aerial. Linear array means that all geophones are arranged along a line. Weighted array means that geophones in a group are arranged in form of parallel lines. In this array sensitivity of each geophone is made different. In aerial array, geophones are laid over an area around the line.

Sample Rate

The time in digital recording, during which discrete samples are recorded.

Record Length

The total length of for recording one short is called as record length.

Coverage (Fold)

Coverage means how many fold data will be obtained from the multifold profiling survey (Robinson &Coruh, 1988). Number of folds from a survey can be determined as follows

$$\text{Fold number} = N\,\Delta x / 2\Delta s$$

Where, N = number of recording channels

Δx = geophone interval

Δs = source interval.

Zero -Offset and Common Offset

Zero offset data is characterized when the source and receiver are present on the same location. There is no move out. For a normal measurement this is seldom the case. When the traces are corrected for the move out and are stacked then a zero offset trace is obtained.

All traces with equal offset between source and receiver. This configuration is often used for several Single channel systems. Also georadar measurements are often carried out with a fixed offset between source and receiver. The minimum possible distance between source and first active channel is called as near offset. The distance from the source to last active channel is called as far offset.

4.3. ENERGY SOURCES

The mechanical disturbance which is at the origin of a seismic observation is generated by displacing momentarily a small volume of rock from its rest position. There are different ways of doing so that whether to apply such an artificial disturbance on land or in water.

Seismic source on land

- **Impact:** Sledge hammer, Drop weight, Accelerated weight
- **Impulsive:** Dynamite. detonating cord, Air gun, Shotgun, Borehole sparker
- **Vibrator:** Vibroseis Vibrator plate, Rayleigh wave generator

Seismic source in water

- **Impulsive:** Air gun. Gas gun. Sleeve gun. water gun, Steam gun, Pinger, Boomer, Sparker
- **Vibrator:** Multipulse. GeoChirp

4.3.1. Explosive/Impulsive Sources

Impulsive sources are capable of liberating broad band signal (energy) in a very short time (Maurice & Sercel, 1997) include mechanical techniques to generate seismic waves. Dr. Burton McCollum has been a pioneer in the application of this principle to seismic prospecting. The first, historically speaking, (and still the most frequently used for land operations) is dynamite shooting. A charge is exploded, generating a shock wave and pushing the surrounding medium away from its original position. Basically, two types of explosives have been used principally: gelatin dynamite and ammonium nitrate (Telford et al., 1990). The velocity of detonation is higher for the explosives used in seismic work, around 6000-7000 m/sec; consequently, the seismic pulses generated have very steep fronts in comparison with other energy sources (Telford et al., 1990). Dynamite shooting is seldom used anymore for off-shore exploration because of the hazards involved in carrying big quantities of explosive materials in a survey boat. On land, the charge is exploded at the bottom of a hole drilled for the purpose, at the surface of the ground, or in the air above the surface. Shot-hole shooting is the technique used most in which the seismic energy so generated can be varied over a wide range by selecting the size of charge best adapted to local conditions (Maurice and Sercel, 1997).

The drawback of shot-hole shooting is the necessity to drill holes. In spite of that, drilling may prove a heavy burden under some terrain conditions. It is to obviate that burden that surface shooting or air shooting are used sometimes. Part of the energy is lost in the air instead of being imparted to the ground, and these techniques require more explosive than buried shooting, but the cost of the additional explosives may be far less than the cost of drilling shot-holes.

4.3.2. Non-Impulsive Sources

4.3.2a Weight Dropping

Historically, the first means of creating a seismic disturbance has been a mechanical device: weight dropping. A heavy weight dropped from a few meters generates an elastic wave in the ground. The amount of energy can be increased within reasonable limits by increasing the mass of the weight or the height from which it is dropped. In all cases, the potential energy is far less than that of a common explosive charge but very little of it is lost in permanent deformations of the ground, and the seismic results are far better than what the mere consideration of potential energy might lead us to think. Weight dropping is not much used nowadays because, it develops a large amount of surface noise and this noise must be minimized before reflections can be interpreted (Maurice and Sercel, 1997).

4.3.2b Vibrators

On land. the main competitor of dynamite is the vibrator. Vibrators offer means of controlling the wave generation that the other sources cannot provide. The basic principle to generate a long duration signal instead of an impulse as shown in fig. 4.1, explained by Crawford et al. 1960. Although the seismic wave attenuation results during propagation through earth but id doesn't modify the sweep signal duration (Maurice and Sercel, 1997). It is let to the objectives that the sweep signal length is estimated as a function of frequency and time (Maurice and Sercel, 1997). When the frequencies are swept from the lower to higher one, the signal is said to be up sweep and down sweep, when from frequencies are swept from higher to lower one (OGTI manuals 1988, Maurice and Sercel, 1997).

The Vibrator has an advantage over impulsive sources because of safety, low cost, fixed band of frequency generation so that the earth does not severely attenuate and the same penetration can be reached with far less energy at the origin. Also cross correlation is a powerful filter which retains only those components common to both functions. This is an additional advantage when the survey is to be carried in densely inhabited districts where cultural noise is impotent (after Maurice and Sercel, 1997).

Among other things, the instantaneous amplitude can be reduced to avoid destructive effects when operating in a built-up area and this loss of intensity can be compensated for by a longer duration of the pulse. Vibrators and weight dropping have been used offshore but they present practical difficulties of operation in the water and thus are essentially restricted to land operations.

4.3.3. Marine Sources

Water shooting or Implosion, is a special case of underwater explosions. The hot gases liberated by the explosion (Air Gun) expand, repelling the water away and creating a cavity which increases in size as the pressure decreases (Maurice and Sercel, 1997, Kearey and Brooks, 2002). When the hydrostatic pressure is reached, the bubble (Kearey and Brooks, 2002) is still expanding and, due to the inertia of the water moving outwards, the volume of

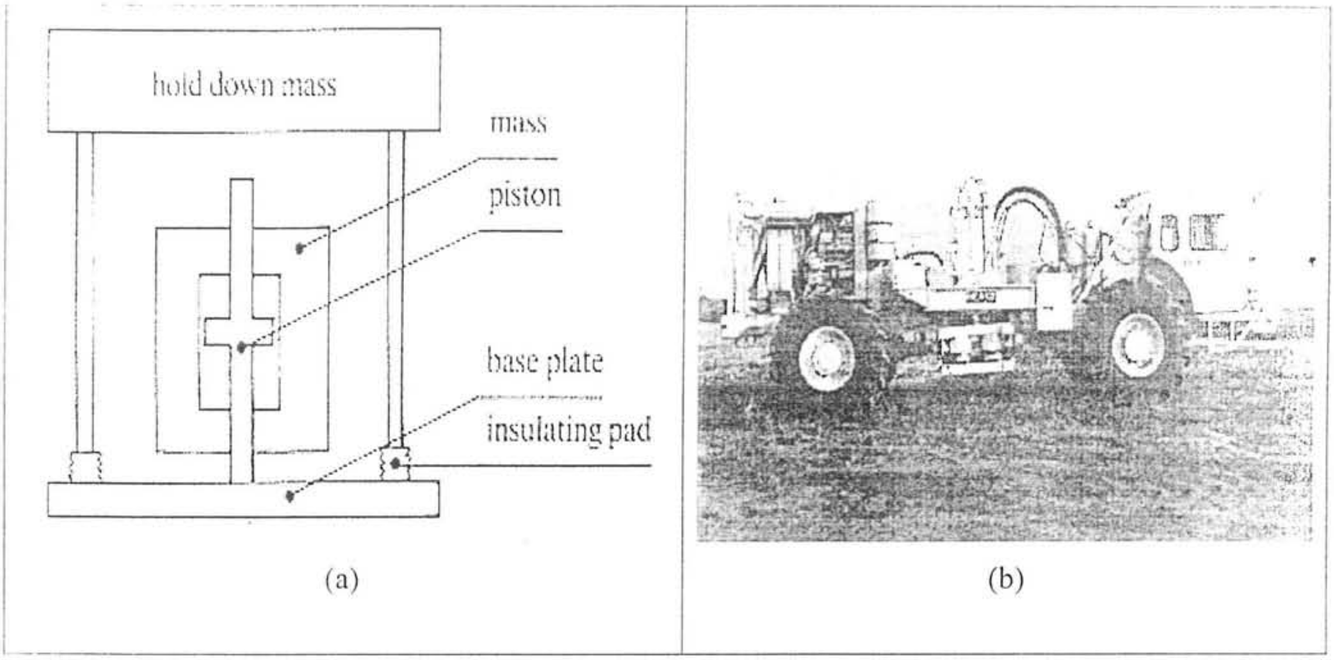


Fig. 4.1 The basic principle of vibrator (a), and the vibrator Truck M-10/601 Mertz Nomad Vibrator Mertz Inc. Ltd. Ponca City).

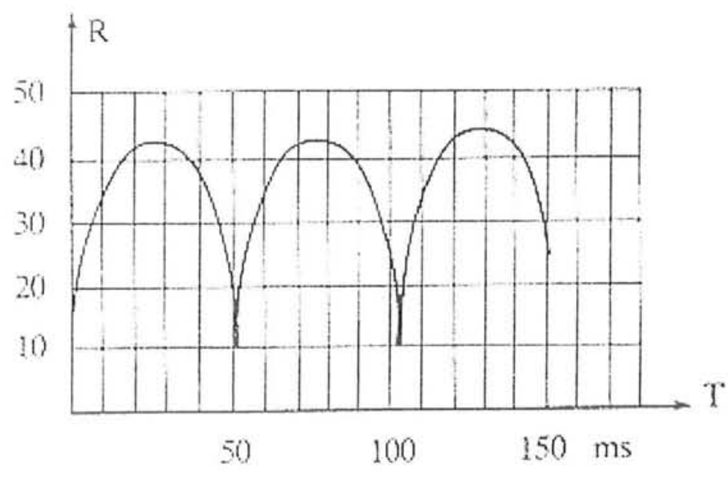


Fig. 4.2 Radius R of the bubble as a function of time T.

cavity keeps on increasing whereas the pressure decreases below the hydrostatic pressure thus generating the inverse pressure force at the surface of bubble which slows the expansion down and stops it. The bubble shrinks, and again hydrostatic pressure builds up. so it transmits the energy down by bubble effect. The oscillations are not symmetric (fig. 4.2): the minimum is very sharp. Such bubbles are repeated several times so recording all event several times. complicating the interpretation. This can be avoided by replacing the explosion by implosion (Water Gun) which is a collapse (Maurice and Sercel, 1997).

The electrical seismic sources constitute the last category worth listing. The spark of an arc releases energy which can be coupled to the surrounding medium. Electrical sources are not much used and they are almost exclusively restricted to offshore operations where water provides a good coupling. studied to determine the nature of the coherent events on the records. Once there is an indication of noise found we can design our array and field layout so that it can be attenuated.

4.4. DETECTORS

4.4.1. Geophone

The captors used on land to detect seismic ground motion (Kearey and Brooks, 2002) are known as seismometers or geophones. Such a motion sensitive instrument, if operated in the water. will term as pressure sensitive geophone or (Maurice and Sercel, 1997). Most of the geophones are based on the principle of a moving coil (Kearey and Brooks, 2002) as in the fig. 4.6. The cylindrical coil is suspended in a magnetic field by a leaf-spring. The passage of a seismic wave at the surface causes a physical displacement of the ground which moves the geophone case and magnet in sympathy with the ground but relatively to the coil because of its inertia. This relative movement of magnet with respect to the coil results in a small voltage being generated across the terminals of the coil in proportion to the relative velocity of the two components. Geophones thus respond to the rate of movement of the ground (i.e. particle velocity) not to the amount of movement or displacement.

Problem: The system is an oscillatory system with a resonant frequency depending upon the mass of the spring and the stiffness of the suspension. It oscillates strongly at the resonance frequency. This can be prevented by damping after Maurice and Sercel, 1997.

- Under critical Damping ($h<1$): Damping is too less, the geophone is not ready for the arrival of the next event.
- Critical Damping ($h=1$): minimum amount required which will stop any oscillation of the system from occurring.
- Over critical Damping ($h>1$): Signal has a large damping and is not sensitive enough for a measurement.

Aim:

- Response in frequency domain should be flat (above natural frequency).
- Response in frequency domain should not have any phase differences.

Most geophones are slightly under damped, typically around $0.6<h<0.66$. This damping can be changed by the addition of a shunt resistor across the coil terminals.

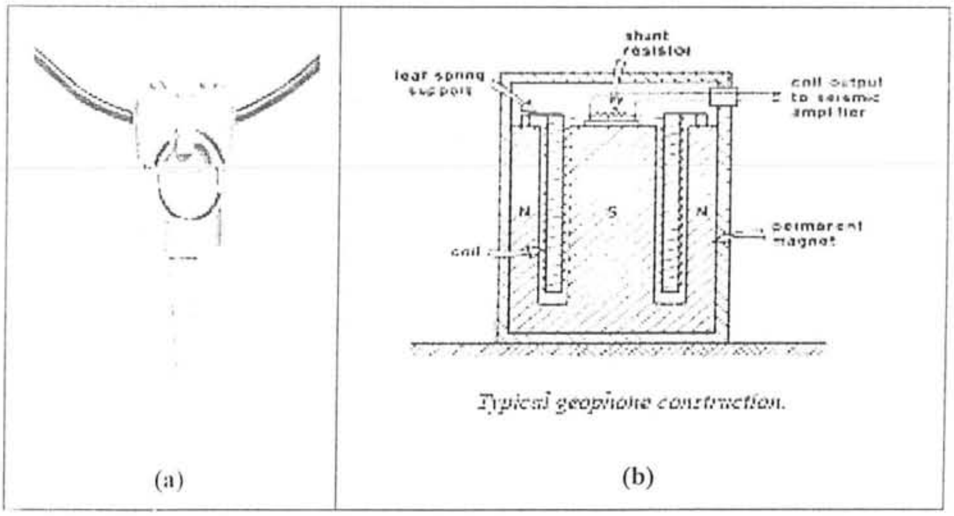


Fig. 4.6 Basic geophone design used in fields (a), and typical geophone construction (b) (after Maurice and Sercel, 1997).

4.4.2. Hydrophones

Hydrophones are use to detect the seismic waves propagating in water. It is composed of ceramic piezoelectric elements (see fig. 4.7) which produce an output voltage proportional to the pressure variations associated with the passage of a compressional seismic wave through water. The sensitivity is typically 0.1 mV Pa^{-1} (Kearey and Brooks, 2002). For multichannel seismic surveying in sea, large numbers of individual hydrophones are made up into hydrophone streamers by distributing them along an oil-filled plastic tube.

4.5. RECORDING SYSTEMS

4.5.1. Analog Recording System:

For the first thirty years or so of seismic exploration, the outputs of the amplifiers were recorded directly on photographic paper by means of a camera. However, about 1952 recording on magnetic tape began. Today few seismic crews are not equipped for magnetic tape recording. The feature which originally led to widespread use of magnetic recording was the ability to record in the field with a minimum of filtering, automatic gain control, mixing etc., and then introduces the optimum amounts of these on play back. Analog magnetic tape recorders usually have heads for recording 26 to 50 channels in parallel (Telford et al. 1990).

Analog systems are systems for which the input and output are analog signals i.e. continuous amplitude signal (Oppenheim, 2002). For an analog seismogram is a continuous record of ground motions as a function of time (fig. 4.12a). The analog recording system is made up of an electric unit normally housed in recording station. Before the signal is recorded by analog system, it can be electronically amplified and filtered. The amplifier is used to increase the strength of weak geophone signals. Some of the signals may be removed by means of electronic filtering before recording the signals. An analog seismic recording system is equipped with a separate amplifier, filter circuit and a magnetic tape for each geophone. These components make up one channel of the recording system.

4.5.2. Digital Recording System:

One of the most significant developments in seismic technology has been introduction of the digital recording in the field first introduced into seismic work early in the 1960s. Digital recording represents the signal by a series of numbers which denote values of the output of the geophone (fig. 4.12b) measured at regular interval, usually 2 or 4 milliseconds. A digital

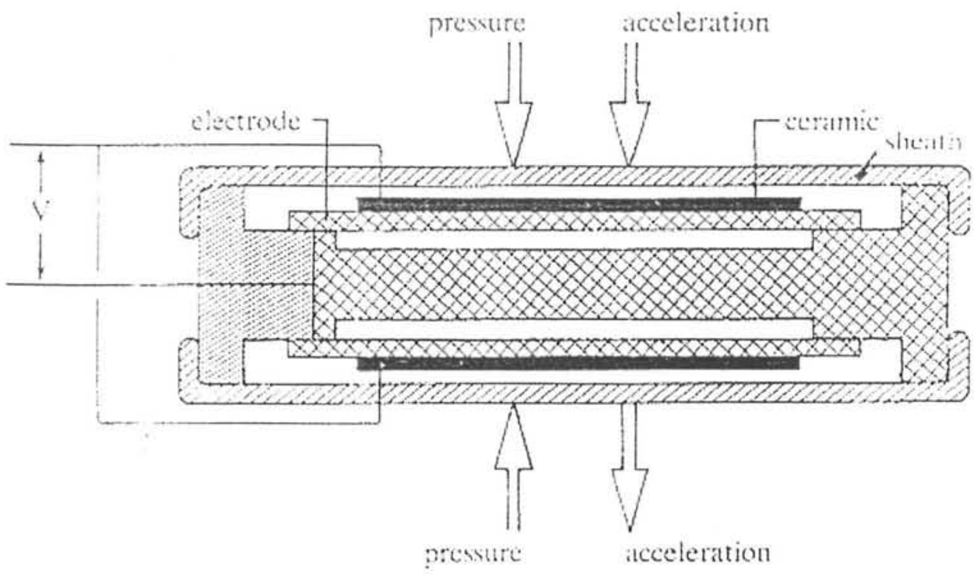


Fig. 4.7. A schematic cross-section of a piezometric hydrophone

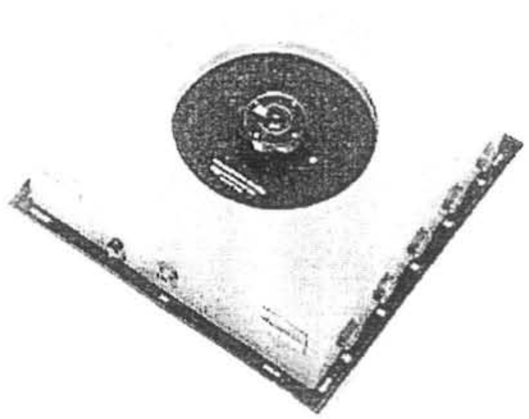


Fig. 4.8 A type of Roll Along Switch: RLX 240 M (after Maurice and Sercel, 1997).

recorder makes use of binary numbers to store the measuring of geophone signal strength. In a multichannel system, each geophone signal is first amplified and filtered by analogue to digital converters (A/D Converter) and a procedure is accomplished by mean of a high speed switch called a Multiplexer. One of the most important advantages of digital seismic recording system is the large increase in dynamic range (i.e. 100 DB) over analogue system (Robinson & Coruh, 1988).

The digital recorder makes use of binary numbers to store the measurement of geophone signal strength and has a significant advantage for the purpose of computer processing. The digital data recorded on a tape is in the form of binary numbers. Each digit of binary number on the tape is called "bit". If the recording head magnetize this bit then it indicates "1" otherwise "0".Digital recording system contains various units including Multiplexer, A/D converter, Formatter, amplifiers, filters etc.

4.5.3. Registration unit

Components of a registration unit, include:

- Signal from geophone
- Preamplifier
- Filter
- (Multiplexer)
- A-D (analog to digital)-converter
- Saving of the result

A signal from the mechanical shake of the ground is being sensed by the geophone as an analog continuous signal, and a seismogram is constructed showing how the amplitude of this signal varies with time. It goes from cable to Roll-along switch (see fig 4.8.), first which has many input seismic channels and outputs the seismic data recording system.

4.5.3a. Pre-amplifier

It is a fixed gain amplifier possibly a front end (fig. 4.9) of the seismic data acquisition (Maurice and Sercel, 1997), consisting of various number of preamplifier-filters equal to the number of seismic channels. The preamplifiers are followed by a multiplexer where the input signals are chopped into short time slices which are intermingled and queued through a single output channel. At this time, samples can be taken and held in a sample-and-hold amplifier. Each sample is raised to a proper voltage for accurate recording.

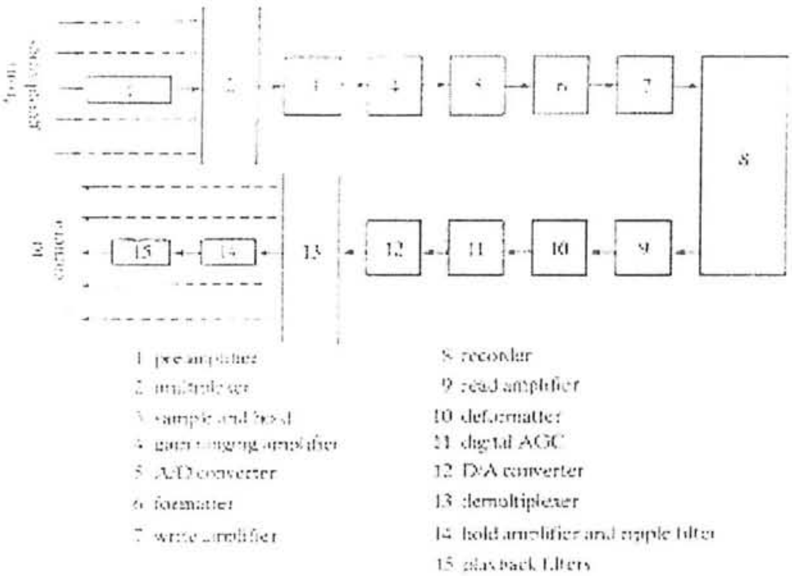
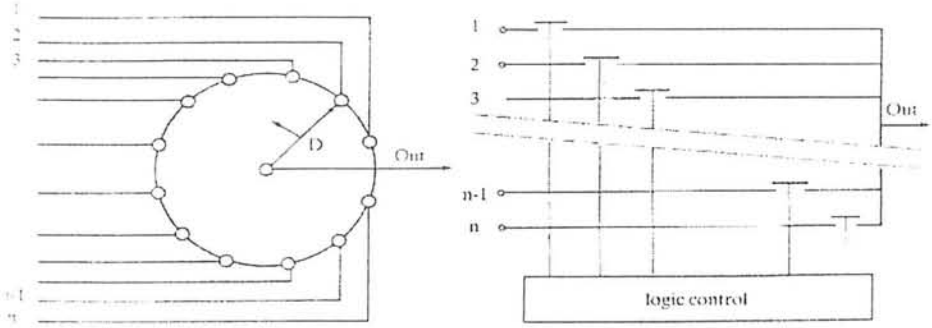


Fig. 4.9. A complete recording unit with preamplifier unit labeled 1 and other parts. It shows how other units do are connected to pass their output to recorder and playback filters.



Multiplexer

Fig. 4.10 Basic principle and procedure of a multiplexer.

4.5.3b. Multiplexer

A multiplexer is a kind of selector switch connecting several inputs successively to a single output (A/D Converter, after Telford et al, 1990). It acts like the distributor of an internal combustion engine but in a reverse manner: the multiple inputs replace the connections to the spark plugs and the single output stands for the connection to the coil (fig. 4.10). The sliding contact D, turning at a constant speed, connects inputs 1, 2, 3, etc... one after the other and in this order to the output. After the last input, n, input number 1 is connected again to the output and the sequence repeats. For the processing all channels must be sorted out which is called demultiplexing.

4.5.3c. AD-Converter

It is much easier to apply mathematical operations on digital data than on analog data. Therefore an analog signal has to be converted into a digital signal. Every Bit (0 or 1) corresponds to a certain voltage that each time differs with a factor two. The A/D-Converter compares the analog signal with the different voltage steps and adds these in such a way that the smallest error between the analog and digital signal occurs. It is clear, that it is impossible to measure a signal which amplitude is larger than the sum of all separate steps or smaller than the smalls step. For a measurement, the gain must be set such that the amplitude of the measured data lies in the range of the AD converter (see fig. 3.12).

We speak of 8-bit, 16-bit, 20-bit und 24-bit sampling:

Example:

- 8-bit: 1 mV-256 mV
- 24-bit: 1 μV-16 V

4.6. SAMPLING

By measurements using a digital system, the data is not continuously measured, but at a specific time interval measured and transported to the AD-converter.

4.7 ALIASING

It is a phenomenon during sampling usually occurs if higher frequencies are folded back into the Nyquest interval. Sampling frequency is the number of sampling points in unit time or unit distance. Thus if a waveform is sampled every two milliseconds (sampling interval: Δt=0.002), the sampling frequency is 500 samples per second (or 500 Hz). Sampling at this rate will preserve all frequencies up to 250 Hz in the sampled function (fig. 4.13a.b). This frequency of half the sampling frequency is known as the Nyquist frequency (f_N) and the Nyquist interval is the frequency range from zero up to f_N

$$f_N = \frac{1}{2\Delta t}$$

No information is lost as long as the frequency of sampling is at least twice as high as the highest frequency component in the sampled data. Seismic measurement systems have often an analog **Anti-Alias-Filter**, which suppresses all Frequencies above the Nyquist-Frequency.

4.8 Seismic Noise

All type of disturbances created and interference with the signal of interest is called a noise.

Noise is divided into two types:

- Coherent noise.
- Incoherent noise.

Coherent Noise

Coherent noise displays some regular patterns on a seismogram. Often it consists of recognizable waves such as surface waves, refracted waves and multiples that are produced by the source. By examining the patterns of coherent noise, we can devise field procedures to reduce it. There are some sources of coherent noise:

- Multiple reflection
- Refracted events
- Diffraction events
- Ground roll
- Direct waves

Incoherent Noise

Incoherent noise displays no systematic pattern on seismogram. There are some sources of random noise, which are:

- Water flow noise
- Small movements with in the earth
- Local noise(people, Traffic etc)
- Wind noise
- Short wave length propagation noise
- Short wave length propagation noise

Noise Control

The basic tools available for controlling noise in the field include:

- Source size
- Source depth
- Electronic filtering
- Receiver arrays
- Electronic mixing

Seismic Data Processing:

Data processing is sequence of operation, which are carried out, according to a predefined program to extract useful information from a raw data set (Al-Sadi, 1980). Indeed the whole set of seismic data processing is to message, seismic data recorded in the filed into a coherent cross-section, indicating significant geological horizons into the earth subsurface, related to hydrocarbon detection and seismic stratigraphy (Hatton et al., 1986, Dobrin and Savit 1988). The purpose of data processing is to produce a perfect seismic section by applying a sequence of corrections so that it is interpretable.

Processing seismic data consists of applying a sequence of computer programs, each designed to achieve one step along the path from field tape to record section. The processing sequence consisting of corrections and adjustments which increases the signal-to-noise ration, correct the data for various physical processes, that obscure the desired geological information of seismic data and reduce the volume of data (Dobrin & Savit 1988, Yilmaz 1987, Keary & Brooks 1991, Al-Sadi 1980, OGTI Manual 1989) (see fig. 5.1).

5.1. Data Reduction:

5.1.1. Demultiplexing:

Demultiplexing in the geophysical sense is the unscrambling of multiplexed field data to trace sequential form (fig. 5.2). The demultiplexed data results in a separate trace for each shot point, sampled at what ever interval has been used during recording (often 4ms) (Badley, 1985). According to Hatton (1986), mathematically demultiplexing data is seen as transposing a big matrix so that the columns of the resulting matrix can be read as seismic trace, recorded at different offsets, with common shot point. Actually the multiplexed data is in time sequential format and demultiplexed data is in trace sequential format which is a convenient format that is used throughout processing. The principle of multiplying in field is adopted when the capacities of the AD converter are not sufficient to digitize and save all channels at the same time. This is common in older measurement systems or for measurements with a large time window and a lot of channels per shot. The separate values of all channels are sorted by samples and not by channels:

$$\boxed{T_1S_1}\boxed{T_2S_1}\boxed{T_3S_1}\cdots\cdots\boxed{T_nS_1}\boxed{T_1S_2}\boxed{T_2S_2}\cdots\cdots\boxed{T_nS_2}\cdots\cdots\boxed{T_nS_m}$$

T_i = Trace i. S_j = Sample j

It is difficult to process the data in this form. It is more convenient and illustrative when the data is sorted by traces i.e. demultiplexed:

$$\boxed{T_1S_1}\cdots\cdots\boxed{T_1S_n}\boxed{T_2S_1}\cdots\cdots\boxed{T_2S_n}\cdots\cdots\boxed{T_nS_1}\cdots\cdots\boxed{T_nS_m}$$

T_i = Trace i. S_j = Sample j

5.1.2. Vibrosies Correlation:

The signal so obtained by non impulsive source is called sweep which may be either upsweep or downsweep.Vibroseis correlation enables us to extract from each of the long overlapping sweep signals a short wavelet much like those obtained with impulsive sources (Robinson and Coruh, 1988), because all reflected and refracted signals in a vibroseis seismogram , overlap and another extensively which is not readable. Another, computer processing is needed to obtain a recognizable record. To make vibroseis record, useable, we "compress" the reflection into short wavelets. This procedure is called **"vibroseis correlation".** This is done by cross correlating the data with original input sweep (fig. 5.3) so that each reflection is compressed to a wavelet which can be used directly to examine subsurface structure (Badley, 1985).

5.1.3. Editing and Summing:

5.1.3.a Editing:

Data editing is the process of removing or correcting any trace which may cause a deterioration of the stack. Individual traces may be affected by polarity reversals or by noise. Entire records may be contaminated by coherent noise or rendered unusable by misfires or auto fires (OGTI manuals, 1988). Several forms of editing are categorized throughout data processing:

a. Editing during demultiplexing

b. Muting

Editing during demultiplexing is don't to reject the traces which contain too many sync errors or rare too short (OGTI manuals. 1988) which may be rejected either automatically or user selected.

Muting is useful to remove useless information from the processing stream in a way that first identifies the information to be removed and then blanked. Muting is categorized as *initial muting*, to remove first arrivals; usually done later in processing, and *surgical muting*, to remove air waves or ground roll energies.

5.1.3b Summing:

It is another option that is available on most demultiplex program to add several records from the same source together to produce a single record of better quality (OGTI manuals, 1989).

5.1.4c Amplitude Adjustment:

Amplitudes of the seismic wavelet is adjusted because it dies out as the input wave travels down to the earth and losses it energy due to the spatial spreading of the wave or absorption. Besides, spherical spreading and energy dissipation in earth, there are other reasons for the observable decay in seismic amplitude with time. Under the knowledge of such reasons amplitude of the seismic wavelet is adjusted:

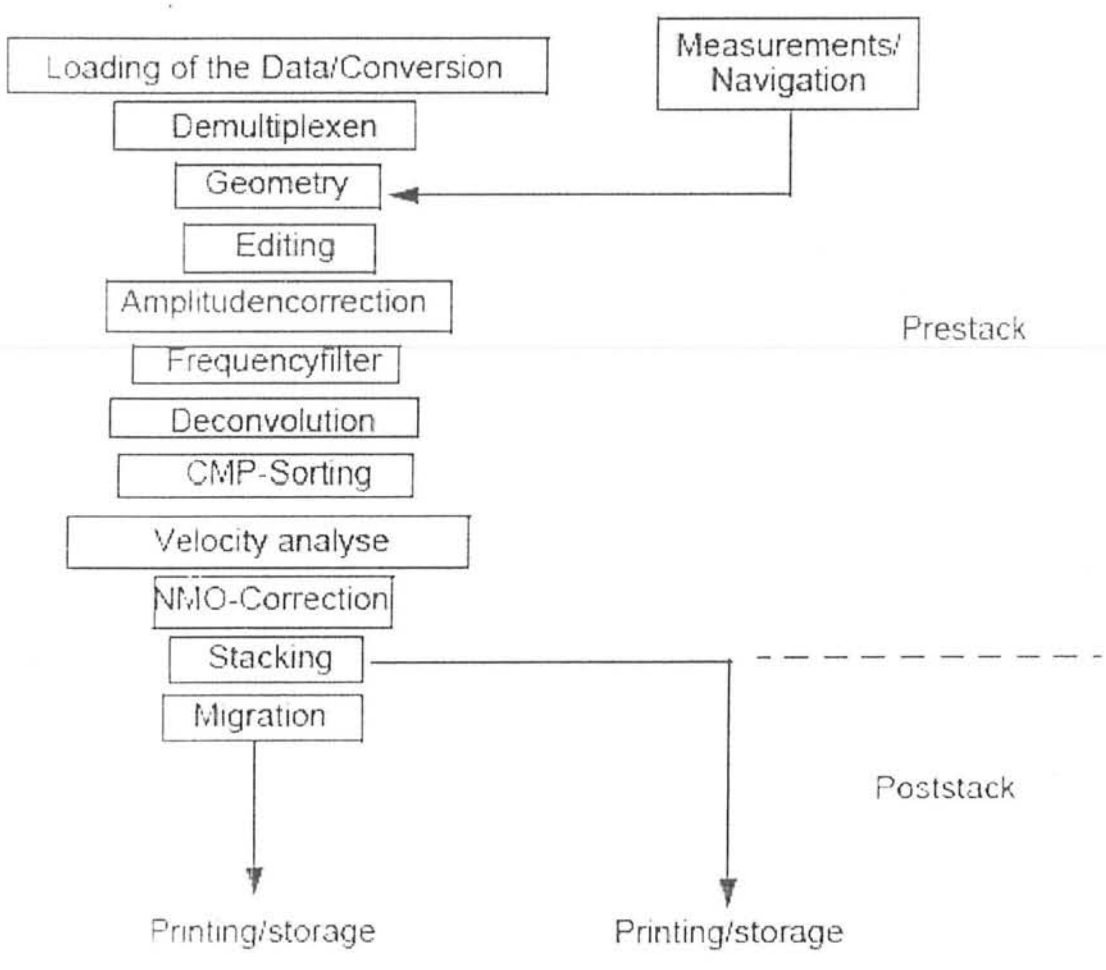
a. Trace Normalization

b. Trace Balancing

a. Trace Normalization:

Trace normalization is an amplitude adjustment applied to the entire trace. It is directly applicable to the case of a weak shot or a poor geophone plant. All absolute values of a trace are summed and compared with a reference value. A scaling factor is determined from the difference between the summation and the reference value, which is used to multiply all data with. Other possibilities of trace normalization could be *Average value* (arithmetic or RMS), Median, Maximum Value to compensate the difference in amplitude which occurs due to the increasing distance between the source and receiver and the lateral differences in amplitudes. But the loss of amplitude with increasing depth is not taken into account.

Fig. 5.1 Detailed processing sequence flow chart.



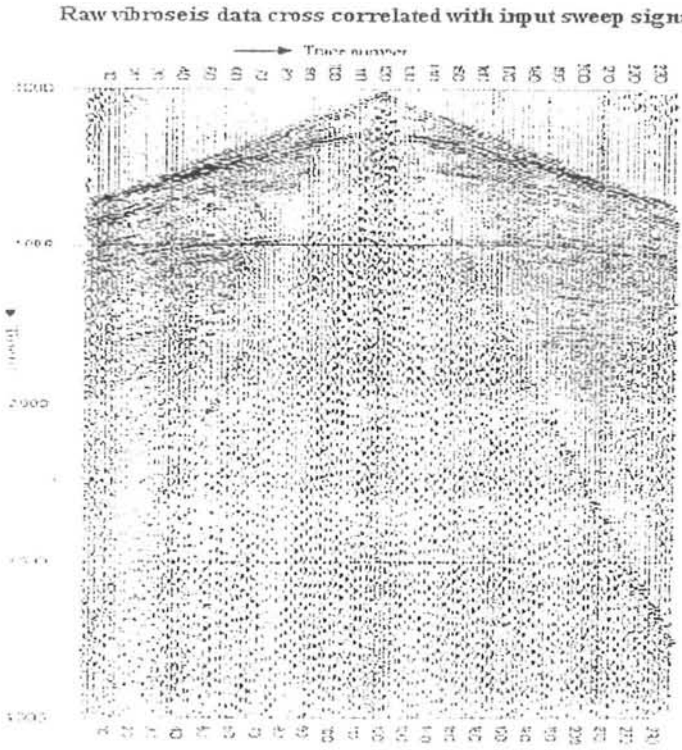


Fig. 5.3 Raw shotgather from Tubbergen data set, cross-correlated with the sweep (Maurice and Sercel, 1997).

b. Trace Balancing-AGC:

The AGC function does not employ a gain to the whole trace, but employs a gain to a certain time sample within a time gate. First, the mean absolute value of trace amplitudes is computed within a specified time gate. Second, the ratio of the desired *rms* level to this mean value is assigned as the value of the gain function. This gain function is then applied to any desired time sample within the time gate; say the n^{th} sample of the trace. The next step is to move the time gate one sample down the trace and compute the value of the gain function for the $(n+1)^{th}$ time sample and so on (fig. 5.4). The time gate is very important. Very small time gates can cause a significant loss of signal character by boosting zones that contain small amplitudes. In the other extreme, if a large time gate is selected, then the effectiveness of the AGC process is lessened. 256- to 1024-ms AGC time gates are commonly chosen.

A disadvantage is that when the AGC gain is applied, it is not possible to reconstruct the original signal again. Therefore, the AGC is only used for display and printing purposes.

5.1.5. Display:

The data so processed is generally displayed in various modes (fig. 5.5) to summarize the information gathered. At any point of processing sequence the seismic analyst can display the data in wiggle trace or other modes. The choice of display is a matter of the client taste, but is not affected by company dictum. Currently, the data provided by OGDCL is the variable area with wiggles plot.

5.2. Geometric Correction

5.2.1. Trace Gathering (Cdp Sort):

There are different possibilities to sort the data:

- Common shot - all traces that belong to the same shot
- Common midpoint (CMP) - all traces with the same midpoint
- Common receiver - all traces, recorded with the same geophone
- Common offset - all traces with the same offset between shot and geophone

For a horizontal layered earth the reflection point lies between source and receiver (midpoint).Using more shots with different positions of the source and receivers several combinations of source and receivers exist which have the same midpoint. When a horizontal layering is present the reflection then also comes from an equal point in the subsurface (Common depth point- CDP). For an inclined layer the point of reflection for traces with the same midpoint are not equal anymore. The nomenclature CDP is not valid anymore. However, several processing programs still use the word CDP in stead of CMP.

Zero offset:

Zero offset data is characterized when the source and receiver are present on the same location. For a normal measurement this is seldom the case. When the traces are corrected for the move out and are stacked then a zero offset trace is obtained.

Common offset:

All traces with equal offset between source and receiver. This configuration is often used for several single channel systems. Also Georadar measurements are often carried out with a fixed offset between source and receiver.

Fold:

The fold indicates the number of traces per CDP. This is often the number of traces in a CMP.

The theoretical formula for the fold is given by:

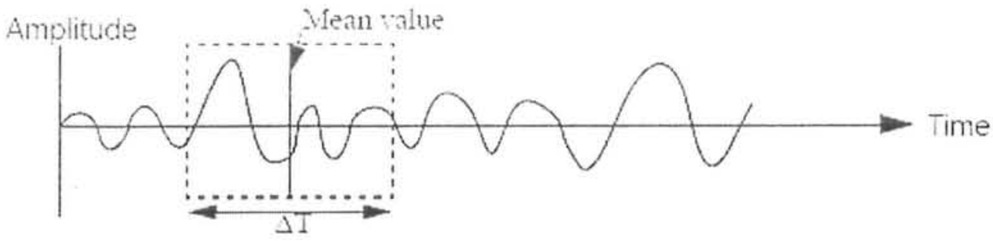
$$\text{Fold} = \frac{\text{Number of Geophones} \cdot \text{Distance between Geophones}}{2 \cdot \text{Distance between shots}}$$

The number of traces which are measured at a certain geophone position is called **"surface fold"**.

5.2.2. Static Correction:

According to Maurice and Sercel (1997), Static Correction is a time-correction, which is constant for an entire trace and meant for the compensation of weathering and elevation time offset. Two main sources of irregularities are:

a. Elevation difference between individual shots and detectors.
b. The presence of a weathered layer which is a heterogeneous surface layer, a few meters to several ten meters thick, of abnormally low seismic velocity which causes a disproportionallity (fig. 5.7) and variable time-delay in the arrival of the desired deeper reflectors. They affect reflection continuity, resolution, the accuracy of velocity analysis and structural form.



Principle of AGC

Fig 5.4 AGC window showing how it works

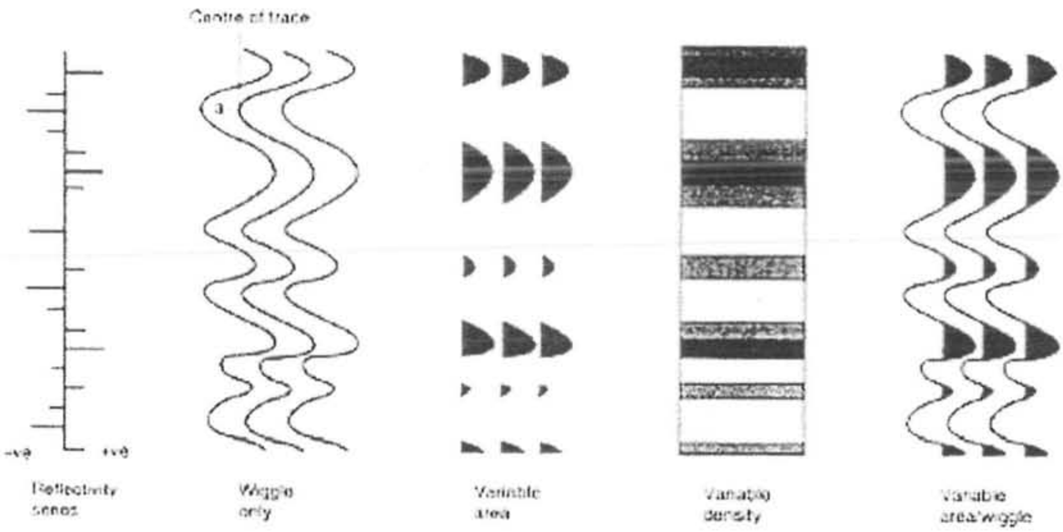


Fig. 5.5 Common display types for seismic trace.

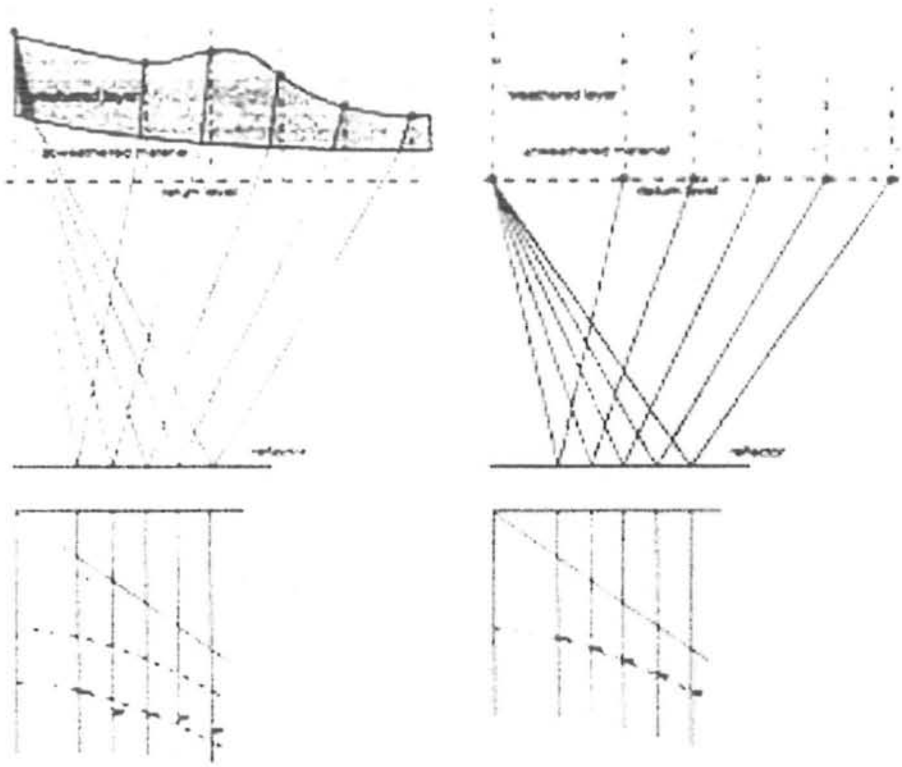


Fig. 5.7 Effect of topography showing that reflections are not perfect hyperbola after static correction the reflections appears as if source and receivers had been positioned to datum level

Aim of static corrections is to adjust the seismic traces in such a way that the sources and receivers are present at one horizontal level. To achieve this, the travel times of the separate traces are corrected. The whole trace is corrected with the same time shift.

It is better to apply it before stack because if there is a low velocity weathering layer delaying the far offset trace and not the near offset traces. then the NMO correction will not accurately align the reflection. The correction is calculated on the assumption that the reflection ray path is effectively vertical immediately beneath any shot or detector. Calculation of static correction requires knowledge of the velocity and thickness of the weathered layer. To determine it, up-hole and refraction information are used. Generally, the elevation correction is done at the same time as weathering correction (Dobrin and Savit 1985, Kearey and Brooks, 2002).

5.2.3. Dynamic Correction:

The principle of dynamic correction is explained in fig. 5.8. The first step in the CDP gather is the normal moveout adjustment (Robinson and Coruh, 1988).For a single constant-velocity horizontal layer, the travel time curve as a function of offset is a hyperbola (fig. 5.9 a). The time difference between travel time at a given offset and at zero offset is called **Normal Moveout (NMO) or dynamic correction.**

The velocity required to correct for normal moveout is called the normal moveout velocity (Yilmaz, 1987).The travel time curve of the reflections for different offset between source and receiver is calculated by using following relationship (Yilmaz, 1987). By using,

$$t^2 = t_0^2 - \frac{x^2}{v_{stack}^2}$$

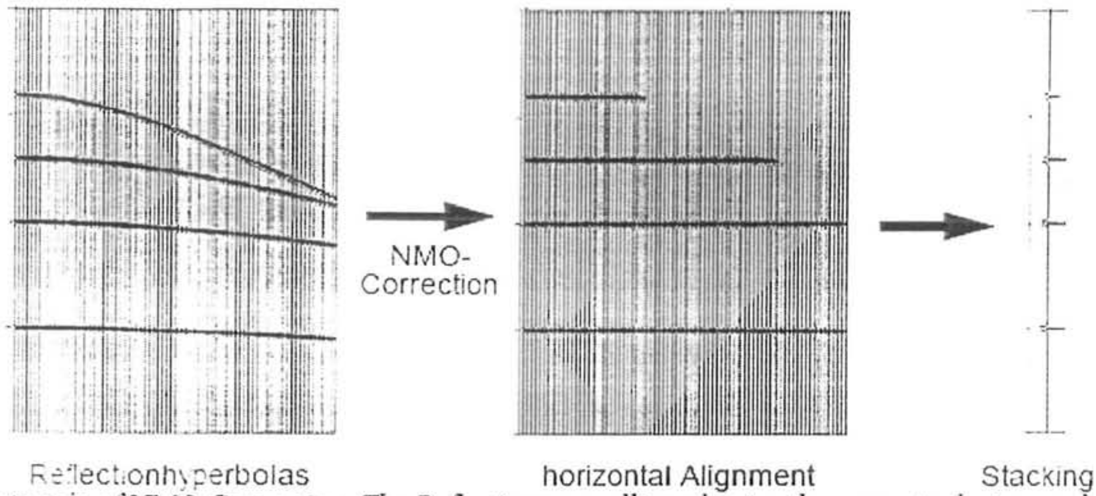
From this formula the NMO-correction can be derived and is given by:

$$\Delta t = t_0 - t(x) \qquad \text{with} \qquad t(x) = \sqrt{t_0^2 + \frac{x^2}{v_{stack}^2}}$$

Where,
V^2_{stack}= stacking velocity computed by velocity analysis.

The Moveout Δt is the difference in travel time for a receiver at a distance x from the source and the travel time t_o for **zero-offset** distance. The NMO-Correction depends on the offset and the velocity. In contrast to the static correction, the correction along the trace can differ.

Principle:



Principle of NMO-Correction. The Reflections are alligned using the correct velocity, such that the events are horizontally. Then all the separate traces are stacked (summed).

Fig. 5.8 Principle of Dynamic Correction

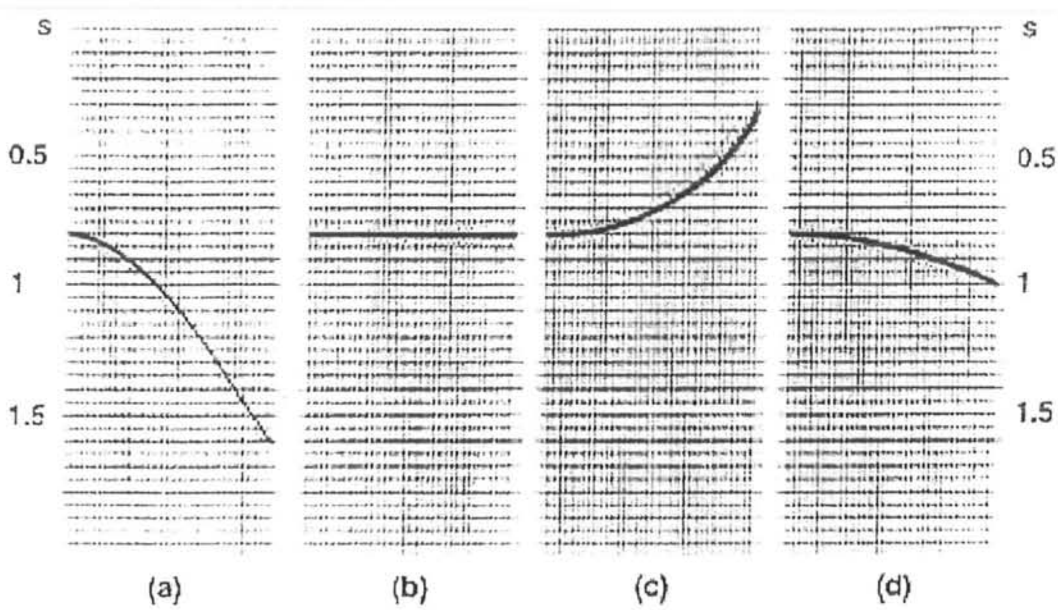


Fig. 5.9 NMO correction (a) not corrected (b) corrected for proper velocity (c) velocity is too high (d) velocity is too low.

5.3. Data Analysis And Parameter Optimizations:

5.3.1. Filters:

Filtering is an operation by which the amplitude and/or phase spectrum of a time signal are altered. Purpose of a filter is to enhance primary reflections by attenuating ambient and source-generated noise whose frequency spectra are separated in tow ways are identical mathematically (OGTI manuals, 1989). In general the aim of filtering is to improve signal to noise ratio (Kearey and Brooks, 2002). There are two basic methods of filtering:

a. Frequency Domain Method

b. Time Domain Method

Other special designed filters are based on one parameter as a basis of filtering e.g. velocity filters, fk-filters, and τ-p filters, dip filters etc (Kearey and Brooks 2002; Yilmaz 1987). The typical application of these filters is given in table 5.1. For various types of filters see fig. 5.10.

Table 5.1

Type of filter	Application
Fk –filters	Suppression of noise signals with specific slopes (Interface waves), multiple reflections and Elimination of Artifacts in stacked Sections (post-stack)
Alias filter	Removes those frequencies which are above Nyquist frequency.
Band pass Filters	Passes on specified band of frequency
Low pass and high pass filters	Passes specific frequency either low or high
Notch Filters	To analog before digitizing, removes specified narrow notch frequency e.g. of power lines 60 Hz.
Tau Filters	time-dependent velocity filter; Suppression of multiples, interpolation between traces, analysis of Guided Waves
Velocity Filters/fan filters/pie slice filtering	Remove coherent noise event on the basis of particular angle at which the event dips (March & Bailey 1983).

5.3.2. Deconvolution or Inverse Filter:

Deconvolution is a process that improves the temporal resolution of seismic data by compressing the basic seismic data by compressing the basic seismic wavelet (Yilmaz, 1987). The aim of deconvolution is reconstruction of the reflectivity function theoretically and practically, shorting of signal. suppression of noise and multiples. From fig. 5.11 after Yilmaz (1987) is divided into two sections before and after deconvolution showing how deconvolution compress and resolves the input signal so that reflections are prominent after the suppression of noises and multiples between 1 and 2 seconds. According to Dobrin and Savit, 1988, inverse filter is designed to deconvolve seismic traces by removing the adverse filtering effects associated with the propagation of the seismic pulse through the layered ground or through a recording system. They only reveal those reflections that stem from real reflector. Two main types of deconvolution are:

a. Deterministic Deconvolution

b. Predictive Deconvolution

Deterministic deconvolution is capable of producing a pulse of any designed shape with an appropriate band width.

Predictive deconvolution attempts to predict event shapes, obtained by statistical studies of the seismic traces (Badley, 1985).

5.3.3. Velocity Analysis:

The aim of the velocity analysis is to find the velocity that flattens a reflection hyperbola, which returns the best result when stacking is applied. This velocity is not always the real RMS velocity. Therefore, a distinction is made between:

• v_{stack}: the velocity that returns the best stacking result.

• v_{rms}: the actual RMS-velocity of a layer.

For a horizontal layer and small offsets, both velocities are similar. When the reflectors are dipping then v_{stack} is not equal to the actual velocity, but equal to the velocity that results in a similar reflection hyperbola.

There are different ways to determine the velocity:

a. (t^2-x^2)-Analysis.
b. Constant velocity panels (CVP).
c. Constant velocity stacks (CVS).
d. Analysis of velocity spectra.

For all methods, selected CMP gathers are used. Example of one method is given in fig.5.12

CVP - "Constant velocity panels"

The NMO-correction is applied for a CMP using different constant velocities. The results of the different velocities are compared and the velocity that results in a flattening of the hyperbolas is the velocity for a certain reflector.

CVS - "Constant velocity stacks"

Similar to the CVP-method the data is NMO-corrected. This is carried out for several CMP gathers and the NMO-corrected data is stacked and displayed as a panel for each different stacking velocity. Stacking velocities are picked directly from the constant velocity stack panel by choosing the velocity that yields the best stack response at a selected event. CVP and CVS both have the disadvantage that the velocity is approximated as good as the distance between two test velocities. Both methods can be used for quality control and for analysis of noisy data.

Velocity-Spectrum:

The velocity spectrum is obtained when the stacking results for a range of velocities are plotted in a panel for each velocity side by side on a plane of velocity versus two-way travel-time. This can be plotted as traces or as iso-amplitudes. This method is commonly used by interactive software to determine the velocities.

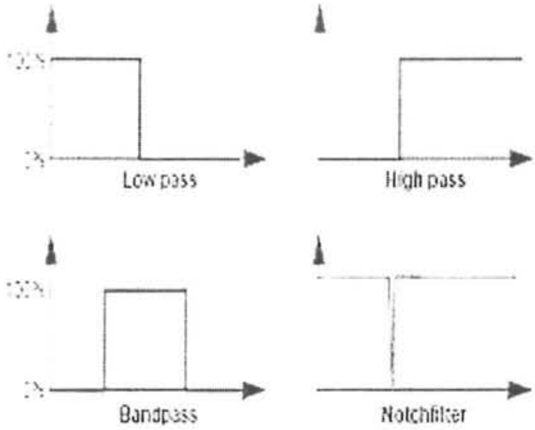


Fig. 5.10 Example of various types of filters

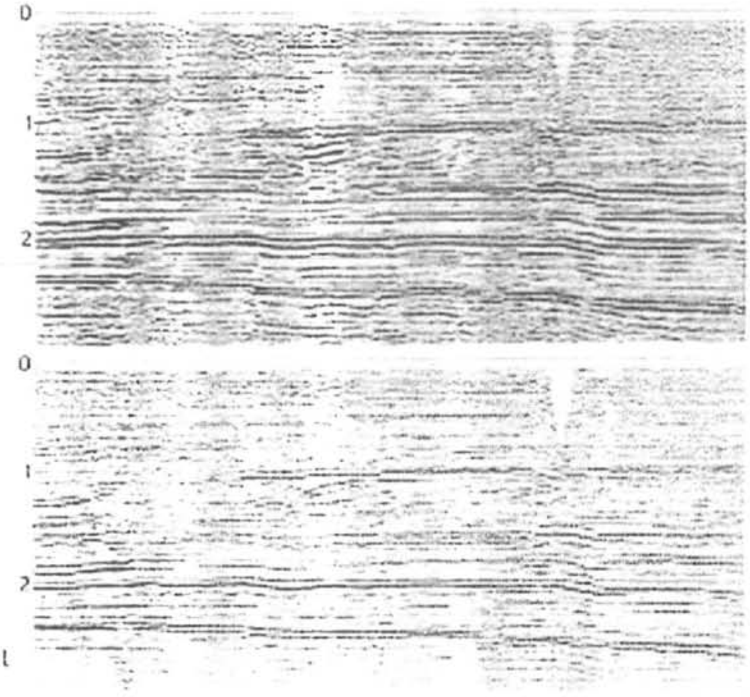


Fig 5.11 An example of deconvolution shows how reflected signals are improved and resolved from those viewed above and as a result of deconvolution below.

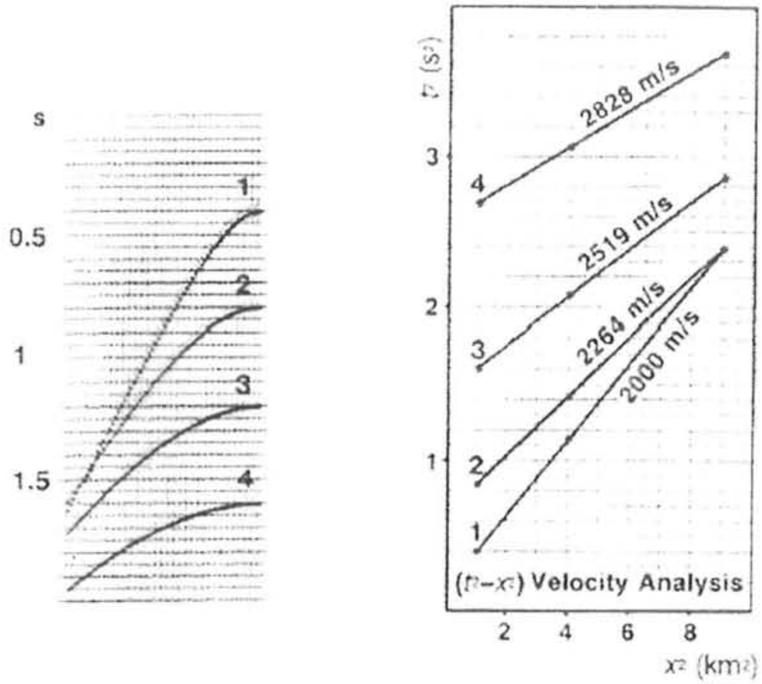


Fig 5.13 Effect of migration moves the reflector back to its true position.

5.4. Data Refinement

5.4.1 Stacking:

Stacking is simply a method to improve signal-to-noise ratio, by adding reflections together in phase and adding noise, out of phase, so that it cancels. According to Badley 1985, stacking discriminates against multiples provided that the velocity of the multiples is such that during stacking, it has a normal-move-out significantly different from primaries arriving at the same time (Badley, 1985). Stacking is performed by summation of the NMO-corrected data. The result is an approximation of a zero-offset section, where the reflections come from below the CMP position.

Several methods can be used to combine the different NMO-corrected traces. The most important are:

noise. This type of stacking is often used to suppress multiples by weighting the large-offset data more heavily than the short-offset traces. because the difference in NMO between primaries and multiples is larger for larger offsets. A weight factor α is introduced.

5.4.2 Residual statics:

The process of residual statics consists of shifting the separate traces in such a way that the optimal reflections are obtained. To make sure that the traces of a single CMP are not shifted randomly, the shift is divided in a value for the source ("source static") and a value for the receiver ("receiver static"). For each source and receiver a value is determined. All traces with a certain source are corrected with the value for that source. Similarly all traces with a certain receiver are corrected with the value for that receiver. The resulting shift (static correction) of a trace consists of the correction value of the source and receiver of the corresponding trace. This processing still assumes that the static shifts are caused by the interface. Therefore, this processing is also called **surface consistent static correction**.

5.4.4. Migration:

Migration may be defined as a process which corrects the distortions of the geological structure inherent in the seismic section (Maurice and Sercel, 1997). As shown in fig 5.13, it maps the image back to its true location.

a. Wavefront Common Envelope Migration:

The dipping reflector is according to the reflection laws; perpendicular to the reflection rays is model of source-receiver common position. This means that the reflector is defined by the tangent-surface (or common envelop) to all the wavefronts drawn for all the incident seismic rays. Hence migration processing is carried out by placing each trace sample on all points of the corresponding wavefront (Al-Sadi, 1980).

b. Diffraction Hyperbola Migration:

This method is based on the assumption that the reflector is made up of a packed series of diffraction points or scatters. According to this model a reflection arrival may be considered as a resultant signal from the interference (constructive and destructive) processes which take place along the diffracted waves from these points (Al-Sadi, 1980).Adopting the following processing sequence, the processed seismic section is obtained.

Processing Sequence:

1. **Editing / Demultiplexing**
   - Processing Sample Interval 2 msec
2. **Preprocessor**
   - Geometry definition
   - Field statics
   - Bad traces are deleted
3. **Datum Statics Correction**
   - Datum plane 150 amsl
   - Datum velocity 1600 m/sec
4. **Geometrical spreading**
5. **Notch filter (50 hz)**
6. **Zone anomaly process (zap)**
7. **Fk dip filter**
8. **Surface Consistent Deconvolution**
   - Surface consistent
   - Operator length 200 msec
   - Prediction distance 2 msec
   - Trace balance
9. **Band pass fileter (12/24-80/48 hz/db)**
10. **Velocity analysis (cvs)**
11. **Normal moveout**
12. **Regional statics (miser)**
13. **Mute**
14. **Stack**
15. **Spectral whitening**
16. **Finite diference migration**

17. **Random noise attenuation**

18. **Shallow averageing (500 msec window)**

19. **Band pass filter**

Time	Low Cut	High Cut
(sec)	(hz/db)	(hz/db)
0.00	12/13	65/38
5.00	12/18	65/36

20. **Peak Gain**

amplitude 2000 ms

21. **Scales**

Horizontal : 80 traces = 1kilometer

Vertical : 10 centimeter per second

22. **Deplay Parameters**

- Display amplitude : 12
- Dipolarity : normal

6.1. Introduction

Velocity is defined as the rate of increase of distance traversed by body or wave in a particular direction. Seismic Velocities are the most important parameter in Seismic technique for interpretation and processing. We routinely use velocities to stack seismic data, to migrate seismic data, and to convert time-recorded seismic sections to depth sections and time maps to depth maps. We also use velocity in more sophisticated ways, such as in attempts to predict porosity, geological age, lithology, fracturing, fluid content and geo pressure.

Seismic velocities vary largely in sedimentary rocks as compared to igneous and metamorphic rocks. Metamorphic and igneous rocks have little or no porosity, and, the seismic wave velocity depend upon the elastic properties of the material making up the rock material itself. In terms of lithology, whenever there is a change in grain size and mineralogical composition of the rock, velocity behavior changes. An increase in grain size will result in the increase in velocity. In many areas, seismic velocity data can be used to identify lithology in discrete formations within the geologic section (Dobrin, 1976). Velocity, as a seismic parameter plays an important role in seismic processing. The accuracy of data reduction, processing and interpretation of seismic data depend mainly on the correctness of velocity measurements. Borehole velocity measurements offer more precision than surface based measurements (Al-Sadi 1980).

Whenever energy is produced by the source, it generates "P" and "S" waves, which enter into the earth. Commonly, the P-wave velocity is considered to be the most important parameter in seismic data processing and interpretation. Together with density velocity variation across an interface decides the efficiency of reflection or transmission at interfaces.

6.2. Factors Affecting Seismic Velocities

The seismic velocities in rocks are affected by several factors. These are listed below:

- Density and Elasticity of the rock
- Porosity of rock

- Geological structure
- Overburden pressure or depth of burial
- Lithology or mineralogical composition of the rock
- Fluid content in the pore
- Age of rock

Density And Elasticity Of Rock

The relationship of density and elasticity with the seismic velocity is given as:

$$(\text{Velocity})^2 = \text{Effective elasticity} / \text{Density}$$

From this relation, the velocity is directly proportional to elasticity and inversely proportional to density. So it is expected that the denser rocks would have low velocity, however the reserve is true in nature. The reason is that as the material becomes more compact its elasticity increases in such a way that it reduces the effect introduced by increased density.

Porosity Of Rock

In the case of porosity higher will be the velocity and vice versa. The relationship between porosity (□) and velocity (V) is given by the following expression,

$$1/V = \square / V_f + (1+\square) / V_m$$

Where,

V_f = Velocity of the Pore Fluid

V_m = Velocity in Rock Matrix

V = Velocity in saturated rock

□ = Fractional porosity

Geological Structure

Discussing the effect of geological structure, it is noticed that in an isotropic medium, the recorded velocity is generally high, when measured along the strike of structure. This difference may be of the order of 5-15%.

Depth of Burial

A quantitative relationship between velocity, depth and age of the rock for the shale and sandstone section, which is given as:

$$V=K\,(ZT)^{1/6}$$

Where.

V = Velocity in feet per second

Z = Depth in feet

T = Age in years

K = Constant

Lithological And Mineralogical Composition Of Rock

Lithological and mineral composition of rocks also affects velocity of seismic waves. As described earlier, that average velocities for igneous rocks is higher than that for other types and they show a narrower range of variation than sedimentary and metamorphic rocks. Mineralogy of rock surely causes variations in seismic velocities. For example the velocity of a same wave will be different for sands, silts and clays etc.

Fluid Content In The Pores

In actual rocks the pores spaces are filled with a fluid. This fluid may be in form of air or any other gas or different liquid solutions. These fluids can affect the seismic velocities. Seismic velocities will be low in case of gases and different solutions present in the pore spaces of rock.

Age of Rock

An older rock might be expected to have a higher velocity, have been subjected for a long time to pressures, cementation, and other factors, which might increase its velocity.

6.3. Seismic Velocities of Earth Materials

The P and S wave velocities of various earth materials are shown below:

Material	P wave Velocity (m/s)	S wave Velocity (m/s)
Air	332	
Water	1400-1500	
Petroleum	1300-1400	
Steel	6100	3500
Concrete	3600	2000
Granite	5500-5900	2800-3000
Basalt	6400	3200
Sandstone	1400-4300	700-2800
Limestone	5900-6100	2800-3000
Sand (Unsaturated)	200-1000	80-400
Sand (Saturated	800-2200	320-880
Clay	1000-2500	400-1000

6.4 Variations in seismic velocities

There are two types of variations in seismic velocities;

1. Lateral variation in seismic velocity
2. Vertical variation in seismic velocity

Lateral variations in seismic velocities

These variations are supposed because of slow changes in density and elastic properties due to changes in lithology or physical properties. Lateral variations make events appear to move up or down on time sections (Robinson & Coruh, 1988).

Vertical variations in seismic velocities

These variations are due to lithological changes of layering and increasing pressure due to increasing depth. Normally seismic velocities increase with the increase in depth (Robinson & Coruh, 1988). Vertical variation in velocity cause differences in the two way travel times of layers of equal thickness

6.5 Types of velocities used in seismic exploration

The different types of velocities used in seismic exploration are;

6.5.1 Average Velocity

6.5.2 Interval Velocity

6.5.3 Root-Mean-Square Velocity

6.5.4 Normal-Move out Velocity

6.5.5 Instantaneous velocity

6.5.1 Average Velocity

Average velocity is simply the total distance traveled divided by the total time traveled. The average seismic velocity is the distance traveled by a seismic wave from the source location to some point on or within the earth divided by the recorded travel time (Al. Sadi, 1980). The Figure 22 shows a two-layer case by which the average velocity is calculated.

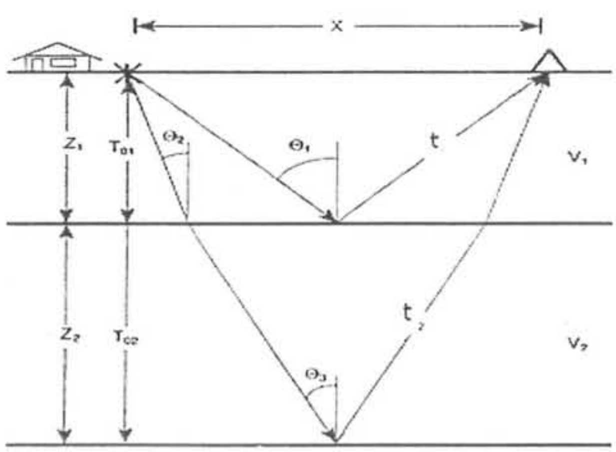


Figure 22: A two-layer case by which the average velocity is calculated (Dobrin, 1988)

If concerned with the distance and time from the surface of the earth to a point at depth, then the one-way distance and time is used (Dobrin, 1988). The average velocity in this case is simply:

$$V_a = \frac{\sum_{i=1}^{n} z_i}{\sum_{i=1}^{n} t_i} = \frac{\sum_{i=1}^{n} v_i t_i}{\sum_{i=1}^{n} t_i}$$

Nevertheless, if considering with the distance from the surface of the earth to a point at depth and back to the surface, then two-way distance and travel time is used, and average velocity equals 2Z/T (Dobrin, 1988). So, average velocity can be expressed as

$$V_a = \frac{z}{t} = \frac{2z}{2t} = \frac{2z}{T}$$

Where, t is one-way travel time, and T is the two-way travel time.

6.5.2 Interval Velocity

Interval velocity, V, is defined as the thickness of a particular layer divided by the time it takes to travel from the top of the layer to its base (Dobrin, 1988). The interval velocity is ΔZ (the thickness of a stratigraphic layer) divided by Δt (the time it takes to travel from the top of the layer to its base). The equation for interval velocity is:

$$V_i = \frac{Z_m - Z_n}{t_m - t_n} = \frac{Z_m - Z_n}{T}$$

The thickness $\Delta Z = Z_m - Z_n$ is also equal to the isopach value of the interval. A typical interval-velocity-versus-time curve compared to the average velocity is shown in Figure 23. The discrete boundaries in the interval-velocity curve indicate stratigraphic and velocity differences between two contiguous layers (Yilmaz, 2001). The average velocity can be determined by averaging the weighted summation of the interval velocities.

If we sum the interval velocities for a series of rock layers, and weight them according to the two-way travel time within each layer, ΔT. the average value would be equal to the average velocity (Yilmaz, 2001). The equation for average velocity, V_A, in terms of interval velocity is:

$$V_a = \frac{\sum V_i \Delta T}{\sum \Delta T} = \frac{2\sum \Delta Z}{\sum \Delta T}$$

Where ΔZ is the interval thickness or isopach thickness.

6.5.3 Root-Mean-Square (RMS) Velocity

The root-mean-square (RMS) velocity is a weighted average. It is used as weighting process where the amount of weighting is determined by the value of the interval velocities. The weighting is accomplished by squaring the interval velocity values (Robinson & Coruh, 1988). So, in this approach, greater weight is given to the greater interval velocities. The equation for RMS velocity is given below

$$v_{rms}^2 = \frac{\sum_{i=1}^{n} v_i^2 t_i}{\sum_{i=1}^{n} t_i}$$

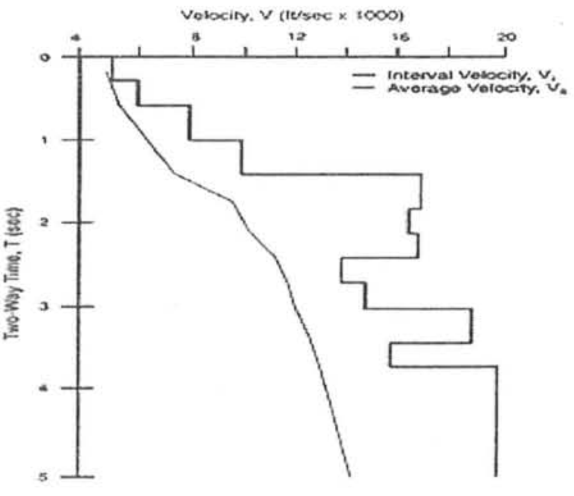


Figure 23: A typical interval-velocity-versus-time curve compared to the average velocity (Yilmaz, 2001)

By comparing the equations it is clear that the RMS velocity is always greater than the average velocity. RMS velocity is strictly a mathematical weighted average and has no intrinsic meaning (Robinson & Coruh, 1988). Figure 24 shows a graphical comparison between the root mean square velocity and the average velocity.

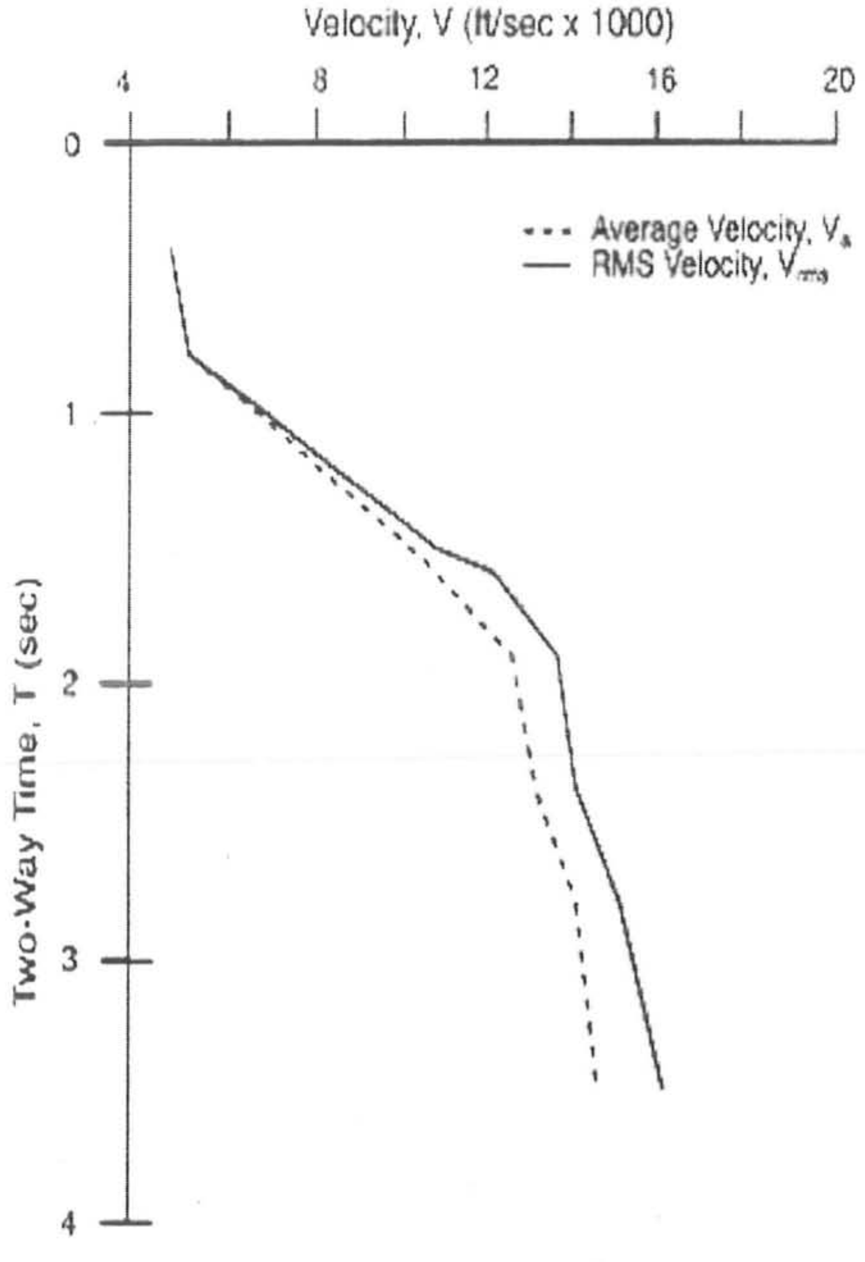


Figure 24: A graphical comparison between the root mean square velocity and the average velocity (Yilmaz, 2001)

6.5.4 Normal-Move out Velocity

The normal-move out (NMO) velocity, or stacking velocity, V_{nmo}, has a horizontal component (X). Therefore, it is dependent on the offset, depth, and spread length. Seismic records with source-to-receiver distances will yield different NMO velocity values. The NMO velocity increases as the value of X increases (Rehman, 1989). The equation for NMO velocity is:

$$V_{nmo} = \frac{X}{\sqrt{T_x^2 - T_x^2}} = \frac{X}{\sqrt{2T_0 \Delta T_{nmo}}}$$

Where, X = the offset distance from source to receiver, T_x = the two-way travel time of a seismic wave reflected off a particular interface and recorded at the receiver location, and T_0 = the two-way travel time of the seismic wave reflected off the particular interface at the zero-offset location.

We can calculate the NMO correction, ΔT_{nmo}, from the average velocity (by Figure 22). From the Pythagorean Theorem, we know that

$$d^2 = \Delta z^2 + \left(\frac{X}{2}\right)^2$$

If we know the normal-movement velocity, it can be related to the average velocity using the equation $T = T_0 + \Delta T_{nmo}$ (Telford, 2004). Then approximate the NMO correction as a function of average velocity:

$$T_0 + \Delta T_{nmo} \approx \sqrt{T_0^2 + \frac{X^2}{V_a^2}}$$

$$\Delta T_{nmo} \approx \sqrt{T_0^2 + \frac{X^2}{V_a^2}} - T_0$$

The NMO correction can also be approximated from the RMS velocity. In this case,

$$\Delta T_{nmo} \approx \sqrt{T_0^2 + \frac{X^2}{V_{rms}^2}} - T_0$$

$$\Delta T_{nmo} \approx \frac{X^2}{2T_0 V_{rms}^2}$$

6.5.5 Instantaneous velocity

If the velocity varies continuously with depth, its value at a particular depth Z is obtained from interval velocity by contracting the interval Z1-Z2 to an infinitesimally thin layer having a thickness dZ (Telford, 2004). The interval velocity then becomes the derivative of Z with respect to "t", which is the instantaneous velocity, defined as follows:

$$V_{inst} = \frac{dz}{dt}$$

6.6 Correlation between velocity types

In seismic prospecting we are dealing with a medium which is made up of a sequence of layers of different velocities. In dealing with this kind of situation, it is necessary to specify the kind of velocity we are using. When velocity is measured for a defined depth interval, it is called as interval velocity and when it is determined for several layers it is called as average velocity (Al-Sadi, 1980).

Relationship between interval velocity, root mean square velocity and average velocity is given by "Dix Formula" (Al-Sadi, 1980). If root mean square velocities (V_{rms}) is given then interval velocities (V_{int}) can be determine by using the following form of Dix formula.

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

If, on the other hand ,if given the average velocity (V_a), interval velocity(V_{int}) can be determined by another form of Dix- formula (Al-Sadi,1980).

$$V_{int} = \frac{V_{a,n} * t_n - V_{a,n-1} * t_{n-1}}{t_n - t_{n-1}}$$

Now, if we are given with interval velocities (V_{int}) and we have to determine average velocities (V_a), (Al-Sadi, 1980), then Dix formula attains the form as given below

$$V_{a,n} = \frac{(V_{int,n} * T_n - T_{n-1}) + (V_{a,n-1} * T_{n-1})}{T_n}$$

So if we are given with any of the interval, root mean square velocity or average velocity, the remaining two by using the corresponding form of Dix-formula.

INTERPRETATION

7.1. Introduction

Interpretation is the transformation of seismic data into structural and stratigraphic picture through a series of different steps. Thus threading together all the available geological and geophysical information including the seismic and then integrating them all in a single picture can only give a picture closer to the reality.

The main purpose of seismic reflection survey is to reveal as clearly as possible, the structures and stratigraphy of the subsurface. The geological meanings of seismic reflection are simply indications of different boundaries where there is a change in acoustic impedance. These observed contrasts are associated with different geological structures are stratigraphic contacts.

To distinguish different formations by means of seismic reflection is an important question in interpreting seismic reflection data. For this purpose the data is correlated with the well data and geology of the area under observation, which is already known (previous literature). The well data provides links between lithology and seismic reflections. The reflector identification is the next stage by which the actual interpretation starts and it establishes a stratigraphic frame block for the main interpretation.

Extracting from seismic data the geological structures, such as foldings and faultings are referred as structural interpretation (Dobrin & Savit 1988). On the other hand, extracting non-structural information from seismic data is called, "Seismic Facies Analysis".

There are two main approaches for the interpretation of seismic section:

- Stratigraphic Analysis
- Structural Analysis

Stratigraphic Analysis

Stratigraphic analysis involves the subdivision of seismic sections into sequences of reflections that are interpreted as the seismic expression of genetically related sedimentary sequences. Unconformities can be mapped from the divergence pattern of reflections on a seismic section. The presence of unconformable contacts on a seismic section provides important information about the depositional and erosional history of the area and on the environment existing during the time, when the movements took place.

The success of seismic reflection method in finding stratigraphic traps varies with the type of trap involved. Most such entrapment features are reefs, unconformity, disconformity, facies changes, pinch-outs and other erosional truncations. Some of the parameters used in seismic stratigraphic interpretation are

- Reflection configuration
- Reflection Continuity
- Reflection Amplitude
- Reflection Frequency
- Interval Velocity
- External Form

Structural Analysis

It is the study of reflector geometry on the basis of reflection time. In structural analysis, the main objective is to search out traps containing hydrocarbons. The most common structural features associated with the oil, are anticlines and faults. In Qadirpur area, faults associated with the extensional regime, resulted in the formation of normal faults and series of horst and graben structures (more detailed exaggeration is presented in chapter 2,of geology of the area).

7.2. Seismic Time Section

A time section is actually a reproduction of an interpreted seismic section. It consists of two scales; horizontal scale consists of SPs while the vertical scale consists of two-way time in seconds. The time for each reflector is marked from the seismic section and plotted against the short points. Since the total time acquired for the section is 5 seconds but the reflectors tend to be only at a depth of 2.4 seconds approximately, therefore the depth taken for the display of the time section and the depth section is 2.4 second rather than 5 seconds. Each reflector depth according to time is read and interpreted in terms of velocity and depth. The total CDP points of the seismic line 985-QPR-08 were from SP-210 to SP-690, out of which SP-400 to SP-700 are interpreted. There were 33 seismic velocity windows provided in the seismic section; only thirteen (S.P. # 400 to S.P. # 700.

7.3. Steps in Interpretation

Marking of horizon

The first step when starting the interpretation process is to judge the reflections and unconformities, if present, on the seismic time section. Those reflectors are selected which are real, show good character and continuity, and can be followed throughout the area (Badely, 1985).

On the time section horizons are marked by picking the continuous train of wavelets running across the section. Confusion arise in marking the continuity because the wavelets or the traces tend to mix up or the sequence might break due to subsurface structural changes or abrupt lithological changes or the most common problem faced is the presence of different types of noises, such noises causes the distortion of the signal. Therefore, in order to decide that whether the sequence continues towards the upper horizon or the lower, a broader view of the interpreter, knowledge about the area and the considerable experience would help in marking a correct pick.

Seven horizons were picked on Line # 985-QPR-04 of Qadirpur area. These horizons were marked through the same steps and not much of difficulty was faced because of simple structure the area. The reflectors were strong enough to be picked due to variation in acoustic impedance that is eventually caused by changes in lithology.

Construction of Iso-velocity Contour Map

Calculate the Dix Average Velocity and these varies from1500m/s to 5000m/s for 5 sec data and have been used to construct a velocity contour map at an interval of 200m/s. This contour map not only gives the vertical variation of velocity, but also the lateral variation in the velocities. The vertical variation is due to the lithological changes, depth burial, porosity, temperature, pressure, and density. When lateral variations in velocity overlying our objective level causes a distortion in the time structure to be significantly different from the real.

- To calculate accurate reservoir volumes for reserve estimates.
- In preparing the well prognosis – when it is necessary to know the anticipated depths of interpreted reflections etc.
- Depth information can be crucial for the locations of casing points, coring or knowing at what depth we are likely to encounter drilling hazards, which may have been identified from regional knowledge or, in favorable circumstances, directly from the seismic.

Depth Section

Generally the depth section gives the configuration of reflectors in the same way as the time section. Remove the kinks for the horizons as much as possible in order to obtain the smoother horizons. To determine the depth of the marked reflectors on the seismic section, the formula employed is:

$$S = V * T/2$$

Where,

S = Depth of the reflector

V = Average velocity

T = Two-way time of the reflector read from the seismic section

7.4. Estimation of horizon depth

An accurate measurement of the seismic velocities is an important step in the seismic interpretation and processing. Two different methods were adopted in order to construct the depth section:

- Mean Average Velocity Line Method
- Dix Average Contour Map Method

Mean Average Velocity Line Method

Mean average velocity graph is formed by plotting the two-way times and the V_{ave} given in the velocity panels at different CDPs of the seismic section. The times are plotted on the x-axis and the V_{ave} is plotted on the y-axis. An average or best-fit line is drawn through the plots of the seismic velocity panels given. This is the Average Velocity line of the seismic section. These velocities are then used to convert the two-way times of the reflectors into depth, for the formation of the depth section. The picked times and the average velocities of the reflectors were used to construct the depth section.

Dix Average Contour Map Method (Iso-Velocity Map)

In this method, the velocities used for the determination of the depth of every reflector are estimated with the help of Average velocity contour map (Iso-velocity map). The different average velocities under the SPs were plotted along their respective times. Dix average velocity varies form 1500 m/s to 5000 m/s for 5sec data and has been used to construct the Iso-velocity map at interval of 200 m/s. The time of the reflectors were read by overlaying the Iso-velocity map and the seismic time section. The velocities were then used in calculating the depth of the horizons first by converting the two-way time in to one-way time, the multiplying it with the velocity obtained.

Results of Line Interpratation

Reflector 1 (R1)

The reflector R1 varies between at the time range of 0.68 sec (680 msec) to 0.69 sec (690 msec) in the time section.

By using mean average velocity method, the velocity ranges from 1950 m/sec to 2000 m/sec and the depth ranges from 663m to 690m.

Reflector 2 (R2)

The reflector R2 varies between SP 200 to SP 450 at the time range of 0.79 sec (790 msec) to 0.80 sec (800 msec) in the time section.

Whereas in mean average velocity method, the velocity ranges from 2060 m/sec to 2075 m/sec and the depth ranges from 814 m to 830 m.

Reflector 3 (R3)

The reflector R4 varies between SP 200 to SP 450 at the time range of 1.09 sec (1090 msec) to 1.16sec (1160msec) in the time section.

Whereas in average velocity method, the velocity ranges from 2240 m/sec to 2315 m/sec and the depth ranges from 1221 m to 1343 m.

Reflector 4 (R4)

The reflector R5 varies between SP 200 to SP 450 at the time range of 1.4 sec (1400msec) to 1.43sec (1430sec) in the time section.

Whereas in average velocity method, the velocity ranges from 2430 m/sec to 2500 m/sec and the depth ranges from 1701 m to 1800 m.

Reflector 5 (R5)

The reflector R6 varies between SP 200 to SP 450 at the time range of 2.37 sec (2370msec) to 2.41 sec (2410msec) in the time section.

Whereas in average velocity method, the velocity ranges from 3015 m/sec to 3060 m/sec and the depth ranges from 3600 m to 3687 m.

Reflector 6 (R6)

The reflector R6 varies between SP 200 to SP 450 at the time range of 2.64 sec (2640msec) to 2.7 sec (2700msec) in the time section.

Whereas in average velocity method, the velocity ranges from 3140 m/sec to 3200 m/sec and the depth ranges from 4145 m to 4320 m.

Whereas in average velocity method, the velocity ranges from 3140 m/sec to 3200 m/sec and the depth ranges from 4145 m to 4320 m.

Reflector 7(R7)

The reflector R6 varies between SP 200 to SP 450 at the time range of 2.84 sec (2840msec) to 2.89 sec (2890msec) in the time section.Whereas in average velocity method, the velocity ranges from 3330 m/sec to 3390 m/sec and the depth ranges from 4728 m to 4898 m.

Depth section based on the Average Velocity method represents the true picture of the area, which is very similar to the time section of the area. Also the shape and dimensions of the reflectors resemble the reflectors marked on the provided seismic section.

7.5. Conclusion

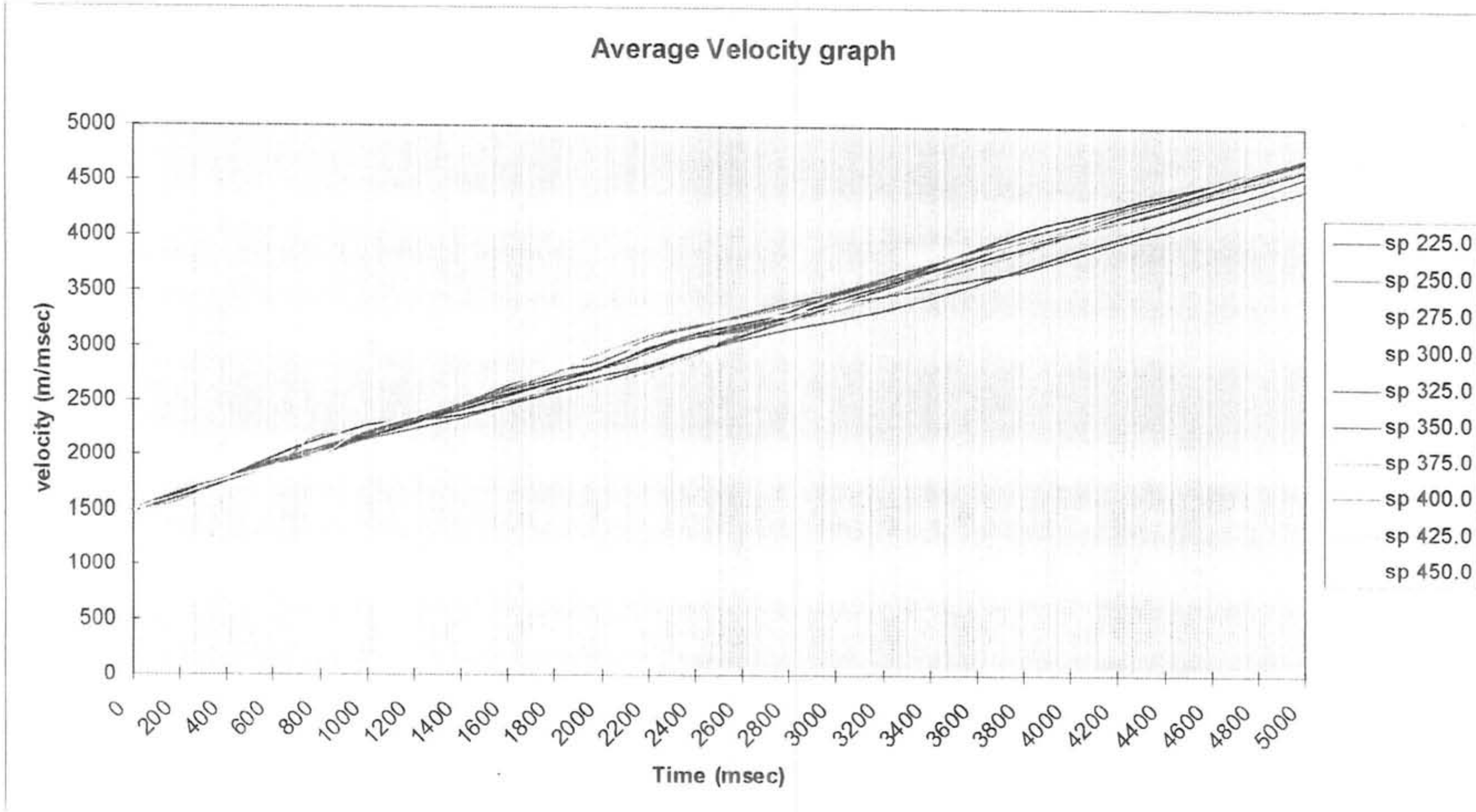
After interpreting of the given seismic section, following conclusions are made:

- The data quality is not so good in the given section and can be attributed to poor quality of migration.
- Observed reflectors are nearly horizontal showing very slight undulations.
- As the seismic line is along the dip so there was great chance of a structural trap. The seismic section shows a thick sedimentary deposit in the Central Indus Basin.
- The interpretation is mainly done based on reflection of different interfaces and stratigraphy of the area and the surrounding well.
- The stratigraphic correlation indicated that reflector R1 is of Miocene age, reflector R2 is of Middle Eocene age, reflector R3 is of Middle Eocene age, and reflector R4 is Lower Eocene age, R5 is of Paleocene age , R6 is of Upper Cretaceous and R7 also is of Upper Cretaceous.
- F3 normal fault present in reflectors R5,R6 and R7
- F1 and F2 normal faults present in R6 and R7, also F2 which form Horst and Graben structure combinally with R3.

- Horst and graban structure are formed because of normal faulting.
- This extensional regime satisfy the regional tectonics of the area.

Time msec	S.P 225	S.P 250	S.P 275	S.P 300	S.P 325	S.P 350	S.P 375	S.P 400	S.P 425	S.P 450
0	1500	1500	1500	1500	1500	1500	1500	1500	1500	1500
200	1638	1614	1652	1621	1614	1639	1652	1621	1574	1627
400	1784	1763	1783	1747	1754	1764	1764	1779	1779	1762
600	1975	1917	1933	1910	1897	1917	1925	1945	1896	1900
800	2142	2034	2089	2059	2031	2012	2042	2077	2178	2006
1000	2273	2142	2250	2202	2180	2165	2215	2205	2230	2238
1200	2322	2244	2341	2314	2314	2287	2332	2333	2320	2357
1400	2371	2347	2421	2415	2425	2458	2460	2473	2472	2498
1600	2466	2475	2522	2504	2546	2571	2595	2641	2615	2654
1800	2577	2622	2615	2598	2669	2676	2704	2781	2752	2782
2000	2688	2740	2672	2692	2811	2796	2798	2868	2945	2911
2200	2821	2840	2729	2773	2976	2996	2947	3068	3112	3044
2400	2965	2961	2843	2936	3126	3082	3090	3183	3208	3175
2600	3109	3077	3002	3136	3209	3178	3147	3302	3280	3248
2800	3253	3162	3057	3242	3335	3285	3275	3404	3375	3314
3000	3397	3242	3189	3347	3482	3441	3400	3509	3474	3386
3200	3541	3343	3314	3478	3602	3574	3475	3635	3593	3514
3400	3668	3458	3435	3609	3747	3696	3549	3760	3711	3662
3600	3786	3588	3564	3741	3905	3815	3624	3885	3835	3811
3800	3904	3725	3694	3873	4061	3957	3699	4007	3962	3957
4000	4022	3863	3832	4006	4170	4066	3819	4128	4082	4058
4200	4140	4001	3972	4138	4279	4193	3943	4249	4216	4197
4400	4258	4139	4112	4271	4388	4320	4066	4370	4343	4336
4600	4376	4277	4251	4403	4497	4448	4189	4492	4469	4474
4800	4494	4415	4391	4515	4607	4575	4313	4613	4596	4614
5000	4612	4553	4531	4628	4715	4702	4436	4728	4723	4753

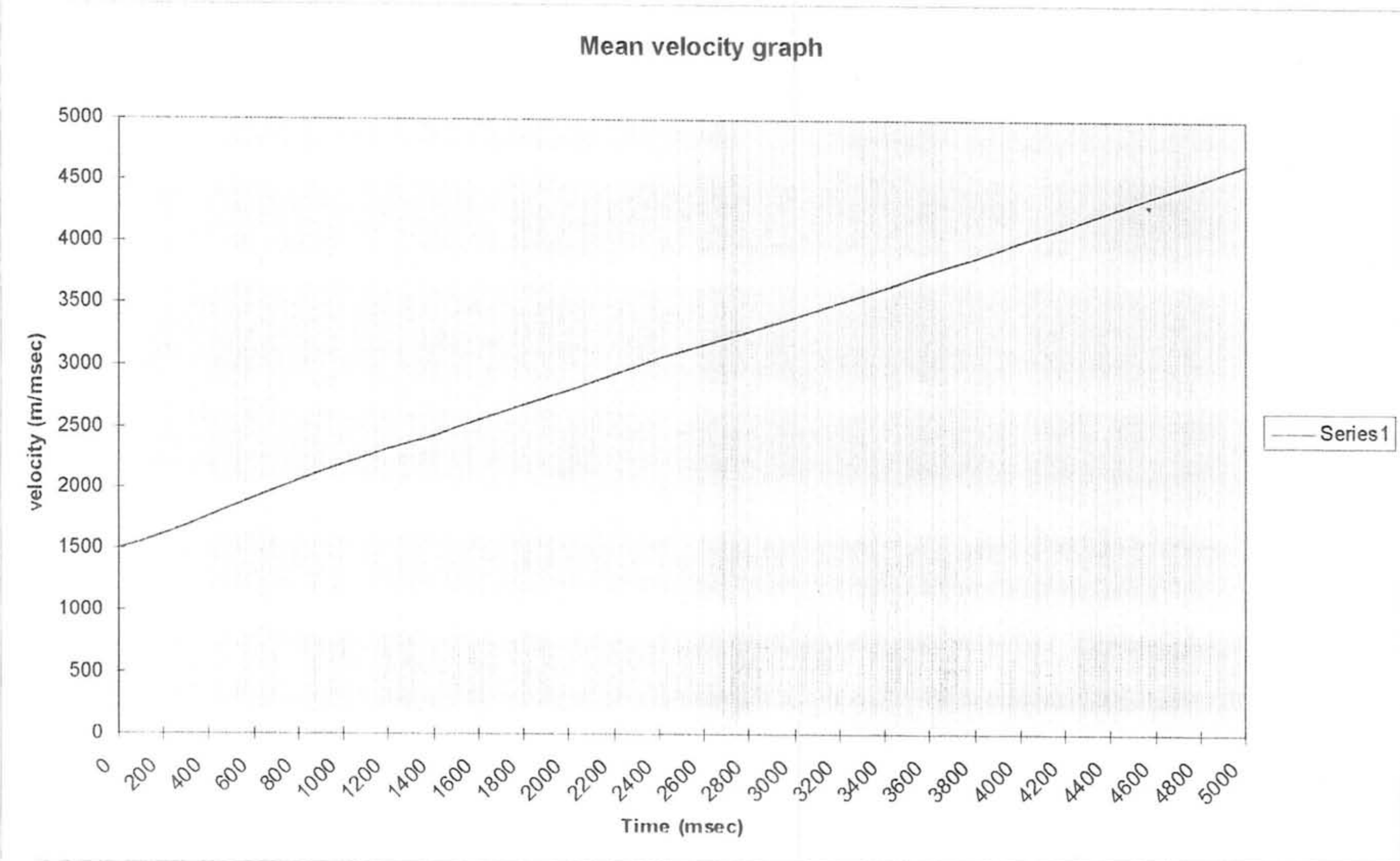
Table 1.1: Table for the Average Velocities of the line 985-QPR-04



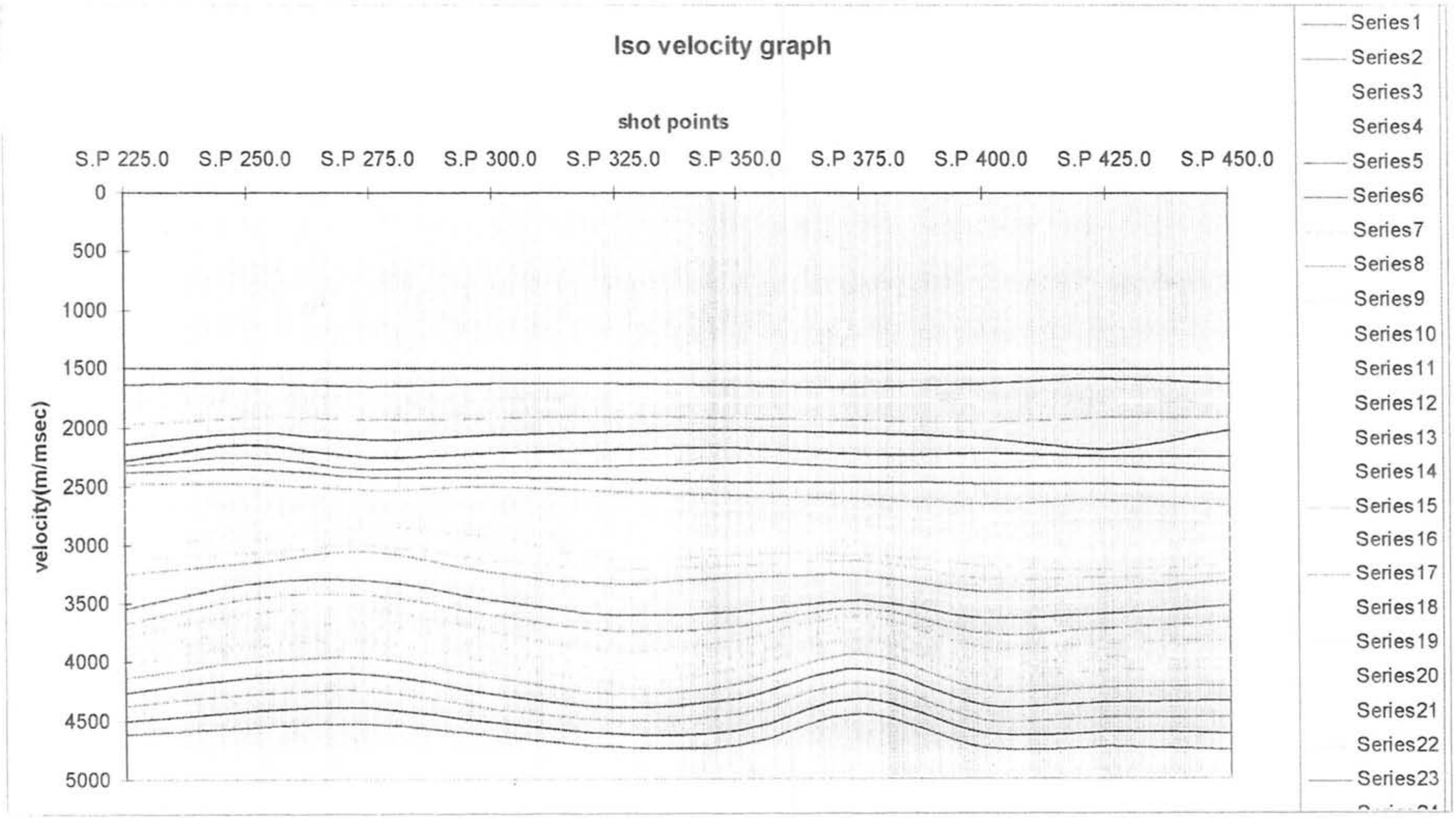
Average velocity graph of the line 985-QPR-04

Time(msec)	Mean Velocity
0	1500
200	1625
400	1768
600	1921
800	2067
1000	2205
1200	2316
1400	2434
1600	2559
1800	2678
2000	2792
2200	2931
2400	3057
2600	3169
2800	3270
3000	3387
3200	3507
3400	3629
3600	3755
3800	3884
4000	4005
4200	4133
4400	4260
4600	4388
4800	4513
5000	4638

1.2 Table for the Mean Velocity of the line 985-QPR-04



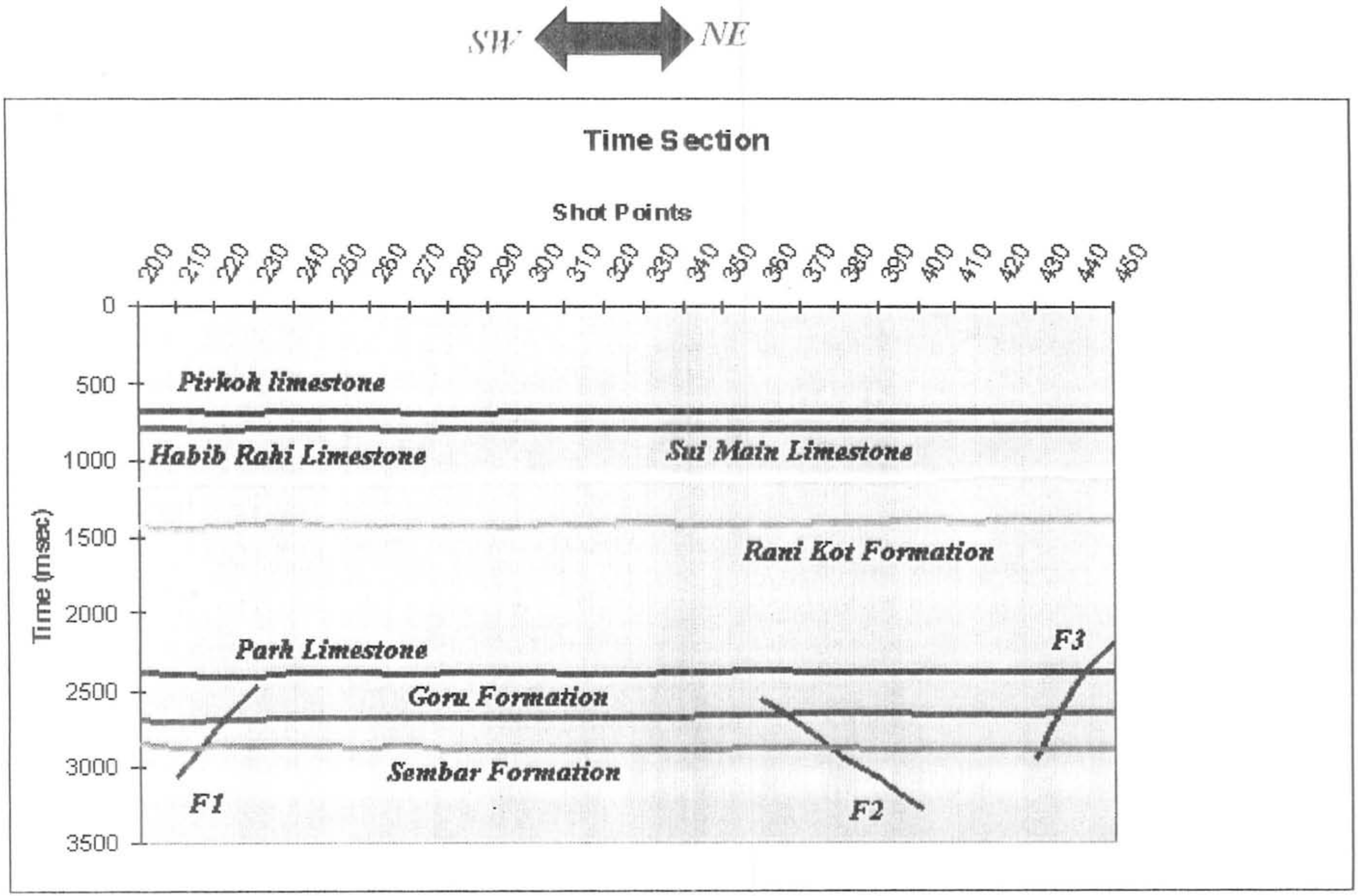
Mean velocity graph of the line 985-QPR-04



Iso velocity graph of the line 985-QPR-04

Shot Points	R1	R2	R3	R4	R5	F3	R6	F1…F2	R7
S.P	msec	msec	msec	msec	msec	msec	mse)	msec	msec
200	680	800	1160	1430	2390		2690		2840
210	680	800	1140	1440	2400		2700	3050	2860
220	690	810	1140	1430	2410		2690	2750	2850
230	690	800	1140	1420	2410		2690	2480	2850
240	680	795	1140	1410	2390		2680		2850
250	680	790	1140	1430	2390		2680		2850
260	685	800	1140	1430	2390		2680		2860
270	690	810	1150	1430	2400		2680		2850
280	690	800	1140	1430	2390		2680		2880
290	690	800	1120	1430	2390		2680		2880
300	685	800	1130	1430	2390		2680		2880
310	680	800	1120	1420	2400		2680		2880
320	680	800	1120	1420	2400		2680		2880
330	680	800	1120	1410	2400		2680		2880
340	680	800	1120	1420	2390		2680		2880
350	680	800	1120	1420	2380		2670		2880
360	680	800	1110	1420	2370		2670	2560	2870
370	680	800	1120	1420	2390		2640	2730	2870
380	685	800	1110	1410	2390		2640	2930	2870
390	685	800	1110	1410	2390		2640	3080	2880
400	680	800	1110	1400	2390		2650	3280	2890
410	680	800	1100	1410	2390		2650		2890
420	680	800	1090	1400	2390		2650		2890
430	680	800	1090	1400	2390	2940	2650		2890
440	680	795	1100	1400	2380	2480	2640		2890
450	680	800	1100	1400	2380	2190	2640		2890

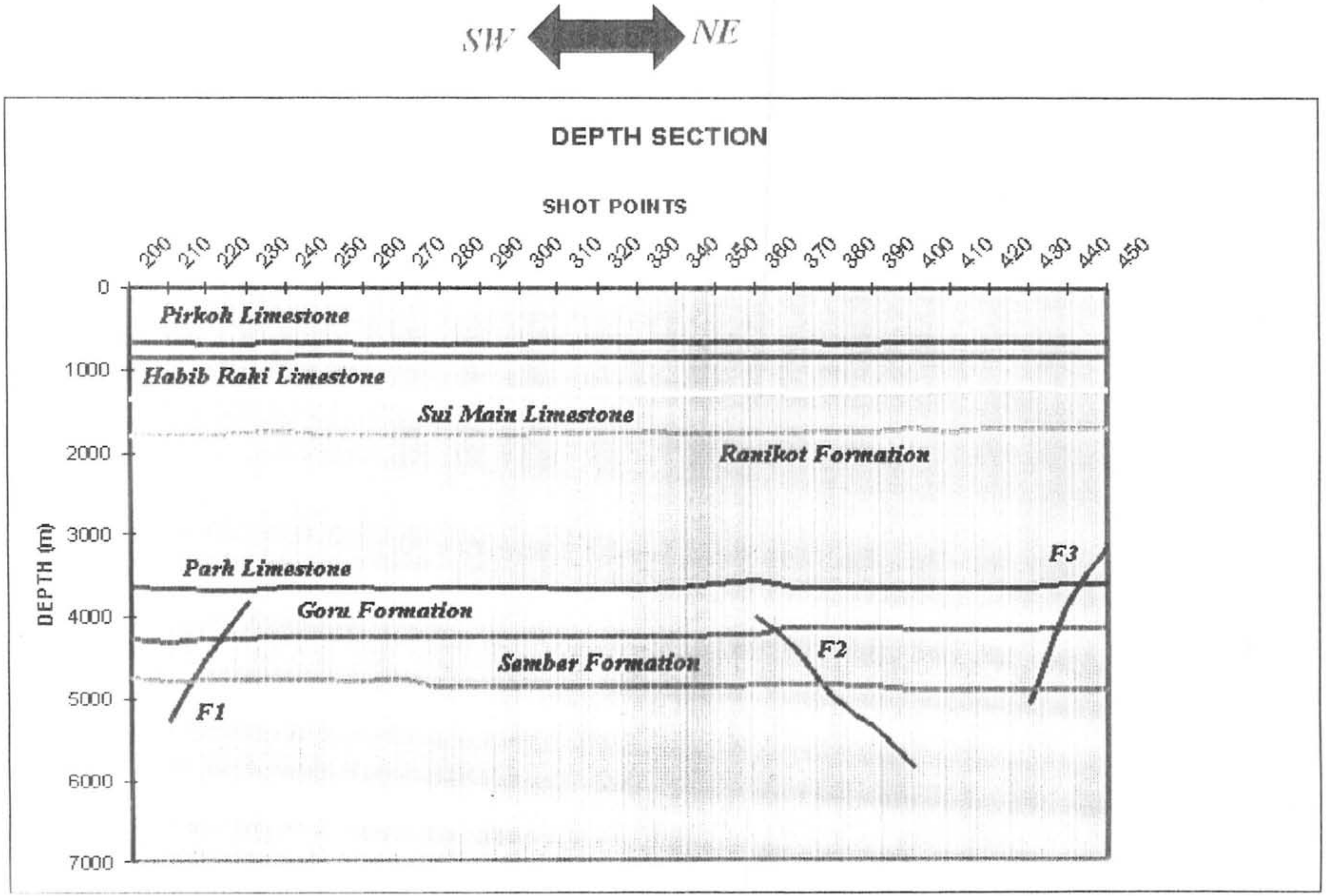
Table 1.3: Table for the Time Section of the line 985-QPR-04



TIME SECTION FOR THE LINE 985-QPR-04

S.P	R1(m)	R2(m)	R3(m)	R4(m)	R5(m)	F3(m)	R6	F1.......F2(m)	R7
200	663	830	1343	1780	3633		4291		4729
210	663	830	1305	1800	3660		4320	5261	4791
220	690	851	1305	1780	3687		4291	4496	4760
230	690	830	1305	1757	3687		4291	3844	4760
240	663	823	1305	1727	3633		4261		4760
250	663	814	1305	1780	3633		4261		4760
260	676	830	1305	1780	3633		4261		4791
270	690	851	1323	1780	3660		4261		4760
280	690	830	1305	1780	3633		4261		4860
290	690	830	1271	1780	3633		4261		4860
300	676	830	1288	1780	3633		4261		4860
310	663	830	1271	1757	3660		4261		4860
320	663	830	1271	1757	3660		4261		4860
330	663	830	1271	1727	3660		4261		4860
340	663	830	1271	1757	3633		4261		4860
350	663	830	1271	1757	3600		4225		4860
360	663	830	1254	1757	3573		4225	4013	4829
370	663	830	1271	1757	3633		4145	4416	4829
380	676	830	1254	1727	3633		4145	5003	4829
390	676	830	1254	1727	3633		4145	5352	4860
400	663	830	1254	1701	3633		4174	5830	4899
410	663	830	1238	1727	3633		4174		4899
420	663	830	1221	1701	3633		4174		4899
430	663	830	1221	1701	3633	5035	4174		4899
440	663	823	1238	1701	3600	3844	4145		4899
450	663	830	1238	1701	3600	3148	4145		4899

1.4 Table For the Depth Section of the line 985-QPR-04


SW
NE
DEPTH SECTION
SHOT POINTS
200
210
220
230
240
250
260
270
280
290
300
310
320
330
340
350
360
370
380
390
400
410
420
430
440
450
DEPTH (m)
0
1000
2000
3000
4000
5000
6000
7000
Pirkoh Limestone
Habib Rahi Limestone
Sui Main Limestone
Ranikot Formation
Park Limestone
Goru Formation
Sembar Formation
F1
F2
F3