Department of Exploration Geophysics

CO₂ Quantification Using Seismic Attributes in Laboratory Experiments

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This thesis is presented for the Degree of Doctor of Philosophy of Curtin University of Technology

June 2007
Declaration

To the best of my knowledge and belief this thesis contains no material previously published by any other person except where due acknowledgment has been made.

This thesis contains no material which has been accepted for the award of any other degree or diploma in any university.

Signature: ..............................................

Date: .................................
This work is dedicated to

My mother and my wife, Mina
for their continuous support,
encouragement and motivation to me
during my PhD research
ABSTRACT

Sequestration has been suggested as a solution for resolving the problem of increasing greenhouse gas emissions. CO₂ is the major greenhouse gas which results from using fossil fuels for domestic and industrial purposes. Different geological targets have been suggested as reservoirs for CO₂ sequestration with saline aquifers being the focus of this research. Monitoring and verification of injected CO₂ into the ground is an essential part of CO₂ sequestration because there is a strong requirement to understand and correctly manage the CO₂ flow and movement within the reservoir over time. This includes a need to understand mobile CO₂ in its all phases (gas, liquid, supercritical and dissolved in formation water).

It is now well recognised that monitoring injected liquids in the sub-surface can be done remotely using surface seismic monitoring techniques. Seismic waves are sensitive to the contrast in the physical properties of formation water and CO₂. As a gas, the migration path of CO₂ has been shown to be easily imaged but such images provide only a qualitative rather than a quantitative solution, which is inadequate to remotely verify storage volumetrics. The complexity of saline aquifer reservoirs containing the different phases of CO₂ (a function of reservoir pressure, temperature, and chemical composition and the state of phase of injected CO₂) requires a good knowledge base of how the seismic response changes to such changes in CO₂ phase and reservoir heterogeneities for verification purposes.

In this research, transmission ultrasonic seismic experiments were performed under controlled pressure, temperature and CO₂ dissolution conditions in water. Different forms of simulated rock matrix were used to understand how seismic attributes changed with changing sequestration conditions. Data analysis showed that the commonly used approach of seismic velocity analysis is not particularly sensitive to dissolved CO₂ whereas seismic amplitude was very sensitive to dissolved CO₂ content and is the seismic attribute of choice for the future quantification of CO₂. The density increase in formation water brine as a result of CO₂ mixture was found to be directly related to transmission amplitude and provides the potential for prediction and thus, remote quantification.
Also, there was confirmation during the transmission experiments that seismic amplitude changes markedly when CO₂ changes phase from its dissolved form into a gas, as a result of significant attenuation by CO₂ bubbles. Analysis showed that the dominant and centre frequency of the spectra also responded to CO₂ phase when it changed from dissolved to its free gas form. However, these attributes appear to be of use in a qualitative manner rather than quantitative.

The CO₂ pre-bubble phase was studied in an attempt to obtain a basic knowledge of the effect on seismic amplitude variation for quantifying dissolved gas amounts with some success. This knowledge has an application in Gas-to-Oil-Ratio mapping in depleting oil fields and can assist the future management of production from fields which are at the stage of near-bubble point due to pressure depletion.

The results of this research have an application in time-lapse seismic monitoring and operational management of greenhouse gas sequestration operations. In particular, the VSP and cross-well seismic methods are immediate beneficiaries of this research, with further work required for application to 3-D reflectivity methods in time-lapse surface seismic monitoring.
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1.1 - World Energy Consumption
The IEO2006 (International Energy Outlook, 2006) predicts an increase in world consumption of marketed energy from all sources over the next two and one-half decades. Fossil fuels (oil, gas, and coal) will continue to supply much of the increment in marketed energy use worldwide throughout its projections into the future. Based on that report, oil remains the dominant energy source over the projection period (2003-2030), but its share of total world energy consumption declines from 38 percent in 2003 to 33 percent in 2030 (Figure 1.1).

Furthermore, according to IEO2006, worldwide oil consumption rises from 80 million barrels per day in 2003 to 98 million barrels per day in 2015 and then up to 118 million barrels per day in 2030. Worldwide, the transportation section and industry are the major growth sectors for oil demand. On a global basis, the transportation sector—where there are currently no alternative fuels that compete widely with oil—accounts for about one-half of the total projected increase in oil use between 2003 and 2030, with the industrial sector accounting for another 39 percent of the incremental demand.

The higher world oil price pathway in the IEO2006 report also predicts that natural gas would be the fastest growing energy source in the mid-term; however, higher natural gas prices according to IEO2006 make coal more cost-competitive especially in the electric power sector, and as a result natural gas use and coal use increase at similar rates. Natural gas demand rises by an average of 2.4 percent per year over the 2003 to 2030 period and coal use increases by an average of 2.5 percent per year. Total world natural gas consumption rises from 95 trillion cubic feet in 2003 to 134 trillion cubic feet in 2015 and to 182 trillion cubic feet in 2030.

The industrial sector remains the most important end-use consumer for natural gas worldwide, accounting for 52 percent of the total growth in natural gas use in the
projections; however, natural gas also remains an important energy source in the electric power sector, particularly as a fuel for new generating capacity. The electric power sector accounts for 39 percent of the increase in global natural gas demand over the period 2003 to 2030, although the higher price pathway according to IEO2006 leads to a slower growth rate for natural gas consumption in the electricity generation sector than was projected in IEO2005. Natural gas is still seen as a desirable option for electric power in many parts of the world, given its efficiency relative to other energy sources and its low carbon content relative to other fossil fuels, making it a more attractive choice for countries interested in reducing greenhouse gas emissions.

Figure 1.2 shows total consumption of energy from 1980 to 2030 in which there is nearly 30 percent of the total consumption projected for 2030, between the projections for the high and low economic growth cases.

1.2 - World Energy Production

World oil output is depleting with the majority of future production limited to a few countries. Figure 1.3 summarizes the uneven distribution of conventional oil in the world (Zagar, 2000). Produced oil is shown in green. Oil to be produced comprises of the known reserves (shown in red) and the yet-to-find (shown in yellow). The key point here is that about half of the oil to be produced lies in just five Persian Gulf countries, Iraq, Iran, Kuwait, U.A.E and Saudi Arabia which can make up the difference between world supply and demand. As Zagar stated in his report (2000), the rate of production of oil from all countries other than these countries in the Persian Gulf area, cannot be compensated by exploring for new reservoirs. It is vital that existing petroleum reservoirs minimize their depletion rates.

Production from most oil reservoirs includes three distinct phases: primary, secondary, and the tertiary phase which is enhanced recovery (U.S. Department of Energy, 2004). During primary recovery, the natural pressure of the reservoir or gravity drive oil into the wellbore and artificial lift techniques (such as pumps) bring the oil to the surface. But only about 10 percent of a reservoir's original oil is typically produced during primary
recovery. Shortly after World War II, producers began to employ secondary recovery techniques (water and natural gas injection) to extend the productive life of oil fields, often increasing ultimate recovery to more than 20 percent.

Tertiary, or enhanced oil recovery (EOR) techniques includes the use of steam-assisted gravity drainage (SAGD) method (Thomas and Michael, 2004) which causes a reduction in the viscosity of the oil and increase in production. Injecting CO2 is another tertiary recovery method. In this method CO2 mixes and dissolves with the residual oil causing it become lighter and have increased mobility (Stevens et al., 2001).

A drop in pressure with production is a common problem for depleting reservoirs. As a consequence of decreased reservoir pressure, dissolved gas in oil starts to become liberated to its free gas form. This will cause two main problems: decreasing viscosity of oil (Jacoby et al. 1969), and pushing down the reservoir’s oil column; both of these problems cause a reduction in production of oil from reservoirs. It is proposed that if we can predict the gas bubble forming in depleting reservoirs before it appears as free gas in fluid, it will be possible to better manage the reservoir in terms of EOR. The current management approach is through monitoring pressure reduction, and then increasing the pressure through fluid injection at the same point.

As can be seen in Figure 1.4, a huge proportion of oil in place (58%) can’t be recovered by current technologies. This area is the target for improved recovery technology.

1.3 - Greenhouse Gas Effect and Global Warming

According to the US National Academy of Sciences (2001), the Earth's surface temperature has risen by about 1 degree Fahrenheit in the past century, with accelerated warming during the past two decades. There is new and strong evidence that most of the global warming over the last 50 years is attributable to human activities such as using oil and gas resources to produce electricity for industry or providing fuel for transportation and home usage (US Environmental Protection Agency, 2004). Human activities have altered the chemical composition of the atmosphere through the build-up of greenhouse
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gases – primarily carbon dioxide, methane, and nitrous oxide. It is argued that El Nino and La Nina are two phenomena caused by such changes in climate in countries around the Pacific. These cause drought and flood in dry and humid parts of the world, respectively (US Department of Commerce, 2004).

Carbon dioxide is the dominant greenhouse gas resulting from the burning of fossil fuels, and it comprises more than half of all man-made greenhouse gas emissions (Hitchon et al., 1998). An increase in using hydrocarbon resources to produce energy over the next few decades will cause more carbon dioxide gas to be released by the domestic use of cars, and power plants thereby increasing the greenhouse gas effect.

1.4 - CO\textsubscript{2} Storage and Sequestration

One good way to decrease the greenhouse gases which have been released from industrial activities is their sequestration underground. This includes injection into hydrocarbon reservoirs to enhance oil and gas recovery. The EOR industry has been using CO\textsubscript{2} as an injection gas to increase the pressure of reservoirs and reduce the viscosity of oil for many years (Oxy Permian Ltd., 2003). But CO\textsubscript{2} injection offers another benefit: it could be a promising way to sequester carbon dioxide from power plants and other energy facilities resulting in a reduction in gas in the atmosphere (U.S. Department of Energy, 2004).

Besides oil and gas fields, saline aquifers, deep ocean waters, and deep un-mineable coal seams can all be used as sequestration reservoirs. Coal seams in particular can be a good alternatives since they can have existing methane gas recovered by the process of CO\textsubscript{2} sequestration, thereby making the process potentially economic regardless to some technical difficulties (Salehi and Gowlli, 2006; Saghafi et al., 2007).

1.5 - Seismic Monitoring of CO\textsubscript{2} Sequestration

As old hydrocarbon fields deplete, there is an increasing importance in oil recovery, which is not assisted by the growing complexity of recently discovered oil fields, and the growing realization that reservoirs are more heterogeneous than assumed in the past. As a
result there has been a major increase in the use of 3-D seismic methods during the past decade. One of the central aspects of this increase involves the need to better understand the relationship between the seismic properties of reservoir rocks and their production properties (porosity, permeability) and state (mineralogy, saturation, pore pressure, etc.). Some areas of seismic imaging applications include the evaluation of stratigraphic traps, fracture detection, and the spatial distribution of porosity and permeability.

The most pressing issue in CO\(_2\) sequestration is the prevention of CO\(_2\) leakage to the surface. CO\(_2\) which is normally injected as a compressed liquid or as a gas (supercritical form) may change its state of phase to free gas if it meets a higher temperature and gradual reduction in pressure when entering a reservoir. Good reservoir management ensures that CO\(_2\) does not come to the surface or pollute water or soil reserves. So, monitoring the fluid movement during CO\(_2\) injection scenarios used for CO\(_2\) sequestration or enhanced oil recovery in depleting reservoirs is another requirement through the use of seismic methods (Ronald and Marcus, 2000).

Reservoir complexity is typically related to significant spatial heterogeneities in terms of porosity, permeability, clay content, fracture density and other properties. A direct consequence of such heterogeneity is the complexity of reservoir processes, ranging from migration of the gas cap in reservoirs with discontinuous shales, over-pressured zones, and the tracking of any injected water, steam, or temperature during the production of fluids from reservoirs with large spatial variations in permeability. This spatial aerial variability cannot be determined at any level of detail using only well testing data, logs, or rock cores. It can best be obtained using remote geophysical measurements, especially seismic measurements, and only when these seismic measurements can be understood in terms of reservoir properties (Hardy, 2003).

The seismic method which is becoming more commonly used to monitor oil and gas reservoir production is \textit{time-lapse} 3-D which is also one of the main methods available for monitoring the CO\(_2\) injection process. Time-lapse vertical seismic profiling (VSP), which is based on transmission of seismic waves, is another seismic method and with
well logging has commonly been used for injection monitoring purposes. Other methods used for monitoring and verification of CO₂ injection include geochemical measurements of soil, air samples and non-seismic geophysical methods.

The need for requiring many wells to achieve a correct evaluation of a reservoir’s parameters is an expensive issue, which encourages the increased use of seismic methods for reservoir characterization purposes rather than drilling additional holes.

**1.6 - Seismic Verification (Quantification) of CO₂ in-place**
Seismic data are often used for quantitative purposes: Volumetric estimations based on 3-D seismic images, overpressure prediction based on seismic velocity analysis, net-to-gross estimation based on seismic amplitude attribute analysis and so on (Landro, 2003). Seismic verification can be very difficult due to the complexity of a reservoir and the noise level of the recorded data; however experimental work on laboratory data which records data under stable conditions, using controlled seismic source and receiver settings have been shown as a good way for the verification of CO₂ bearing fluids within the rock matrix (Keshavarz and Evans, 2005).

In using seismic methods for CO₂ sequestration underground two aims are to be achieved: one aim is in understanding the location of the CO₂ while the other is quantifying the CO₂ irrespective of phase, which is a far harder issue to tackle. In practice, we must be able to monitor the state-of-phase and its rate of change wherever there is a gas-occupied volume to allow us to choose the best option to control it.
Quantifying the changes in CO₂ content and phase in a reservoir is necessary for more accurate volumetric assessments. We know how much CO₂ has been injected, and if this is different from the amount quantified through any monitoring processes, then we must know where the difference is located- otherwise there is potential for loss of control and hence, loss of reservoir management.

Searching for qualitative and quantitative tools for understanding and predicting the effects of lithology, pore fluid type, saturation, stress, pore pressure, fractures and
temperature on seismic properties are the main research subjects of rock physics. Over the last three decades, enormous studies have been made to understand the relationship between the physical properties of reservoir rocks, and geophysical observations, the science now known as “rock physics”. We have discovered relationships, for example, between sonic velocity and porosity, permeability, and the velocity of p-waves and s-waves versus saturation and lithology. Some of these relationships have involved the effects of pore pressure, stress, temperature, clay content, compaction, fluid type and saturation. Seismic methods are sufficiently sensitive to fluid phase changes that they could be used to quantify CO$_2$ injected into the ground (for sequestration or EOR purposes) in a dissolved form (Keshavarz and Evans, 2005).

The research discussed here mainly involves experiments to find a relationship for fluid phase changes, especially during the pre-bubble condition, which has not been previously studied. This will allow the seismic measurements to be more quantitative for the prediction of fluid saturation, over-pressure and fluid phase-change monitoring, in the future.

1.7 - Aim of the Research
Seismic methods are commonly used for monitoring fluid movement during CO$_2$ injection scenarios in enhanced oil recovery operations over depleting reservoirs (Ronald and Marcus, 2000). In the EOR situation, accurate monitoring of the flow of carbon dioxide can be achieved by recording large amounts of data acquired at the surface and through the use of many production and injection wells. But, geological storage of carbon dioxide in a saline aquifer is a different case by comparison with EOR projects. Since saline aquifers are not economic resources, just a few boreholes drilled into the formation can provide high resolution monitoring for carbon dioxide migration, and when coupled with surface seismic, could provide the optimum way to remotely study the carbon dioxide migration through the aquifer.

CO$_2$ injected into a reservoir may be in the form of a liquid or super-critical fluid, depending on the pressure and the temperature conditions of the reservoir. Some injected
CO₂ will replace the interstitial reservoir formation water (Figure 1.5). The CO₂ saturation level in the rock matrix is often highest at the injection point and would be expected to decrease as it mixes with the water when it moves away from the injection point dissolving with the water as it flows away. So, it is important to note that CO₂ not only displaces the formation water, but also dissolves in the formation water in quantities depending on the temperature, pressure and salinity conditions (Carroll et al., 1991).

Observations from Sleipner 3-D seismic field data have shown that the CO₂ bubble has a similar seismic amplitude response to a conventional gas “bright spot”, which is a problem when attempting to establish precise CO₂ reservoir volumetric measurements (Keshavarz and Evans, 2004). While it has been shown that dissolved gas has little effect on the properties of pore fluid (McKenna, 2004), no models currently exist to correctly quantify the amount of liquids and gas during this phase change from a dissolved gas to a free gas, probably because the phase-change effect can be transient in the laboratory, and hard to simulate under real field conditions. The effects however, are clearly apparent on field data (Keshavarz and Evans, 2004).

Much has been done to study the gas/liquid effect on seismic velocity (Arts et al., 2004; Sinartio, 2002; Lei and Xue, 2006; Siggins, 2006; Wandler, 2007), but quantification of dissolved CO₂ has not been mentioned in any papers. The main aim of this research was to study the effect of changes in dissolved CO₂ concentration on seismic attributes using simple physical experimental laboratory models. A knowledge of the seismic response to changes in dissolved CO₂ in a brine mix may allow us to better monitor the efficiency of CO₂ injection while providing a more precise volumetric assessment for reservoir control purposes. This thesis focuses on a seismic study of dissolved CO₂ in the sequestration process- an area that has not been studied before- to help verify conventional seismic methods and provide corrections for seismic volumetric studies.
Figure 1.5. CO$_2$ replaces the formation water after injecting into the ground. Also some quantity of it dissolves in water. This dissolved CO$_2$ is at its highest concentration close to the zone occupied by CO$_2$ and dilutes when it moves away from the injection point (modified after CO2CRC website).
Another application of this thesis would be in mapping the gas-water ratio. In the petroleum industry, whether in a master development plan or EOR scenario, the gas-oil ratio (GOR) is one of the important parameters which are used to determine the best site for new production or injection wells. GOR values obtained by sampling the fluids of a reservoir from a well at different depths can be a difficult process. Interpolation of values could be invalid due to the large distance between wells or the complexity of a reservoir. In this case seismic methods may be used to give a large aerial coverage to allow the prediction of values of GOR over the entire reservoir.

A final benefit of this research may be in characterizing fizz water. Many high cost wells have been drilled in the belief that their seismic response was the same as large gas quantities. However, low gas saturated reservoirs have been recognized as providing a seismic response. Fizz water is often blamed as a false hydrocarbon indicator (Han and Batzle, 2002). The result has been the production of small amounts of fizz gas instead of economic gas reserves (Poupon et al., 1999). Results of this research can help to solve the problem of fizz gas, because some characteristics of seismic waves may be different for fizz gas compared with commercial gas reservoirs (Zhu et al., 2000).

1.8 - Thesis Configuration
The structure of my thesis is designed to follow the chronological development of my work. A discussion on world energy use and production followed by global warming and the need for my research program has been presented in this Chapter. In Chapter 2, the reader is provided with the concepts of geological storage of carbon dioxide. The Chapter begins by providing a general discussion of the various sites for geological storage, with an emphasis on saline aquifers. Chapter 2 concludes with a general discussion of potential monitoring technologies including seismic and non-seismic methods. The direct transmission seismic method is the main monitoring tool which I used to observe the storage of carbon dioxide in the physical models I used for my laboratory experiments. Chapter 3 provides the reader with the theoretical basis of seismic monitoring. The fundamental theories of seismic wave propagation, rock physics and reflectivity are
examined in this chapter followed by the basic concepts of CO₂ phase relationships due to variation in pressure and temperature conditions.

Chapter 4 then provides the reader with a description of the physical modelling setup and the controls on the recording system, followed by calibration tests and the process of dissolved CO₂ preparation. Chapter 5 presents the development of laboratory experiments including both transmission and reflection tests performed on different pressure cells with different matrices and transducers. The data analysis and seismic response to dissolved CO₂ is discussed in Chapter 6. This Chapter has a discussion on the effect of pressure and dissolved CO₂ content on seismic velocity, amplitude, instantaneous frequency and phase. Empirical relationships for pressure change versus seismic amplitude and dissolved CO₂ content are also presented in this Chapter along with a study of scattering, attenuation and dispersion effects caused by adding a matrix as the rock frame to the cell filled with different CO₂ content fluids.

The final discussions, conclusions and recommendations of my research are outlined in Chapter 7. A full list of references then completes this thesis.
CHAPTER 2 – CO$_2$ STORAGE

After becoming familiar with the CO$_2$ effect as the main contributor of greenhouse gas on climate change and global warming, and studying the increasing trend of CO$_2$ emissions as a result of development by industry and the increasing use of hydrocarbon resources in Chapter 1; in this chapter the main issues of CO$_2$ storage and the methods which have been used for monitoring CO$_2$ flow during the storage process, will be briefly explained.

2.1 - Geological Targets for CO$_2$ Storage
Geological formations considered for CO$_2$ storage are layers of porous rock more than 1000 m deep underground that are capped by a layer of non-porous rock above them. Sequestration engineers drill a well down into the porous rock and inject pressurized CO$_2$ into it. The CO$_2$ is buoyant and flows upward until it encounters a layer of non-porous rock and becomes trapped. There are other mechanisms for CO$_2$ trapping as well. CO$_2$ molecules can dissolve in brine, react with minerals to form solid carbonates, or adsorb in the pores of the porous rock (Johnson et al., 2001).

CO$_2$ is injected into different reservoirs depending on their different storage requirements. Before CO$_2$ was ever considered for injection into the ground as a way to reduce carbon released into the atmosphere, it was routinely injected into oil fields to enhance overall oil recovery (Jacoby et al. 1969; Oxy Permian Ltd., 2003; Orr, 2004). Over subsequent years, a wealth of technical knowledge was obtained by the petroleum industry which forms the fundamental basis upon which future storage projects are now being considered.

CO$_2$ can be injected into depleted oil and gas fields in order to increase the reservoir pressure, thereby using CO$_2$ to enhance oil and gas recovery by reducing the viscosity of heavy oils. It is also injected into coal beds for extracting methane captured in the coal. These are some of the options for which CO$_2$ injection is being used, for economic purposes which could also be considered as potential geological sites for CO$_2$ storage.
Storing CO₂ in coal beds, deep water and saline aquifers are also options for CO₂ sequestration purposes which will be discussed in this chapter.

2.1.1 – Enhanced Oil Recovery (EOR)

World oil output is depleting with the majority of future production limited to a few countries. It has been suggested that the rate of production of oil from all countries other than countries in the Persian Gulf cannot be compensated for by exploring for new reservoirs (Zagar, 2000). This indicates the importance of using enhanced oil production techniques to increase oil production from the old reservoirs.

As old hydrocarbon fields deplete, oil recovery becomes increasingly important, which is not assisted by the growing complexity of recently discovered oil fields, and the growing realization that reservoirs are more heterogenous and their production properties (porosity, permeability) and state (mineralogy, saturation, pore pressure) are more complex than assumed in the past.

Oil production from most reservoirs can be categorized into the primary, the secondary, and the tertiary phase- also known as enhanced recovery (EOR). Based on the 2004 U.S. Department of Energy Annual report, only about 10 percent of a reservoir's original oil is typically produced during primary recovery, which results from the natural pressure of the reservoir or gravity to drive oil to the surface and the rest of the oil in a reservoir needs to be extracted using secondary and tertiary techniques. Secondary production methods typically involve the injection of large volumes of water or natural gas into the reservoir to displace the hydrocarbons, which increases the recovery amount to 20 to 40% of original oil in place. Injecting hot water and steam, re-injecting natural gas, and injecting chemicals such as CO₂ can help to reduce the viscosity of heavy oil. These EOR techniques may increase oil production up to 30-60% of the original oil in place (U.S. Department of Energy, 2004).

The EOR technique that is attracting some interest recently is carbon dioxide (CO₂)-EOR. The first EOR project using CO₂ injection was performed in the U.S. in the early 1970s.
(Holtz et al., 2001). This was successful in improving the mobility of oil and life time of production from a number of fields, resulting in the construction of a network of pipelines to transport CO\textsubscript{2} from industrial sources to the depleting oil reservoirs (U.S. Department of Energy, 2007). Under favourable temperature and pressure conditions, CO\textsubscript{2} mixes and dissolves with the residual oil causing the oil to swell, become lighter and have increased mobility as can be seen in Figure 2.1 (Stevens et al., 2001).

With much of the easy-to-produce oil gone from U.S. oil fields, producers have attempted several tertiary, or enhanced oil recovery (EOR) techniques that offer prospects for ultimately producing 30 to 60 percent of the reservoir's original oil (U.S. Department of Energy, 2004). As Thomas and Michael (2004) have mentioned, the use of steam-assisted gravity drainage (SAGD) as a recovery technique for extraction of heavy oil, causes a reduction in the viscosity of the oil. Their technique has been said to allow an additional 15% to 25% of oil to be recovered from existing wells in regions where it is difficult to bring heavy oils to the surface.

The United States is the world leader in enhanced oil recovery technology, using about 32 million tons of CO\textsubscript{2} per year for this purpose (U.S. Department of Energy, 2004). From the perspective of the sequestration program, enhanced oil recovery represents an opportunity to sequester carbon at a low net cost, due to the increased revenues from the recovered oil/gas. In an enhanced oil recovery application, the integrity of the CO\textsubscript{2} that remains in the reservoir is well-understood and very high, as long as the original pressure of the reservoir is not exceeded. The scope of this EOR application is currently economically limited to point sources of CO\textsubscript{2} emissions that are near an oil or natural gas reservoir.

2.1.2 – Depleted Oil and Gas Reservoirs

In the petroleum industry, sustaining oil production levels is a common problem in depleting reservoirs where a reduction in pressure as a result of oil production is occurring. As a consequence of decreased reservoir pressure, dissolved gas within the oil starts to become liberated to its free gas form thereby increasing the oil’s viscosity
Figure 2.1. Schematic of enhanced oil recovery using CO\textsubscript{2} injection. CO\textsubscript{2} and mobilised oil may be produced from production wells, in which the CO\textsubscript{2} is recycled back down injection wells. A small percentage of CO\textsubscript{2} may be lost from production wells and natural fractures but over 90% of the injected CO\textsubscript{2} remains trapped within the reservoir (after Stevens et al., 2001).
(Jacoby et al. 1969), and pushing down the reservoir’s oil top while expanding the gas cap; both of these problems can result in reducing oil production. Worldwide capacity for CO₂ storage within depleting oil and gas reservoirs has been estimated to be approximately 900 Gigatons (Stevens et al., 2001).

The EOR industry has been using injected CO₂ gas for many years to increase the reservoir pressure while reducing the viscosity of oil (Oxi Permina, 2003).

Steam injection is one of the EOR methods which can reduce oil viscosity resulting in an additional 15% to 25% of oil to be recovered from existing wells in reservoirs where it is difficult to bring heavy oils to the surface; so one method to increase the reservoir’s pressure is through fluid injection.

The injection of CO₂ into depleted oil and gas reservoirs can be conducted in a manner which minimises the change in reservoir pore pressure to reduce the risk of compromising seal integrity which could result from hydraulic failure. Bradshaw and Rigg (2001) suggested that CO₂ can be stored in reservoirs that were produced initially resulting in pressure depletion (rather than a water or steam drive mechanism) to maximise the volume of CO₂ which can be injected without significantly over-pressuring the reservoir.

However, CO₂ injection offers another benefit: as discussed earlier it is a way to sequester carbon dioxide from power plants and other energy facilities thereby reducing its build-up in the atmosphere where it can contribute to the "greenhouse effect" (U.S. Department of Energy, 2004).

2.1.3 – Enhanced Coal-Bed Methane

Of the sequestration options available, geologic sequestration of CO₂ in coal formations is considered one of the methods with the greatest short-term potential (Liang et al., 2003). Coal beds typically contain a large amount of methane-rich gas that is adsorbed
onto the surface of the coal. Tests have shown that CO$_2$ is roughly twice as adsorbing on coal as methane, giving it the potential to efficiently displace methane and remain sequestered in the bed (Liang et al., 2003).

Coal bed sequestration of carbon dioxide may use CO$_2$ that has been collected as a flue gas from large emission sources such as power plants. The current practice for recovering coal bed methane is to depressurize the bed, usually by pumping water out of the reservoir thereby reducing the pore pressure causing the methane to liberate. An alternative approach is to inject carbon dioxide gas into the bed. Tests have shown that the adsorption rate for CO$_2$ is approximately twice that of methane, giving it the potential to efficiently displace methane and remain sequestered in the bed (Liang et al., 2003; U.S. Department of Energy, 2007). When coal adsorbs CO$_2$ replacing the methane, and the produced methane can thereafter be used as an energy source. This cost-effective technology has two merits: CO$_2$ emission reduction and effective use of untapped methane from coal beds.

The potential volume of CO$_2$ that could be sequestrated using enhanced coal bed methane recovery (ECBMR) has been estimated to be more than 150 Gigatons worldwide (Stevens et al., 2001). Limited pilot ECBM operations around the world have shown that combined coal seam gas recovery and CO$_2$ storage can lead to incremental volumes of methane being recovered (Gunter et al., 2005). ECBMR has provided a cost effective CO$_2$ storage since coal burning power stations will often be located close to the energy source. The economic analyses of ECBM pilot operations also suggest that this practice can be profitable with marginal increases in the price of produced gas (Saghafi et al., 2007). Good permeability within the coal seams are required for ECBMR which has the potential to provide significant reductions in atmospheric CO$_2$ emissions. Low permeability seams may need to be fractured using water injection to allow significant volumes of CO$_2$ to be injected, but such fractures may present an obvious risk to CO$_2$ containment and also limits the volume of CO2 which could be injected to the seams.

2.1.4 – Deep Unmineable Coal Seams
CHAPTER 2 – CO₂ STORAGE

Like depleting oil reservoirs, unmineable coal beds are a good opportunity for CO₂ storage. Unmineable coal seams are those which are too deep or too thin to be mined economically, but they are an economic target for methane recovery technologies such as de-watering (and de-pressurizing), or the method of sweeping the coal bed with nitrogen or CO₂. Based on NETL (2007), two or three molecules of CO₂ are adsorbed for each molecule of methane released, thereby providing an excellent storage sink for CO₂. Coal swelling due to CO₂ adsorption is a potential barrier to CO₂ ECBM by reducing the permeability which not only restricts the flow of CO₂ into the formation but also impedes the recovery of displaced CBM. NETL (2007) has recommended angled drilling techniques and fracturing as possible means of overcoming the negative effects of swelling. Beside this potential lack of permeability, high drilling costs may cause an economic impediment to prevent significant volumes of CO₂ being stored within such unmineable coal beds seams.

2.1.5 – Deep-water Disposal (Ocean Storage)

CO₂ disposal in deep ocean geological formations potentially appears to be an attractive alternative to a number of disposal concepts recently investigated. In 1997, the U.S. Department of Energy (DOE) with organizations in Japan and Norway signed the first major research agreement on options for long-term carbon dioxide disposal in Kyoto, Japan. The agreement involved a field evaluation of deep water CO₂ disposal, or sequestration, in the ocean to address the technical feasibility and the environmental impact of CO₂ ocean sequestration. This plan involved a process of pumping liquefied CO₂ more than 1,000 metres below the ocean's surface through a series of pipes. Researchers anticipated that the CO₂ would remain in the deep ocean for several hundred years. Deep water sequestration potentially could be available for about 30 percent of U.S. power plants located throughout the coastal states (DOE, 1997).

Ocean storage potentially could be achieved in two ways: by injecting and dissolving CO₂ into the water column (typically below 1,000 metres) via a fixed pipeline or a moving ship, or by depositing it via a fixed pipeline or an offshore platform onto the sea floor at depths below 3,000 m where CO₂ would be denser than water and hence expected
Figure 2.2. Overview of ocean storage concepts. In “dissolution type” ocean storage, the CO$_2$ rapidly dissolves in the ocean water, whereas in “lake type” ocean storage, the CO$_2$ is initially a liquid on the sea floor (after IPCC, 2005).
to form a “lake” that would delay dissolution of CO$_2$ into the surrounding environment (Figure 2.2). The dissolved and dispersed CO$_2$ would become part of the global carbon cycle and eventually equilibrate with the CO$_2$ in the atmosphere (IPCC, 2001). Sea water itself contains large quantities of carbon dioxide as carbonate and bicarbonate ions and as dissolved carbon dioxide gas. The system is in continual dynamic exchange with atmospheric carbon dioxide and depending on the water’s temperature and salinity, can act as a sink or a source. Thus ocean techniques proposed so far may only displace the problem into the relatively near future without ensuring permanent disposal (Hoffert, et al., 1979).

Another form of CO$_2$ disposal in the ocean is by the dumping of solid CO$_2$ blocks from a ship (Ormerod, 1994). In the case of the pipeline option, a rapid return of an important fraction of the injected volumes to the ocean surface will occur because of the density difference between the liquid CO$_2$ and ocean water. The block disposal option also will be hindered by the fact that a large portion of the solid CO$_2$ will return to the liquid or vapour phase during the slow free-fall of the solid blocks to the sea bottom. In this method, the blocks will not penetrate the seabed sediments, and the solid CO$_2$ will dissolve in the deep ocean bottom waters (Guevel et al., 1996). According to Guevel et al. (1996), This method has been suggested to be more safe and environmentally friendly, however economical issues needs to be considered more.

2.1.6 – Saline Aquifers
Saline formations are layers of porous rock that are saturated with brine. They are much more commonplace than coal seams or oil and gas bearing rock, and represent an enormous potential for CO$_2$ storage capacity (NETL, 2007). However, much less is known about saline formations than is known about crude oil reservoirs and coal seams. As a result there is a greater amount of uncertainty associated with their ability to store CO$_2$, since they have not been the subject of the high level of intensive investigation which is common in the oil and gas industry. Saline formations may have a lower permeability than hydrocarbon-bearing formations, which may subsequently require hydraulic fracturing or other similar field practices to increase their injectivity. They also
may contain minerals that could react with injected CO\textsubscript{2} to form solid carbonates. Table 2.1 compares the capacity of CO\textsubscript{2} storage in different geological targets in the US (Byrer, 2002). Based on this table, deep saline reservoirs have the highest capacity for CO\textsubscript{2} storage of about 130 Gigatons.

There are many sedimentary regions in the world (Figure 2.3) suitable for CO\textsubscript{2} storage. In general, geological storage sites should have an adequate storage capacity and injectivity, a satisfactory sealing cap-rock or confining unit and a sufficiently stable geological environment to avoid compromising the integrity of the storage site. Adequate porosity/thickness (for storage capacity) and permeability (for injectivity) are critical; porosity usually decreases with depth because of compaction and cementation, which reduces storage capacity and efficiency. The storage formation should be capped by extensive confining units (such as shale, salt or anhydrite beds) to ensure that CO\textsubscript{2} does not escape into overlying, shallower rock units and ultimately to the surface. Extensively faulted and fractured sedimentary basins or parts thereof, particularly in seismically active areas, require careful characterization to be good candidates for CO\textsubscript{2} storage.

CO\textsubscript{2} injection into a reservoir will cause significant changes within the reservoir over time. By injecting CO\textsubscript{2} into a reservoir it initially displaces the brine, whilst migrating upward due to buoyancy becoming trapped below an impermeable seal- which is called structural trapping. If the aquifer does not have enough hydrodynamic flow to accommodate the rapid influx of CO\textsubscript{2}, there is the possibility of over-pressurising. After CO\textsubscript{2} is captured within a reservoir for some time, it will dissolve into the formation brine; this process is called solubility trapping. The CO\textsubscript{2} saturated brine becomes heavier and then will descend into the reservoir (Johnson et al., 2001). Dissolution of CO\textsubscript{2} in brine releases hydrogen and bicarbonate ions into the formation water and reduces the pH which will cause an acidic condition to occur. As an acidic brine mix, susceptible minerals are then dissolved within the rock frame and the porosity increases slightly. New minerals such as calcite, dolomite and siderite are then formed when calcium, magnesium and iron ions released into the water interact with carbonate ions which were released before (Johnson et al., 2001). This form of storage is known as mineral trapping.
Range of Estimates for CO$_2$ Sequestration in U.S. Geologic Formations

<table>
<thead>
<tr>
<th>Geologic Formation</th>
<th>Capacity Estimate (GtC)</th>
<th>Source</th>
</tr>
</thead>
<tbody>
<tr>
<td>Deep saline reservoirs</td>
<td>1-130</td>
<td>Bergman and Winter 1995</td>
</tr>
<tr>
<td>Natural gas reservoirs in the United States</td>
<td>25$^a$</td>
<td>R.C. Burruss 1977</td>
</tr>
<tr>
<td></td>
<td>10$^b$</td>
<td></td>
</tr>
<tr>
<td>Active gas fields in the United States</td>
<td>0.3 / year$^c$</td>
<td>Baes et al. 1980</td>
</tr>
<tr>
<td>Enhanced coal-bed methane production in the United States</td>
<td>10</td>
<td>Stevens, Kuuskraa, and Spector 1998</td>
</tr>
</tbody>
</table>

$^a$ Assuming all gas capacity in the United States is used for sequestration  
$^b$ Assuming cumulative production of natural gas is replaced by CO$_2$  
$^c$ Assuming that produced natural gas is replaced by CO$_2$ at the original reservoir pressure

Table 2.1. Capacity of geological targets of CO$_2$ storage in the United State. Deep saline reservoirs have the greater capacity for CO$_2$ storage than other geological storage options (after Byrer, 2002).
Figure 2.3. Distribution of sedimentary basins in the world suitable for CO$_2$ storage (after IPCC, 2005).
A number of sea-borne injection projects are presently ongoing such as those of the Norwegian company Statoil. One of their projects is at the Sleipner field in the Norwegian sector of the North Sea. About 1 million tonnes a year of recovered CO₂ from Heimdel reservoir is injected into the Utsira Formation from the Sleipner West platform, at a depth of about 800-1000 metres below the sea floor (Arts et al., 2004).

Compared with CO₂ sequestration in coal seams and depleted oil fields, sequestration of CO₂ in deep saline formations does not produce value-adding by-products, but it has other advantages. The sedimentary basins suitable for CO₂ storage distributed around the world (Figure 2.3) offer the most readily available option of all of the different geological scenarios, providing relatively easy access for the energy and industrial sectors from an injection point of view. Also, saline aquifers offer the largest capacity for CO₂ storage compared with the other options.

According to the final report of the GEODISC (Geological Disposal of Carbon Dioxide) project, the storage potential of Australian saline aquifers is adequate to store that country’s total CO₂ emissions for many hundreds of years at current rates of emission (Cook et al., 2001). According to a DOE (2007) report, the estimated carbon storage capacity of saline formations in the United States is large enough (up to 500 billion tonnes of CO₂) to make them a viable long-term solution for the U.S.

2.2 – Monitoring Techniques

The need for long term storage requires that monitoring techniques will need to ensure the environmental acceptability and safety of CO₂ storage in saline formations and other geological storage sites. Determining that CO₂ will not escape from formations by either migrating up to the surface or contaminating drinking water supplies is a key aspect of any sequestration research program on the CO₂ monitoring and quantification process.

Geophysics offers a variety of methods that operate over a wide range of geological environments, scales, and reservoir depths. Geophysical technologies represent the most
effective remote monitoring means to provide this assurance, which, coupled with reservoir engineering, geochemical fluid analysis, near-surface and atmospheric techniques, can provide the level of confidence required to verify that storage is effective. Quantitative assessment of the geophysical options includes investigations into changes in rock properties (sensitivity analysis and laboratory measurements); numerical models to simulate the process (space-time sampling, criteria and tradeoffs, monitoring tests on synthetic data sets), and tests with field data (including in-house time-lapse data sets from petroleum reservoirs, and available data sets from sequestration projects).

The development of a large acoustic contrast between CO$_2$ and the formation water as a result of CO$_2$ injection indicates that 3-D seismic techniques are well suited for monitoring CO$_2$ storage (Brevik et al., 2000) which can be tied to well log data to make it a trustworthy monitoring tool. However, other complementary methods may be required to effectively monitor the various trapping mechanisms of CO$_2$ in saline aquifers. Here I provide a review of some potential geophysical and non-geophysical technologies which may be used to monitor CO$_2$ storage in the subsurface. However, a detailed theoretical description of each method will not be done and the reader is referred to the relevant citations if further information is required. Since this thesis deals entirely with analysis of seismic methods, Chapter 3 will present the necessary theory.

2.2.1 – Seismic Methods

Cost-effective time-lapse seismic imaging that can provide continuous monitoring and adapt to changing reservoir conditions is one of the best options for CO$_2$ monitoring and verification (where verification implies quantification of fluids in place). An important principle to be followed is that the monitoring effort must decrease with time, and eventually stop when safe containment is no longer an issue.

Reservoir changes are detected by differencing the seismic snapshots recorded at different times. Different seismic methods are used in monitoring and verification. It has been suggested that in time-lapse cross-well tomography and vertical seismic profiling (VSP) methods, embedding both sources and receivers permanently will maximize
survey repeatability and reduce deployment costs. Surface seismic lines may be added or different sections of a VSP or tomographic survey may be recorded at different times to track the CO$_2$ front or to target specific reservoir zones or problems areas.

This seismic approach may include signal coding to permit the use of low-power sources for continuous operation which enables low cost and low energy real-time monitoring (Hartley, 2006). Seismic methods can be used at different scales depending on the required and achievable resolution. Surface seismic provides large lateral coverage but the vertical resolution can be restricted to tens of metres in the 2-D case and in the 3-D case it could be dependent on the quality of the recorded data. Conversely, sonic logging can provide resolution in the order of centimetres, but the necessary high frequency signal can often only penetrate a few metres into the formation from the borehole rather than the relatively large distances needed to sense the presence of CO$_2$. VSP and cross-well seismic are routinely used to bridge the gap between sonic logging and surface seismic methods. A rare method known as sonic reflection or single-well imaging, is a technology yet to fill the final gap between the different methods (McKenna, 2004). Figure 2.4 shows the resolution that each seismic method provides.

A combination of different seismic methods may be used to provide a satisfactory resolution and adequate coverage of the sequestration site.

2.2.1.1 – Surface Seismic
Repeated surface seismic measurements have been used over a number of oil reservoirs to monitor the efficiency of enhanced recovery during CO$_2$ injection (Miller et al., 2005) and also saline aquifers (Arts et al., 2004; McKenna, 2004) during CO$_2$ sequestration. These measurements have observed changes in compressional (P-) and shear (S-) wave velocity, attenuation and anisotropy, as well as seismic attributes extracted from changes in rock physics as a consequence of CO$_2$ flooding (Raef et al., 2005; Salehi and Gowelly, 2006).
Figure 2.5 (a) shows the velocity push-down caused by injecting CO₂ into formations in Weyburn Field- the largest of this kind in western Canada. The data is processed for this project which is performing both enhanced tertiary oil recovery and greenhouse gas sequestration (Li, 2003). In Figure 2.5 (b) change can be seen in seismic data amplitude over a period of 2 years of CO₂ injection via horizontal wells. The change in amplitude of reflection events is considerable and the largest changes are related to CO₂ accumulations in the direction of the horizontal injection wells (Li, 2003). Combined rock physics and reservoir simulations predicted the P-wave reflection amplitude of the reflection increased by 15 to 20% after CO₂ was injected into the formation (Li, 2003). The change in amplitude over time successfully imaged CO₂ migration within the reservoir. In particular, movement of CO₂ along high permeability streaks or fracture zones can be clearly seen (Figure 2.5(b)).

To date, changes in P-wave velocity and impedance in repeated surface seismic measurements have been interpreted to represent changes in either pore pressure or gas saturation (Li and Siggins., 2005; Landro et al., 2003). By contrast, seismic S-waves may be more sensitive to changes in pore pressure while remaining relatively insensitive to changes in fluid saturation. Long-offset surface seismic data may be used to interpret pressure prediction using amplitude variations with offset (AVO) analysis, as shown in Figure 2.6. Different seismic attributes have also been mentioned which are able to monitor CO₂ floods through a reservoir. According to Salehi et al. (2006), spectrally decomposed seismic data at different frequencies can be used to monitor and map the expansion of CBM or CO₂ fronts through coal seams for ECBM purposes.

A major issue in repeating seismic measurements over time is that both sources and receivers should be the same in time-dependent surveys to maximize survey repeatability and reduce deployment costs.
Figure 2.4. Schematic representation of the range and resolution provided by various forms of seismic data. Seismic resolution is directly related to the frequency content of the received signal (after Esmersoy et al., 1998).
Figure 2.5 (a). P-wave velocity push-down caused by CO$_2$ accumulation after 2 years (1999-2001) of injection in Weyburn Field (Canada). (b) Time-lapse amplitude difference map from horizon amplitudes extracted from the upper reservoir unit- porous Midale Marly dolomites were used for reservoir simulation in order to calculate the volume of the different CO$_2$ plumes underground (after Li, 2003).
2.2.1.2 – Vertical Seismic Profiles (VSPs)

Vertical seismic profiles (VSP) record signals between seismic sources located at the surface and receivers positioned in a borehole. This type of data can be used to tie surface seismic data in depth using borehole information and provides more accurate seismic measurements. A VSP generally provides higher frequency data than surface seismic methods because the seismic energy travels dominantly one way and therefore is not affected so much by signal attenuation as a result of passing through the rock matrix just once compared with the surface seismic data. Consequently the resolution of a VSP is usually appreciably better than that of surface seismic data (Sheriff and Geldart, 1982) and it has an improved signal-to-noise ratio compared with surface seismic. VSP can often provide a much higher resolution image of the immediate vicinity of the borehole.

VSP is recorded using different source-receiver geometries. Walk-away VSP is the ideal method for providing a seismic profile, in which the source locations are moved to large offsets from the well head, in order to provide data for the prediction of lateral and azimuthal variations in reservoir properties.

VSP measurements can use a smaller source providing higher frequencies than surface seismic because of the relatively short travel distances surrounding the well bore. 3-D VSP is a good way of studying fluid movements in a reservoir, but because of the high costs of performing VSP surveys, walkaway VSPs are rarely performed except in the extreme case where a knowledge of subtle geological features is required. It should be mentioned that VSP measurements have more repeatability than conventional surface seismic, since the receivers’ locations in a well bore during the recording process are measured more accurately than that of the receivers’ positions during marine surface seismic recording where the receivers are continually towed from one point to another.

VSP measurements have been shown to be the seismic method which is able to monitor CO₂ plumes of a relatively small size (Daley et al., 2005). Figure 2.7(a) shows VSP sections recorded at the Frio site (Daley et al., 2005) in which two-way travel time of the
Figure 2.6. Estimations of changes in (a) fluid saturation and (b) pore pressure using time-lapse surface seismic data. The left panel of each figure is the true saturation or pressure field from a flow simulator and the right panel is the estimated saturation of pressure field using time-lapse AVO analysis (after Tura and Lumley, 1999).
signals has increased due to CO\textsubscript{2} injection, while a very large increase in transmitted amplitude (80\%) was observed pre- and post- CO\textsubscript{2} injection (Figure 2.7(b)).

Micro-hole technology developed with DOE is a new low-cost VSP method which is becoming more practical as a reservoir imaging option. NETL’s Microhole Technology Initiative accelerated research and development of a suite of special miniaturized tools and sensors for drilling ultra small-diameter (3½-inches and smaller) boreholes with scaled-down, purpose-built coiled tubing rigs that are easily transportable (NETL, 2007). According to NETL (2007), a number of these microhole tools have now been successfully field-tested and the results show that significant cost savings, minimized risks, and reduced environmental impact (drilling waste can be decreased to as little as one-fifth that of conventional drilling) are achievable. Microhole technology may provide the means for recording a much clearer, sharper view of the subsurface in the future. Using microhole technology, VSP operators can justify the small expense of drilling microholes specifically fit for occupation by seismic sensors, thereby improving their ability to find by-passed pockets of oil and natural gas. Denser sampling (in terms of three times normal) could achieve much higher resolution in VSP methods.

2.2.1.3 – Cross-well seismic
Cross-well seismic surveys involve the sources and receivers in two different boreholes. Usually, several receiver locations in one well record data from a number of source locations in another well which have a few hundred metres of separation. One major issue in seismic measurements over time is that both source and receiver arrays should be the same in terms of type and location during time-lapse surveys. The best option is to permanently embed the sources and receivers to maximize survey repeatability and reduce deployment costs as suggested earlier.

Analysis of one-way travel times recorded for each source-receiver pair then uses ray theory to produce a velocity image or tomogram across the interval between the wells. Figure 2.8 shows the cross-well method with different ray paths. Similar to the VSP method, frequency content of data could be high due to the short ray path propagation.
Figure 2.7. (a) VSP records pre- and post- CO$_2$ injection in the Frio reservoir. (b) A change in two-way travel time due to the presence of CO$_2$ can be seen, but there is an 80% increase in amplitude after injection which could be used for volumetric assessments (after Daley et al., 2005).
distance. This can help in reservoir characterization through improved fault imaging and higher resolution stratigraphic interpretation (McKenna, 2004). Time-lapse cross-well seismic data were recorded during the supercritical CO$_2$ flood of an 1100 metres deep saline aquifer in Nagaoka in Japan (Xue et al., 2005). The cross-well tomography of the area between the monitor wells provided the location of the gas bubble. This was found to be in good agreement with the results of the 3-D simulation of the experiment (Sato et al., 2006). The change in P-wave velocity between the 3200 and 6200 tonne CO$_2$ injection can be seen in Figure 2.9.

2.2.1.4 – Multi-component Seismic

Multi-component, time-lapse seismology has great potential for monitoring fluid movements in reservoirs. The main reason is simply the presence of fluid-filled fractures. Shear waves (S-waves) are much more sensitive than compressional waves (P-waves) to the presence of fractures or micro-fractures and the fluid content within the fracture network (Benson and Davis, 2001).

Seismic shear wave anisotropy is a key to monitoring fluid property changes in fractured media. Seismic wave conversion in the contact of rock units causes two shear modes to form (S1 and S2) and to propagate with different velocities in different directions with respect to fracture directions. The faster mode (S1) propagates with its particle motion parallel to the open fracture direction, perpendicular to the minimum horizontal stress (S3) in the reservoir – a phenomenon called S-wave splitting, or birefringence.

The first time-lapse (4-D), multicomponent (9-C) seismic survey was acquired at Vacuum Field in Lea County, New Mexico (Benson and Davis, 2001). Shear wave (S-wave) and compressional wave (P-wave) seismic data were used to monitor the reservoir fluid property changes associated with a CO$_2$ tertiary flooding within the carbonate host rock. Reservoir fluid properties, including viscosity, density, saturation and pressure changes, occur in response to CO$_2$ injection. These fluid property changes alter the interval velocity and attenuation of S-waves passing through the reservoir interval by up to 10 percent, but cause little (1 to 2 percent) or no measurable change in P-wave velocity
Figure 2.8. Cross-well methods. (a) Sources and geophones in nearby bore holes provide straight ray paths. (b) Ray-path bending and concentration of energy across high velocity cells can leave many cells poorly sampled. (c) Velocities are then determined for the actual situation shown in part. (b) (after Sheriff and Geldart, 1982)
Figure 2.9. Seismic cross-well methods in Nagaoka field, Japan. (a) Sources and geophone geometry. (b) P-wave velocity difference tomograms between observation wells at a distance of 160 m after 3,200 tonne- CO₂ injection and after 6,200 tonne- CO2 injection (after Xue et al., 2005).
and attenuation on the surface seismic data (Benson and Davis, 2001).

The Weyburn Field in Canada discussed earlier, has also been investigated using multicomponent seismic methods to monitor CO₂ injection for EOR purposes. According to Davis et al. (2003), rock and fluid physics measurements and modelling suggested that carbon dioxide should cause a 4-6% decrease in P-wave velocity with a corresponding 15-20% change in reflection amplitude and shear-wave velocity. Modelling suggested that where fracture zones occurred, the shear wave velocity anisotropy could change by 5-10%. The P-wave amplitude difference map between 2000 and 2001 is shown in Figure 2.10 (a). This map shows the CO₂ movement along fracture zones. The S-wave amplitude difference map shows some large anomalies as well. Based on these two maps, fracture zones caused channelling of CO₂ away from the injection pattern. Using these data and the S-wave splitting map (Figure 2.10 (c)), a classification for anisotropy was predicted (Figure 2.10 (d)).

2.2.1.5 – Passive Seismic

Different data are sensitive to different property changes caused by CO₂ injection. Surface seismic data show the reflectivity changes. VSP tomography gives detailed velocity models that show the pressure and saturation changes associated with CO₂ injection.

Seismic activities which do not involve any controlled sources, are called passive seismic. Microseismic locations can indicate the fluid front. Joint passive and active seismic tomography and imaging can provide a more reliable monitoring method at a relatively low cost and is a way to maximize the use of field data (Quan and Harris, 2006).

Within a reservoir, activities such as fluid flow within a fracture system, can produce passive seismic energy due to periodic strain accumulation and can occur during injection into or extraction of a fluid from a reservoir, and rock breakage during hydraulic fracturing (which is an option for increasing the permeability of a reservoir) may create microseismic events (Sheriff and Geldart, 1982). Seismicity data has been used in coal
Figure 2.10. (a) P-wave amplitude difference map. (b) S-wave amplitude difference map. (c) Time-lapse changes in amplitude derives shear-splitting where arrows indicate fast shear wave propagation directions. (d) Anisotropy classification based on multi-component seismic survey (after Davis et al., 2003).
Figure 2.11. CO$_2$ injection underground can cause microseismic activity for different processes. Each process is shown schematically in this figure (after Sminchak et al., 2001).
mining as a hazard or rock fall indicator during underground mining particularly over the last 15 years (Hatherly and Evans, 2006).

If the permeability and porosity of a reservoir’s rock is so low that the fluids do not communicate easily, a pressure build up may result when injecting CO$_2$ into the reservoir and the reservoir could become over-pressurised. Increasing pore pressure resulting from CO$_2$ injection can reduce the frictional resistance along existing fault planes and cause fault blocks to slip, or even may change loose rock unit contacts to form new faults. Both sliding fault blocks and newly forming faults can induce seismicity by releasing seismic energy as the fault moves. Figure 2.11 (Sminchak et al., 2001) shows schematically the processes which may cause passive seismicity.

Recording passive seismic events in conjunction with surface seismic as a 3-D time-lapse study has been shown to be suitable for fluid flow monitoring during the production or injection process in low permeability chalks (Hornby et al., 2006). In passive seismic surveys, three component geophones should be permanently embedded in a position (triaxial sensors in the case of borehole measurements) to give high repeatability of recordings and reduce noise in order to be able to determine the raypath orientation of the arrivals. The microseismic monitoring of fractures, as a key factor in the selection of an optimum well location, is another use of passive seismic methods (Lappierre, 2006).

In the passive seismic method, continuous signals are recorded and analysed using earthquake detection algorithms to determine when an impulsive energy source has occurred and the event location is calculated at the point in space that matches the observed arrival times of different phases and ray path orientations recorded by the triaxial sensors (McKenna, 2004). To invert time data to depth positions, detailed velocity models can be constructed from sonic logs.

The intrinsic seismic activity rate depends on the strength of the rock and the strained energy stored in the rock. Mohr-Coloumb failure criteria may be used to estimate the likelihood of CO$_2$ injection triggering fault reactivation; however, many assumptions
regarding fault stress and pressure need to be utilised and this produces less certainty compared to direct evidence from passive seismic monitoring (Sminchak et al., 2001).

2.2.2 – Other monitoring methods

Non-seismic methods include microgravity, electromagnetic (EM) and self potential (SP) methods which have been used in CO₂ monitoring projects (Gasperikova and Hoversten, 2006). These methods have different resolution scales but may offer significant cost savings (Caruso and Milanese, 2004). The feasibility of each geophysical technique depends on the magnitude of the change in the measured geophysical property produced by increasing the concentration of CO₂, and on the inherent resolution of the technique. Noble gas chemical tracers, satellite-based detectors which show deformations caused by pressure change due to CO₂ injection, and atmospheric CO₂ measurements are other non-seismic methods which make us able to predict the CO₂ movement within a reservoir. In this section on the application of non-seismic methods, I provide a brief review for completeness of the thesis since the thrust of my research was not associated with such methods.

2.2.2.1 - Micro-Gravity

Gravity recording methods respond to changes in the density of the Earth. For most of the depth interval of interest for sequestration, the gravity method offers a reasonable approach for observation of CO₂ injection into brine or oil-bearing formations since CO₂ is less dense than brine or oil, so the density contrast between CO₂ and brine may produce an observable gravity anomaly when CO₂ is stored in a saline aquifer. Shallow reservoirs, large porosity, and a large volume of CO₂ are the best situation for the success of surface micro-gravity measurements.

The micro-gravity method was successfully applied in CO₂ monitoring through an EOR project in Bluff field, Alaska, USA (Gasperikova and Hoversten, 2006). In this example, borehole gravity measurements in conjunction with surface measurements guided the interpolation of gravity measurements between wells. According to Gasperikova and Hoversten (2006) there is a large scale relationship between the vertical components of
the measured gravity field within a CO₂ saturated reservoir, which can provide a good time-lapse view of the CO₂ movement within the reservoir.

A plan view of a numerical model of the density field for the top of the Bluff field reservoir can be seen in Figure 2.12 (a). The modelled surface vertical component of gravity measurements was in good agreement with the density field (Figure 2.12 (b)) because a 20 µGal gravity response could be observed from the CO₂ brine interface (within a 20-m reservoir at 1900 m depth), in which there was 30% CO₂ and 70% brine, compared to the model without CO₂. In this case, changes in the CO₂ saturation of 10% produced changes in the vertical component of gravity of approximately 6 µGal, which would be the limit of detectability (which was 4 µGal). Lateral movement of the CO₂ - brine front of some 500 m from the changes in the vertical component of gravity could also be observed (Gasperikova and Hoversten, 2006).

The low cost of onshore gravity acquisition makes it a good method which could be integrated with more expensive seismic data to monitor CO₂ distribution underground.

Offshore gravity data has also been recorded successfully over the Troll gas filed (Eiken et al., 2000). These measurements are more expensive and difficult to acquire than those onshore.

Figure 2.13 (a) shows the simulated change in density caused by an 18 month CO₂ injection into Otway gas field in Australia in order to enhance gas production (Toomey et al., 2005). In this example, supercritical CO₂, which has a density higher than methane but lower than water, replaced formation water. This would result in a net decrease in density, while replacing methane with CO₂ results in a net increase in density.

The result is a change in the gravity response. As can be seen in this figure, the maximum density change occurs in the CO₂ layer underneath the residual methane cap in the crestal structure. The simulation results show that monitoring of CO₂ movement due to a density change is possible. Gravity inversion results confirm that CO₂ injection
Figure 2.12. (a) Plan view of the density field model at top of the reservoir for Bluff field for CO$_2$ having a water saturation of 30% and 70%, respectively. (b) Surface vertical component of gravity has shown the density change very well (after Gasperikova and Hoversten, 2006).
causes an increase in density at the injection point and a reduction in density at the production well (Figure 2.13 (b)).

Within this modelling, there was no account for the effect of a dry or wet season (water table) prior to or during injection. The area is covered with near-surface karst limestones causing caverns and sink holes, and if such gravity measurements were taken in the dry season then in the wet, there could be a substantial difference in the gravity response irrespective of the amount of injected CO$_2$. Therefore, gravity measurements would not necessarily be useful at this site.

**2.2.2.2 – Electromagnetics**

Electrical methods primarily respond to earth material resistivity. Since CO$_2$ is resistive, electrical methods are candidates for monitoring CO$_2$ in brine-bearing formations. The electrical resistivity of reservoir rocks is more sensitive to changes in water saturation than P-wave velocity, as can be seen in Figure 2.14 (a). Archie’s law describes the electrical resistivity of sedimentary rock as a function of water saturation ($S_w$), porosity ($\phi$), and pore fluid resistivity ($\rho_{\text{brine}}$) (Gasperikova and Hoversten, 2006). All petroleum fluids (oil, condensate, and hydrocarbon gas) as well as CO$_2$ are electrically resistive- hence, Archie’s law is appropriate for any combination of oil, hydrocarbon gas, condensate, or CO$_2$. In Figure 2.14 (b) the bulk resistivity is plotted on a log scale to span the large range of resistivity values as a function of the gas saturation.

The cross-well electromagnetic (EM) method is designed to map the inter-well resistivity distribution in a 2-D (or 3-D) sense. These data can be used to characterize reservoirs structurally and stratigraphically as well as to track on-going processes where pore fluid is replaced or moved. A cross-well EM field system consists of a transmitter tool deployed in one well and a receiver tool deployed in a second well located up to 1000 m from the source well (Figure 2.15 (a)).
Figure 2.13. (a) Cross section of density change after 18 months of CO$_2$ injection in Otway basin. Supercritical CO$_2$ is injected from a deeper point in a well to the right side and a mixture of CO$_2$ and methane is extracted up-dip from the left well. (b) Plan view of the area shows a reduction in density due to the methane movement towards the production well and an increase in density around the injection well due to CO$_2$ injection (after Toomey et al., 2005).
EM techniques respond as a function of the rock bulk electrical resistivity, due to the electrical sensitivity to water saturation. Hovorka (2005) used cross-well EM surveys at the Frio injection site for pre- and post CO$_2$ injection. Her analyses to date showed success in measuring the subsurface distribution of CO$_2$. Wilt (2003) used the EM method to map the CO$_2$ flood in the Vacuum oil field carbonates in West Texas, USA. As can be seen in Figure 2.15 (b) the inter-well resistivity image is in good agreement with the porosity log image in dip and overall structure of the reservoir. The resistivity section derived from the cross-well EM data replicates the log-based resistivity but at a lower resolution. Superimposing the log based resistivity and the cross-well resistivity difference sections (Figure 2.16 (a)) may help explain the nature of the arcuate-shaped zone of decreasing resistivity (Figure 2.16 (b)) and also could be used in quantification of water saturation, porosity, etc.

Passive EM monitoring has also been investigated as a monitoring method in the Lost Hills oil field to detect seismicity (Kirkendall and Roberts, 2003). Figure 2.17 shows both seismic and an EM signal recorded in response to a hydrofracture event. As can be seen in this figure, similar to the passive seismic method, if multiple receivers were used it may be possible to locate the position of any fracturing.

2.2.2.3 – Near Surface Monitoring

Techniques which measure CO$_2$ migrating upward out of the storage formation and towards the surface, are called near surface monitoring methods. Ground and formation water sampling and soil gas sampling are the main techniques used in near surface CO$_2$ monitoring. To detect any chemical change caused by CO$_2$ migration out of the target formation, it is required to do a base line survey before CO$_2$ injection. According to the IPCC (2005) report, an initial concentration of CO$_2$, HCO$_3^-$, CO$_3^{2-}$, and trace elements will help to quantify the solubility and mineral trapping of CO$_2$, and study the CO$_2$⁻ water-rock interactions, as well as detect any CO$_2$ leakage into shallow aquifers.
Figure 2.14. (a) Comparison between P-wave velocity and resistivity on brine saturation. Resistivity is far more sensitive to brine saturation than P-wave velocity (after Kirkendall and Roberts, 2003). (b) Archie’s equation can be used to predict the resistivity of a rock as a function of gas saturation, porosity and density (after Gasperikova and Hoversten, 2006).
Figure 2.15. (a) Schematic of cross-well seismic. In one well the EM transmitter transmits the EM waves and secondary field produced by induced electromagnetic waves in the reservoir’s rocks, is received in the other well. (b) The resistivity and porosity sections agree in the dip direction and overall structure in this part of the field and the zones of highest porosity (red on plots) match with the higher resistivity (red-yellow) horizons in the resistivity section (after Wilt, 2003).
Figure 2.16. (a) Resistivity and time-lapse resistivity difference cross-section between two wells in Vacuum oil field. (b) Superimposed section made of log-based resistivity and cross-hole EM sections introduced the possibility of faults in the reservoir (after Wilt, 2003).
Near surface CO$_2$ monitoring methods involve geochemical monitoring, chemical tracer detection and borehole measurements which will now be briefly discussed.

### 2.2.2.3 – Geochemical Monitoring

Direct geochemical measurements can clarify the nature and help quantify the fluid-rock interaction over time. Geochemical measurements can be performed using both surface (to observe any potential leakage to the soils on the surface) and long-term mineral alteration using core plugs taken at regular intervals in the storage reservoir.

Geochemical measurements coupled with water chemical tests and a study of the natural and introduced tracers may help to determine the effectiveness and efficiency of the various CO$_2$ trapping mechanisms. Geochemical CO$_2$ monitoring may involve collecting soil-gas samples, measuring CO$_2$ soil flux, and calculating the CO$_2$ soil-gas concentration profiles with depth (Doughty et al., 2005).

Man-made structures such as damaged or abandoned wells and natural geological features such as faults and outcrops are two types of potential leakage pathways which are most suitable for geochemical sample collection. Since CO$_2$ is not expected to reach the surface, it is also critical to do pre CO$_2$ baseline monitoring to identify the CO$_2$ signature of naturally-occurring processes (Doughty et al., 2005). Geochemical sampling at the monitor well will provide direct evidence of breakthrough of CO$_2$ and tracers. Moreover, CO$_2$ charged water acts as a weak acid, and reacts to some extent with the minerals in the aquifer, producing a distinct chemical signature in the water collected at the monitoring well.

### 2.2.2.4 – Tracers

CO$_2$ storage site-characterization activities can include a tracer test, employing a conservative solute (fluoresce) in a balanced doublet (injection rate at the injection well equals pumping rate at the monitor well) to create a steady flow-field between the injection and monitor wells (Doughty et al., 2005). Tracers are used to indicate fluid travel time between injection and monitoring wells by introducing them into the injection
Figure 2.17. Passive EM and seismic data recorded during a hydrofracture event in the Lost Hills oil reservoir. The picture shows potential for passive EM signals to monitor microseismicity via the seismoelectric effect (after Kirkendall and Roberts, 2003).
stream. The source of CO$_2$ measured in geochemical samples could be identified by tracing the CO$_2$ movement within the storage formation, quantifying solubility trapping and tracing any leakages (Gibson-Poole et al., 2006). Tracers may be naturally occurring (isotopes of carbon, hydrogen, oxygen, and noble gasses often associate with CO$_2$) or introduced into the injection flow (such as noble gasses, SF$_6$, and fluorocarbons).

By adding a tracer to all fluids that are introduced into the subsurface, collected samples will contain an easily detectable signature, facilitating a reconstruction of fluid movement and interactions. In down-hole sampling, the samples are collected using a U-tube which sustains the pressure of the interval in which the sample has been taken. The main criterion for sampling methodology is to allow accurate re-creation of down-hole conditions. Real-time tracer analysis can be performed at the surface using a portable mass spectrometer.

In a tracer-test breakthrough curve, the three main attributes are: the time of first arrival, the peak concentration, and the overall shape of the curve being measured. The time of the first arrival determines the velocity through the fastest flow paths. Also the overall shape of the breakthrough curve provides information on the distribution of flow paths (for example, uniform versus strong preferential flow).

2.2.2.5 – Bore-hole Measurements
Conventional geophysical measurements recorded as logs in wells provide very high resolution data from physical properties of the reservoir which are directly obtained from the reservoir zone. Wire-line logging, which routinely is used by petrophysicists, has been mentioned as an essential tool in monitoring of CO$_2$ injection (Hovorka et al., 2006) and usually provides a detailed geological description for other monitoring techniques. Repeated wire-line measurements (such as resistivity measurements) can be used to detect the CO$_2$ arrival at bore-holes, in order to calculate saturation and interactions with the reservoir. The presence of CO$_2$ was also identified using induction, sonic and neutron logging of the reservoir at the observation well in Nagaoka field in Japan (Xue et al., 2005).
A combination of temperature and pressure data within a well can provide an accurate indication of CO$_2$ state-of-phase. Down-hole pressure can be measured using permanent pressure gauges in an observation well while temperature can be achieved using temperature logs. Flow meters may provide an indication of the CO$_2$ migration rate and help long term flow prediction.

Density and neutron logs can provide an estimation of formation porosity, so repeated measurements may indicate changes in porosity caused by geochemical interactions between CO$_2$ and the formation mineralogy. Resistivity logs are also sensitive to water saturation and can be used in CO$_2$ saturation prediction. Changes in CO$_2$ saturation may be obtained by repeated measurements using resistivity logs.

2.3 - Summary of This Chapter
This chapter has reviewed some potential sites for geological storage of CO$_2$ and discussed monitoring techniques, including seismic and non-seismic methods, which have been successfully used to observe the long-term CO$_2$ displacement with brine in the formation and its movement underground. In each CO$_2$ monitoring program, the technical feasibility at each particular site and the associated cost of the various options have to be considered in order to optimise the best methods for monitoring and verification.

Since my research was focused on the seismic method of monitoring dissolved CO$_2$ using controlled repeated ultrasonic experiments on a physical model, therefore the following chapter will now present a more in-depth review of the relevant seismic theory.
3.1 – Seismic Theory
The experiments to be performed would involve recording seismic transmission and reflection data over and through a physical model representing a CO$_2$ storage reservoir. Repeated transmission experiments were considered as an appropriate tool to observe the seismic response to dissolved CO$_2$ in distilled water. This approach was based on the fact that dissolving different quantities of CO$_2$ in known volumes of water would probably cause a density and bulk modulus change in the water, which in turn could cause a change in seismic velocity and amplitude variation (attenuation) or consequently a change in quality of transmission. Comparing the experiments through a matrix filled with different mixes of fluids to the experiments without a matrix would allow a study of seismic attenuation as well as the scattering effect of seismic wave transmission due to the presence of a matrix. This chapter reviews the relevant seismic theory required for my analysis of the data recorded during changes in pressure and dissolved CO$_2$, to achieve my research objectives.

3.1.1 – Elastic Wave Propagation in porous rocks
Seismic waves propagate through the Earth as elastic waves. Elastic waves are propagated through solid bodies by the transfer of forces that act upon them as an elastic deformation of a body. The equation of wave propagation in elastic solids is derived by combining Hook’s law and Newton’s law of motion (Sheriff and Geldart, 1982).

3.1.2 – Stress-Strain and Hook’s Law
An elastic body will be deformed by a force acting upon it. Stress is defined as force per unit area, and thus when a force is applied to a body the stress is the ratio of the force to the area over which the force is applied. If the force is perpendicular to the area, the stress is called “Normal stress” and when the force is tangential to the area, the stress is “Shear stress”. When the force is neither parallel nor perpendicular to any element of stress on
the surfaces, this force can be resolved into components of normal and shearing stresses
(Sheriff and Geldart, 1982). Figure 3.1 illustrates the components of stress acting on an
infinitesimally small volume surrounding a point within an elastic solid. The normal and
shear stresses can be seen on the surfaces of the body. For the cube to retain its shape,
nine stress components must act on the volume. These components formulate the second
order stress tensor (Nye, 1985):

\[
\sigma_{ij} = \begin{bmatrix}
\sigma_{xx} & \sigma_{xy} & \sigma_{xz} \\
\sigma_{yx} & \sigma_{yy} & \sigma_{yz} \\
\sigma_{zx} & \sigma_{zy} & \sigma_{zz}
\end{bmatrix} .
\] (3.1)

The diagonal elements of the stress tensor are the normal stress components and the off-
diagonal elements are shear stress components. The sum of the moments acting on each
surface of the cube must be zero (Yilmaz, 2001), which causes \(\sigma_{ij} = \sigma_{ji}\) and the tensor
becomes symmetrical about the main diagonal.

Deformation of a body due to stress would be described as an elastic strain, when
changes in shape and dimensions occur. It assumes that no permanent deformation of the
rock particles occurs along the propagation path of seismic waves which are considered
as elastic waves. The relative change in a dimension of a body in length due to normal
stress is referred to as “Normal strains”. The change in shape of the medium caused by
shearing stress is known as “Shearing strain”. The strain that a unit cube undergoes in
response to stress can also be expressed in components, as shown in Figure 3.2. These
three-dimensional components form the second order strain tensor (White, 1983):

\[
\varepsilon_{ij} = \begin{bmatrix}
\varepsilon_{xx} & \varepsilon_{xy} & \varepsilon_{xz} \\
\varepsilon_{yx} & \varepsilon_{yy} & \varepsilon_{yz} \\
\varepsilon_{zx} & \varepsilon_{zy} & \varepsilon_{zz}
\end{bmatrix} .
\] (3.2)

If the dimensions of the unit cube become infinitesimally small, \(\varepsilon_{xy} = \varepsilon_{yx}\), \(\varepsilon_{xz} = \varepsilon_{zx}\), and
\(\varepsilon_{yz} = \varepsilon_{zy}\) (Yilmaz, 2001). Thus, the stress tensor is also symmetrical about the main
diagonal.

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Figure 3.1. Forces acting on a unit cube within a homogeneously stressed body. The first index $i$ defines the direction of the stress and the second component $j$ defines the plane on which the forces act (after Sheriff and Geldart, 1982)).
Figure 3.2. Types of strain on a unit cube made by different stress types (after White, 1983).
Stresses acting on an elementary volume of an elastic solid can be expressed as a linear combination of strains (White, 1965). The mechanical behaviour of a linear elastic material can be described by Hooke’s law. The generalised Hooke’s law for a linear elastic body states that stress and strain can be connected through a fourth-order tensor $C_{ijkl}$ (Mavko et al., 1998);

\[ \sigma_{ij} = C_{ijkl} \varepsilon_{kl} \]

or

\[ \varepsilon_{ij} = S_{ijkl} \sigma_{kl} \quad \text{i,j,k,l= 1,2,3} \]  

where $\sigma_{ij}$ is stress, $\varepsilon_{kl}$ is strain, $C_{ijkl}$ is the fourth-order stiffness tensor and $S_{ijkl}$ is the elastic compliance tensor.

In practice we cannot measure strains however we can relate stress to displacement $(u)$ rather than strains:

\[ \sigma_{ij} = C_{ijkl} \frac{\partial u_k}{\partial x_j} \]  

Since the stress and strain tensors are symmetrical, the number of independent stress and strain components is reduced to six. These tensors are typically expressed in matrix form (White, 1983):

\[
\sigma = \begin{bmatrix}
\sigma_{xx} & \sigma_{xy} & \sigma_{xz} \\
\sigma_{yx} & \sigma_{yy} & \sigma_{yz} \\
\sigma_{zx} & \sigma_{zy} & \sigma_{zz}
\end{bmatrix} \rightarrow \begin{bmatrix}
\sigma_{xx} \\
\sigma_{xy} \\
\sigma_{xz}
\end{bmatrix}
\]

and
However, some of these components are independent and stiffness tensor $C_{ijkl}$ is invariant under permutation of indices $i$ and $j$ as well as $k$ and $l$ (Federov, 1968). The symmetry of stresses and strains satisfy the condition that $C_{ijkl} = C_{jikl} = C_{ijlk} = C_{jilk}$ and also due to strain energy consideration (adiabatic process) $C_{ijkl} = C_{klji}$ the eighty-one component reduces the total number to thirty-six independent components.

The simplification of the stiffness tensor $C_{ijkl}$ reduces its four subscripts $3 \times 3 \times 3 \times 3$ to two, allowing it to be replaced by a $6 \times 6$ matrix (Mavko et al., 1998):

$$
C_{ij} = \begin{bmatrix}
    c_{11} & c_{12} & c_{13} & c_{14} & c_{15} & c_{16} \\
    c_{12} & c_{22} & c_{23} & c_{24} & c_{25} & c_{26} \\
    c_{13} & c_{23} & c_{33} & c_{34} & c_{35} & c_{36} \\
    c_{14} & c_{24} & c_{34} & c_{44} & c_{45} & c_{46} \\
    c_{15} & c_{25} & c_{35} & c_{45} & c_{55} & c_{56} \\
    c_{16} & c_{26} & c_{36} & c_{46} & c_{56} & c_{66}
\end{bmatrix}.
$$

The stiffness matrix is symmetric about the main diagonal, and therefore only comprises of twenty-one independent components. Thus, in matrix form and expressed in terms of displacement, Hook’s law simplifies to:
The symmetry of physical properties of a medium can reduce the number of stiffness components used for describing the media (Mavko et al., 1998).

### 3.1.3 – The Elastic Wave Equation

If stresses on the rear face of the element of volume are as shown in Figure 3.3, the stresses on the front face are respectively,

$$\sigma_{xx} + \frac{\partial \sigma_{xx}}{\partial x} \, dx, \quad \sigma_{yx} + \frac{\partial \sigma_{yx}}{\partial x} \, dx, \quad \sigma_{zx} + \frac{\partial \sigma_{zx}}{\partial x} \, dx. \quad (3.9)$$

Because these stresses are opposite to those acting on the rear face, the net (unbalanced) stresses are:

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

These stresses act on a face having an area \((dy \, dz)\) and affect the volume \((dx \, dy \, dz)\); hence, we get for the net forces per unit volume in the directions of the x-, y-, and z axis the respective values:

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

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

$$\begin{bmatrix}
\sigma_{xx} \\
\sigma_{yx} \\
\sigma_{zx}
\end{bmatrix} =
\begin{bmatrix}
c_{11} & c_{12} & c_{13} & c_{14} & c_{15} & c_{16} \\
c_{12} & c_{22} & c_{23} & c_{24} & c_{25} & c_{26} \\
c_{13} & c_{23} & c_{33} & c_{34} & c_{35} & c_{36} \\
c_{14} & c_{24} & c_{34} & c_{44} & c_{45} & c_{46} \\
c_{15} & c_{25} & c_{35} & c_{45} & c_{55} & c_{56} \\
c_{16} & c_{26} & c_{36} & c_{46} & c_{56} & c_{66}
\end{bmatrix}
\begin{bmatrix}
\frac{\partial u_x}{\partial x} \\
\frac{\partial u_y}{\partial x} \\
\frac{\partial u_z}{\partial x}
\end{bmatrix} =
\begin{bmatrix}
\frac{\partial u_x}{\partial z} + \frac{\partial u_z}{\partial y} \\
\frac{\partial u_y}{\partial z} + \frac{\partial u_z}{\partial x} \\
\frac{\partial u_z}{\partial y} + \frac{\partial u_y}{\partial x}
\end{bmatrix}. \quad (3.8)
Figure 3.3. Unbalanced stress acting on a unit cube along $x$-axis (after White, 1983).
\[ \frac{\partial \sigma_{xx}}{\partial x} + \frac{\partial \sigma_{xy}}{\partial y} + \frac{\partial \sigma_{xz}}{\partial z} . \quad (3.10) \]

Newton’s second law of motion states that the sum of all forces \( f \) acting on a body in an arbitrary direction \( i \), is equal to the product of the mass \( m \) and acceleration \( a \) of the body:
\[ f_i = ma . \]

Figure 3.3 shows that the force applied to a unit volume can be expressed in terms of applied stress:
\[ f_i = \frac{\partial^2 \sigma_{ij}}{\partial x_j} . \quad (3.11) \]

Newton’s second law can be expressed in terms of the displacement \( u \) of the unit volume (equation of motion along the \( x \)-axis):
\[ f_i = \rho \frac{\partial^2 u}{\partial t^2} = \rho \left( \frac{\partial \sigma_{xx}}{\partial x} + \frac{\partial \sigma_{xy}}{\partial y} + \frac{\partial \sigma_{xz}}{\partial z} \right), \quad (3.12) \]

in which \( \rho \frac{\partial^2 u}{\partial t^2} \) is unbalanced force in the \( x \)-direction on a unit volume and \( \rho \) is the density (assumed to be constant). Similar equations can be written for the motion along the \( y \)- and \( z \)-axis.

Combining equations (3.11) and (3.12) allows Newton’s second law to be expressed in terms of stress and displacement:
\[ \rho \frac{\partial^2 u}{\partial t^2} = \frac{\partial^2 \sigma_{ij}}{\partial x_j} . \quad (3.13) \]

Substituting equation (2.2) into (2.13) defines the elastic wave equation (Mavko et al., 1998):
\[ \rho \frac{\partial^2 u_k}{\partial t^2} = C_{ijkl} \frac{\partial}{\partial x_j} \left( \frac{\partial u_k}{\partial x_l} \right) = C_{ijkl} \frac{\partial^2 u_k}{\partial x_j \partial x_l} . \] (3.14)

### 3.1.4 – Isotropic Medium

Isotropic symmetry of the stiffness tensor is the simplest form of symmetry class for an elastic medium. In this case, P- and S-wave velocities do not vary with the direction of wave propagation. This simplifies the elastic stiffness matrix \( C_{ij} \) from twenty-one independent constants to only two (Mavko et al., 1998):

\[
C_{ij} = \begin{bmatrix}
    c_{11} & c_{12} & c_{12} & 0 & 0 & 0 \\
    c_{12} & c_{11} & c_{12} & 0 & 0 & 0 \\
    c_{12} & c_{12} & c_{11} & 0 & 0 & 0 \\
    0 & 0 & 0 & c_{44} & 0 & 0 \\
    0 & 0 & 0 & 0 & c_{44} & 0 \\
    0 & 0 & 0 & 0 & 0 & c_{44}
\end{bmatrix} , \quad C_{12} = C_{11} - 2C_{44} . \] (3.15)

The two independent constants, \( C_{11} \) and \( C_{44} \) can be related to the bulk \( (K) \) and shear \( (\mu) \) moduli of a rock (Mavko et al., 1998):

\[
C_{11} = K + \frac{4}{3} \mu ,
\]

\[
C_{44} = \mu
\]

For seismic compressional wave propagation along the \( x \)-axis, the direction of particle displacement is also along the \( x \)-axis. This simplifies equation (3.14) to:

\[
\rho \frac{\partial^2 u_x}{\partial t^2} = \left( K + \frac{4}{3} \mu \right) \frac{\partial^2 u_x}{\partial x_x} . \] (3.16)

Rearrangement gives the P-wave velocity for a seismic wave travelling in an isotropic medium as:

\[
\frac{\partial u_x}{\partial t} = \sqrt{\frac{K + \frac{4}{3} \mu}{\rho}} . \] (3.17)
For a seismic shear wave propagation in the $x$-direction, causing particle displacement in the $y$-direction, equation (3.14) becomes:

$$\rho \frac{\partial^2 u_x}{\partial t^2} = \mu \frac{\partial^2 u_y}{\partial x^2}. $$

Rearrangement gives the S-wave velocity for a seismic wave travelling in an isotropic medium as:

$$\frac{\partial u_x}{\partial t} = \sqrt{\frac{\mu}{\rho}}. $$

Therefore, only estimates of bulk modulus ($K$), shear modulus ($\mu$) and density ($\rho$) are required to predict P-wave ($V_p$) and S-wave ($V_s$) velocities in homogeneous isotropic elastic media.

### 3.2 – Rock Physics

The rate at which elastic waves propagate through elastic solid bodies depends on the elasticity of the material. In the Earth, the elasticity of porous rocks will be dependent on mineralogy, compaction history, porosity, cementation, pressure, temperature, and the fluid contained within the available pore space (McKenna, 2004). Rock physics analysis is used to predict the elastic properties of porous rocks from knowledge of the rock properties.

#### 3.2.1 - Estimating Dry Frame Elastic Moduli

The dry frame elastic moduli are essential to calculate the effective moduli of a saturated rock with different fluids using Gassmann’s equation. In this chapter, theories for calculating the dry moduli of a rock frame from an empirical equation will be explained. Estimating the dry moduli for a granular matrix and cemented sand is discussed.

#### 3.2.1.1 – Empirical Relationships

Estimation of the elastic moduli for both the grains and the pores are required to predict the dry elastic moduli of a porous rock. Mavko et al. (1998) presented an equation for the
dry frame elastic moduli estimation:

\[ K_{dry} = K_{grain} (1 - \beta) \quad , \quad (3.19) \]

where \( \beta \) is referred to as Biot’s coefficient and represents the compressibility of the pore space. As the porosity of a rock increases, the more compliant the rock will become. For most porous materials there is a critical porosity \( (\phi_c) \) which separates their mechanical and acoustic behaviour into two distinct domains (Mavko et al., 1998). For porosities lower than \( \phi_c \), the grains are load bearing, while for porosities higher than that, the fluid is load bearing (suspension domain). In the suspension domain, the effective elastic moduli typically follows the Reuss lower bound of the fluid-mineral mixture in the load bearing domain, the effective elastic moduli lie between the upper and lower Hashin-Shtrikman bounds of the fluid-mineral mixture (Mavko et al., 1998). This can be seen in Figure 3.4.

Since the porosity of most of the reservoir rocks is typically well below \( \phi_c \), a number of empirical relationships have been developed to predict the elastic moduli of dry sandstones from porosity.

Nur et al. (1995) introduced a formula for estimating dry bulk and shear moduli of porous rocks using a linear trend between the zero porosity and critical porosity limits for sandstone:

\[
\begin{align*}
K_{dry} &= K_{grain} \left(1 - \frac{\phi}{\phi_c}\right) \\
\mu_{dry} &= \mu_{grain} \left(1 - \frac{\phi}{\phi_c}\right)
\end{align*}
\]

(3.20)

These formulas seems to be a modified version of the Voigt average where one end member is the effective elastic moduli of the grains and the other end member is the effective elastic moduli of a suspension (Mavko et al., 1998).

Geertsma (1961) presented his empirical relationship as:
Figure 3.4. Mechanical properties of porous media divides into two distinct domains with regard to critical porosity, $\phi_c$. For porosities lower than this, the rock frame is load-bearing and for porosities greater than $\phi_c$ the fluid is load-bearing (after Mavko et al., 1998).
\[
\frac{1}{K_{\text{dry}}} = \frac{1}{K_{\text{grain}}} (1 + 50\phi) \quad .
\] (3.21)

### 3.2.1.2 – Granular Theory

The effective elastic properties of packing of spherical particles depend on normal and tangential contact stiffness of a two-particle combination (Figure 3.5) (Mavko et al., 1998):

\[
S_n = \frac{\partial F}{\partial \delta} \quad .
\]

\[
S_\tau = \frac{\partial T}{\partial \tau} \quad .
\] (3.22)

For a random packing of spherical grains, the effective dry bulk and shear moduli of a medium with porosity of \( \phi \), coordination number of \( C \), normal and tangential stress \( S_n, S_\tau \) of a two-sphere, and radius of the grains \( R \), can be calculated as:

\[
K_{\text{eff}} = \frac{C (1 - \phi)}{12 \pi R} S_n
\] (3.23)

\[
G_{\text{eff}} = \frac{C (1 - \phi)}{20 \pi R} (S_n + 1.5S_\tau)
\]

and the effective P- and S- wave velocities are (Winkler, 1983):

\[
V_p^2 = \frac{3C}{20 \pi R \rho} \left( S_n + \frac{2}{3} S_\tau \right)
\] (3.24)

\[
V_s^2 = \frac{C}{20 \pi R \rho} \left( S_n + \frac{2}{3} S_\tau \right)
\]

\[
\frac{V_p^2}{V_s^2} = 2 \left( \frac{3 S_n + 2}{2 S_n + 3} \right)
\]

where \( \rho \) is the grain material density.

An approximate of coordination factor \( C \) could be determined from Murphy (1982). Figure 3.6 represents the coordination number as a function of porosity.
When a hydraulic confining pressure ($P$) is applied to a random packing of spherical grains, as shown in Figure 3.5, the confining force ($F$) is defined as:

$$ F = \frac{4\pi R^2 P}{1 - \phi} . $$  \hspace{1cm} (3.25)

The Hertz model defines the radius of the contact area ($a$) and the normal displacement ($\delta$) for two identical spherical grains experiencing normal compression as (Mavko et al., 1998):

$$ a = R \sqrt[3]{\frac{3FR}{8G} (1 - \nu)} $$  \hspace{1cm} (3.26)

and $\delta = \frac{a^2}{R}$,

where $G$ and $\nu$ are the shear modulus and the effective Poisson’s ratio of the grain materials respectively. As a result, normal stiffness ($S_n$) is defined as:

$$ S_n = \frac{4aG}{1 - \nu} . $$  \hspace{1cm} (3.27)

And the bulk modulus could be representing as:

$$ K_{dry} = \sqrt[3]{\frac{C^2 (1 - \phi^2)^3 G^2}{18\pi^2 (1 - \nu)^2} P_{effective}} . $$  \hspace{1cm} (3.28)

From Mindlin (1949), the tangential stiffness of a random packing of spherical grains which experiences a tangential applied force, after a normal confining force has been applied, could be expressed as:

$$ S_t = \frac{8aG}{2 - \nu} . $$  \hspace{1cm} (3.29)

In which the dry shear moluli could be calculated as:

$$ \mu_{dry} = \frac{5 - 4\nu}{5(2 - \nu)} \sqrt[3]{\frac{3C^2 (1 - \phi^2)^3 G^2}{2\pi^2 (1 - \nu)^2} P_{effective}} . $$  \hspace{1cm} (3.30)
Figure 3.5. Normal and tangential stiffness of two spherical grains against normal and shear forces acting upon them (after Mavko et al., 1998).
Figure 3.6. Coordination number represents the average number of contacts between grains in a randomly packed spherical grain medium (after Mavko et al., 1998).
Mindlin theory assumes no slip at contact surfaces between two particles. In fact such a slip will occur at the edges of contact regions (Mavko et al., 1998).

### 3.2.1.3 – Cemented Sand

Bulk and shear modulus of dry sand in which cement is deposited at grain contacts could be calculated from the *cemented sand model*. In this model cement is assumed to be elastic and the framework of it is a dense, random pack of identical spherical grains with porosity about 36% and the average number of grains in contact assumed to be 9. Properties of the cement may differ from those of the spheres (Mavko et al., 1998).

Dvorkin and Nur (1996) calculated the dry rock bulk and shear moduli for the cemented sand (Mavko et al., 1998) as:

$$K_{\text{eff}} = \frac{1}{6} C (1 - \phi) M_c \tilde{S}_n$$

$$G_{\text{eff}} = \frac{3}{5} K_{\text{eff}} + \frac{3}{20} C (1 - \phi) G_c \tilde{S}_\tau$$

$$M_c = \rho_c V_{pc}^2$$

$$G_c = \rho_c V_{sc}^2$$

where $\rho_c$ is the cement’s density, $V_{pc}$ and $V_{sc}$ are its P- and S-wave velocities, and $\tilde{S}_n$ and $\tilde{S}_\tau$ are the normal and shear stiffness respectively, which calculate as the following relationships:

$$\tilde{S}_n = A_n \alpha^2 + B_n \alpha + C_n$$

$$A_n = -0.024153 \Lambda_n^{-1.3646}$$

$$B_n = 0.20405 \Lambda_n^{-0.89008}$$

$$C_n = 0.00024649 \Lambda_n^{-1.9864}$$

and
\[
\begin{align*}
S_r &= A_r \alpha^2 + B_r \alpha + C_r \\
A_r &= -10^{-2} (2.26 \nu^2 + 2.07 \nu + 2.3) \Lambda_r^{(0.079 \nu^2 + 0.1754 \nu - 1.342)} \\
B_r &= (0.0573 \nu^2 + 0.0937 \nu + 0.202) \Lambda_r^{(0.274 \nu^2 + 0.0529 \nu - 0.8765)} \\
C_r &= -10^{-2} (9.654 \nu^2 + 4.945 \nu + 3.1) \Lambda_r^{(0.01867 \nu^2 + 0.4011 \nu - 1.8186)}
\end{align*}
\]

where \(G\) and \(\nu\) are the shear modulus and the Poisson’s ratio of the grains, respectively; \(G_c\) and \(\nu_c\) are the shear modulus and Poisson’s ratio of the cement, respectively; \(a\) is the radius of the contact cement layer; and \(R\) is the grain radius. Assuming that all porosity reduction from initial porosity is attributed solely to cement deposition, there are two simplistic schemes by which we can determine the parameter \(\alpha = \frac{a}{R}\) as shown in Figure 3.5 (Mavko et al., 1998). The first scheme models cement deposition at grain contacts only:

\[
\alpha = 2 \left( \frac{\phi_c - \phi}{3C(1-\phi_c)} \right)^{1/2}
\]

and the second scheme models cement deposition evenly across the grain surfaces:

\[
\alpha = 2 \left( \frac{2(\phi_c - \phi)}{3(1-\phi_c)} \right)^{1/2}
\]

3.2.2 – Estimating Fluid Elastic Moduli

Pore fluids strongly affect the seismic properties of rocks. Properties of pore fluids vary systematically with their composition, pressure and the temperature (Batzle and Wang, 1992). The effects of fluids on the elastic modulus of a rock can be estimated using...
Gassmann’s equation. This equation assumes that the pore fluid is a Newtonian fluid having no resistance to shear deformation but does resist compression. Thus in Gassmann’s equations, the shear modulus of an ideal fluid which is zero has not been included and only an estimation of the fluid’s effective bulk modulus is required to calculate the saturated elastic moduli.

Wang and Batzle (1992) have studied the effects on salinity, pressure, temperature and gas (methane) dissolved in water on density and bulk modulus of the fluid. The density of brine could be calculated as (Wang and Batzle, 1992):

$$\rho_w = 1 + 1 \times 10^{-6}(-80T - 3.3T^2 + 0.00175T^3 + 489P - 2TP + 0.016T^2P - 1.3 \times 10^{-5}TP - 0.333P^2 - 0.002TP^2)$$

(3.36)

and

$$\rho_\beta = \rho_w + S(0.668 + 0.44S + 1.44 \times 10^{-6}[300P - 2400PS + T(80 + 3T - 3300S - 13P + 47PS)])$$

where $\rho_\beta$ and $\rho_w$ are the densities of water and brine at g/cm$^3$ and S is the weight fraction (ppm/1000000) of sodium chloride. The calculated density is shown in Figure 3.7.

A vast amount of acoustic data is available for brines, but generally only under the pressure, temperature, and salinity conditions found in the oceans (e.g., Spiesberger and Metzger, 1991). Wilson (1959) provides a relationship for the velocity $V_w$ of pure water to 100°C and about 100 MPa.

$$V_w = \sum_{i=0}^{4} \sum_{j=0}^{3} w_{ij} T^i P^j$$

where constants $w_{ij}$ are given in Table 2.1. Millero et al., (1977) and Chen et al., (1978) gave additional factors to be added to the velocity of water to calculate the effects of salinity. A simplified form of the velocity function was provided by Chen et al., (1978) with the constants modified to fit the additional data presented by Batzle and Wang (1992) as:
Figure 3.7. Brine density as a function of pressure, temperature, and salinity. The solid circles are selected data from Zarembo and Fedorov (1975). The lines are the regression fit to these data. “PPM” refers to the sodium chloride concentration in parts per million (after Batzle and Wang, 1992).
Figure 3.8. Calculated brine modulus as a function of pressure, temperature, and salinity (after Batzle and Wang, 1992).
| \( W_{00} \) | \( W_{01} \) | \( W_{02} \) | \( W_{03} \) | \( W_{10} \) | \( W_{11} \) | \( W_{12} \) | \( W_{13} \) | \( W_{20} \) | \( W_{21} \) | \( W_{22} \) | \( W_{23} \) | \( W_{30} \) | \( W_{31} \) | \( W_{32} \) | \( W_{33} \) | \( W_{40} \) | \( W_{41} \) | \( W_{42} \) | \( W_{43} \) |
| 1402.85 | 1.524 | 3.437 \( \times 10^{-3} \) | -1.197 \( \times 10^{-5} \) | 4.871 | -0.0111 | 1.739 \( \times 10^{-4} \) | -1.628 \( \times 10^{-6} \) | -0.04783 | 2.747 \( \times 10^{-4} \) | -6.503 \( \times 10^{-7} \) | -2.135 \( \times 10^{-6} \) | 1.478 \( \times 10^{-4} \) | -6.503 \( \times 10^{-7} \) | -1.455 \( \times 10^{-8} \) | -2.197 \( \times 10^{-7} \) | 5.23 \( \times 10^{-11} \) | 1.327 \( \times 10^{-10} \) | -4.614 \( \times 10^{-13} \) | 7.987 \( \times 10^{-10} \) | 1.327 \( \times 10^{-10} \) |

Table 3.1. Coefficients for water property computation (after Batzle and Wang, 1992).
\[ V_b = V_w + S \left( 1170 - 9.6T + 0.055T^2 - 8.5 \times 10^{-5}T^3 + 2.6P - 0.0029TP - 0.0476P^2 \right) + S^{1.5} \left( 780 - 10P + 0.16P^2 \right) - 820S^2 \]  

(3.37)

The calculated bulk moduli using equations (3.38) and (3.39) is shown in Figure 3.8.

### 3.2.3 – Gassmann-Biot Theory (Fluid Substitution)

Gassmann’s equation is widely used to predict fluid content from seismic data and in the case of 3-D seismic data, a real distribution of hydrocarbons (Calvert, 2005). By far the most widely and successful method for determining the effect of fluid changes is the Gassmann (1951) fluid substitution formulation which can be described in both isotropic and anisotropic media (Brown and Korringa, 1975).

To get a useful formula for the velocity of a fluid filled porous rock, the effect of porosity and the pore fluid must be taken into account. Gassmann (1951) derived a general relationship between the dry and the saturated rock moduli. His formulation proved to be reasonably accurate and very useful in seismic exploration. While Gassmann’s formulation is quite general and valid for all pore geometries, there are several important assumptions (Mavko, 2005):

- The rock is isotropic.
- The rock is homogeneous (mono-mineral rock, pore spaces are connected).
- The frequency is low.

The relation between the dry and saturated bulk moduli of rock can be written simply as (Mavko, 2005):

\[
\frac{K_{sat}}{K_g - K_{sat}} = \frac{K_{dry}}{K_g - K_{dry}} = \frac{K_f}{\phi(K_g - K_f)}
\]

(3.38)

and
or it may be written as:

\[ K_{\text{sat}} = K_{\text{dry}} + \eta^2 M \]  \hspace{1cm} (3.40)

where \( \eta = 1 - \frac{K_{\text{dry}}}{K_g} \).

\( M \) is Gassmann’s pore space modulus:

\[ M = \frac{K_g}{\eta - \phi \left( 1 - \frac{K_g}{K_f} \right)} \]  \hspace{1cm} (3.41)

where

\( K_{\text{sat}} \) is bulk modulus of the saturated rock,
\( K_g \) is bulk modulus of the mineral grains,
\( K_{\text{dry}} \) is bulk modulus of the dry rock matrix,
\( K_f \) is bulk modulus of the saturating pore fluid,
\( \mu_{\text{sat}} \) is shear modulus of the saturated rock,
\( \mu_{\text{dry}} \) is shear modulus of the dry rock and
\( \phi \) is porosity.

It can be seen that fluid has no effect on a rock’s shear modulus, but the bulk modulus of saturated and dry rocks are different. Equation (3.40) can be reformulated in various ways to suit exploration objectives and strategies. Typical forms are, according to Dvorkin (2001):
Fluid substitution can be performed by the following recipe (Mavko, 2005):

1. Begin with measured velocities and density (i.e. from log data); $V_p$, $V_s$ and $\rho$.

2. Extract Moduli from Velocities measured with fluid 1:

\[
K_1 = \rho \left( V_p^2 - \frac{4}{3} V_s^2 \right),
\]

and

\[
\mu_1 = \rho V_s^2.
\]

3. Transform the bulk modulus using the Gassmann equation:

\[
\frac{K_2}{K_g - K_2} - \frac{K_2 f_2}{\phi(K_g - K_{f2})} = \frac{K_1}{K_g - K_1} - \frac{K_{f1}}{\phi(K_g - K_{f1})}
\]

where

$K_1$ and $K_2$ are dynamic rock moduli with fluids 1 and 2,

$K_{f1}$ and $K_{f2}$ are bulk moduli of fluids 1 and 2,

$\rho$, and $\rho_s$ are density of rock with fluids 1 and 2

$K_g$ and $\phi$ are mineral modulus and porosity respectively and

$\rho_{f1}$ and $\rho_{f2}$ are density of fluids 1 and 2.
4. $\mu_1 = \mu_2$ shear modulus stays the same.

5. Transform density

$$\rho_2 = (1 - \phi)\rho_g + \phi \rho_{f_2} = \rho_1 + \phi(\rho_{f_2} - \rho_{f_1}). \quad (3.47)$$

6. Reassemble the velocities:

$$V_p = \sqrt{\frac{K_2 + \frac{4}{3}\mu_2}{\rho_2}}, \quad (3.48)$$

and

$$V_s = \sqrt{\frac{\mu_2}{\rho_2}}. \quad (3.49)$$

It is important to remember that the Gassmann equation assumes that the fluid has time to cross flow between pores during wave propagation (pressure in equilibrium); therefore it is only valid for a low frequency range and usually called the low frequency solution. Biot (1956) assumed that the fluid could flow through the pore spaces to give a “high frequency solution”; this introduced the additional factor of fluid viscosity and matrix permeability.

Isotropy is one of the fundamental assumptions in the Gassmann equation, however with modifications it has been discussed for anisotropic conditions. Because of the anisotropic nature of rock containing fractures in different directions, ordinary Gassmann equations are not applicable for fractured porous rocks which often host hydrocarbons and only suit the sorted grain sediments in porous media.

### 3.2.4 – Estimating Effective Grain Elastic Moduli

Gassmann’s equation for fluid substitution assumes a homogeneous mineral modulus, whereas in reality usually there are different proportions of grains within a rock matrix. The effect of different minerals in a matrix need to be averaged to allow use in Gassmann’s equations and is called an *effective* grain (McKenna, 2004).
Figure 3.9. Schematic presentation of the upper and lower bounds on the elastic bulk and shear moduli (after Mavko et al., 1998).
To calculate the effective elastic moduli of a mixture of grains and pores theoretically, it is necessary to know (Mavko et al., 1998):

1- The volume fraction of the various phases.
2- The elastic moduli of the various phases.
3- The geometric details of how the phases are arrange relative to each other.

Figure 3.9 shows the schematic presentation of the upper and lower bounds on the elastic bulk and shear modulus. Having the volume fraction of constitutions, the effective modulus will fall between the bounds (upper and lower bounds). The precise value of the effective modulus depends on how accurate the geometry of particles was defined. The best bounds will give the narrowest possible range without specifying anything about the geometries of the constitutions (Mavko et al., 1999), which Hashin and Shtrikman (1963) presented. Based on Hashin and Shtrikman (1963):

\[
K^{HSE} = K_1 + \frac{f_2}{(K_2 - K_1)^{-1} + f_1\left(K_1 + \frac{4}{3}\mu_1\right)^{-1}} \tag{3.50}
\]

\[
\mu^{HSE} = \mu_1 + \frac{f_2}{(\mu_2 - \mu_1)^{-1} + \frac{2f_1(K_1 + 2\mu_1)}{5\mu_1(K_1 + \frac{4}{3}\mu_1)}} \tag{3.51}
\]

where

\[K_1, K_2 = \text{bulk moduli of individual phases}\]
\[\mu_1, \mu_2 = \text{shear moduli of individual phases}\]
\[f_1, f_2 = \text{volume fraction of individual phases}\]

Generally, the expression gives the upper bound when the stiffest material is termed 1 in the expression above and the lower bound when the softest material is termed 1. It should be mentioned that this relationship has limitations which assumes that each constitution is isotropic, linear and elastic and the rock is also isotropic and linear elastic. Berryman (1995) later extended Hashin and Shtrikman’s formula to more than two phases.
Other efforts to provide upper and lower bounds of the elastic moduli for a homogeneous mineral aggregate introduced by Voigt (1890) and developed by Reuss (1929). Kuster and Toksöz (1974) later improved the prediction of the average elastic moduli by certain assumptions made about the geometry of the mineral constitution. Wood (1955) calculated the velocity of a wave through fluid suspensions where heterogeneities were small compared with the wavelength, based on the Reuss average.

3.2.5 – Seismic Attenuation and Quality Factor

Knowledge of attenuation can be very useful in seismic data processing, as its removal increases resolution. But its greatest potential lies as a direct hydrocarbon indicator. Attenuation depends on fluid mobility (Batzle et al 2003, Kumar et al 2003) which depends on the viscosity and bulk modulus of the pore fluid, and the permeability of the rock (Hedlin and Margrave, 2001). If the pore space is completely filled with fluid, the fluid has less mobility than if there is some gas saturation. It has been shown that attenuation is highest in a partially fluid saturated rock (Kumar et al 2003; Winkler and Nur, 1982). Much of our knowledge of attenuation comes from laboratory measurements (Kumar et al 2003). Attempts to extract attenuation from seismic data recorded at the surface have met with little success, so measurements of attenuation in terms of quality factor, are often made at a well using Vertical Seismic Profiles (VSP).

3.2.5.1 – Seismic Attenuation

The attenuation of compressional and shear waves in rocks strongly depends on the physical state and saturation conditions (Toksöz et al., 1980). From Toksöz et al. (1980), attenuation generally varies much more than the seismic velocities, as a result of change in the physical state of materials. However the experimental determination of attenuation is more difficult than the measurement of velocities. The amplitude of events on a seismic record depends on many factors. Some of these factors are in our control and the effect of others can be estimated and compensated for. There are three major source of attenuation: geometric scattering effects of the rock matrix grains, energy absorption of the wave in a viscous fluid, and an-elasticity of the rock frame particles (Kuster and Tosöz, 1974).
These three source of attenuation later expanded and categorized to six known loss mechanisms for the interaction of sound with a dispersed system: (1) viscous; (2) thermal; (3) scattering; (4) intrinsic; (5) structural; and (6) electrokinetic (Dukhin and Guetz, 1996).

Beside of the introduced mechanisms for wave attenuation, divergence is usually the major factor causing “time-dependent amplitude change” (Sheriff and Geldart, 1982), which is actually distance dependent. If the medium were homogeneous, the amplitude weakening would be inversely proportional to distance. However because velocity generally increases with depth, ray path curvature makes the wave spread out more and it reduces unit energy level. While a seismic wave propagates, the energy is spread along the surface of the wave front. For spherical waves the surface increases as it passes further from the location of the source and the energy of the wave decreases. Geometrical spreading causes the intensity and the energy density of spherical waves to decrease inversely as the square of the distance from the source (Newman, 1973). This is called spherical divergence ((Sheriff and Geldart, 1982).

Absorption causes wave energy conversion to heat, while seismic waves interfere with each other by diffraction and scattering weakens the transmission energy (Sheriff and Geldart, 1982). Basic mechanisms of transforming wave energy to the heat are not clearly understood. Internal friction is in the form of sliding friction and viscous losses in the interstitial fluids, which are probably the most important mechanisms, while the latter is more important in high permeability rocks (Sheriff and Geldart, 1982). The decrease in amplitudes due to absorption appears to be exponential with distance for elastic waves in rock, which can be written as:

\[ A = A_0 e^{-\alpha x} \] (3.52)

where \( A \) and \( A_0 \) are the amplitudes at two locations with a distance \( x \) and \( \alpha \) is the absorption coefficient.
3.2.5.2 – Quality Factor

The quality factor, $Q$, is a measure of how dissipative the material is. Generally seismic amplitude decreases exponentially with time (Sheriff and Geldart, 1982). Quality factor shows the ability of a rock to attenuate seismic waves. Physically, $Q$ is defined as the ratio of a wave’s energy to the energy dissipated per cycle of oscillation. If follows that a loss-less material has a $Q$ of infinity while a completely lossy material has a $Q$ of zero. Rocks are found to have $Q$ values in the range 10-400 with the typical sedimentary rock having a value near 100 (Hedlin and Margrave, 2001). Higher frequencies are attenuating more than lower frequencies so that the spectrum of a seismic wavelet changes with time. The lower the $Q$, the larger is the dissipation (Mavko et al., 1998).

Many postulated mechanisms predict that in solids, quality factor (quality of transmission or $Q$) should depend on frequency; however, $Q$ can be independent of frequency (since absorption is directly proportional to frequency) (Toksöz et al., 1978), due to frictional seismic energy loss as passing waves cause sliding at grain boundaries and converts seismic energy to heat. Rate independency of fractional heating seems to be the cause of reported observations that $Q$ is nearly independent of frequency (Winkler et al., 1979).

Often no distinction is made between “attenuation” and “absorption”. Because the difficulties in measuring absorption and also because the quantity of interest is usually the net decrease in wave amplitude, measurements are often made of total attenuation without regard to its cause. The absorption is often described by the quality factor ($Q$-factor) that is given by:

$$Q = \frac{\Delta E}{E} = \frac{2\pi}{\Delta E} \text{ fraction of energy loss per cycle.}$$  \hspace{1cm} (3.53)

The absorption coefficient $\alpha$ can be considered as a first approximation proportional to the frequency:

$$\frac{1}{Q} = \frac{\alpha V}{\pi f} = \frac{\alpha \lambda}{\pi}$$  \hspace{1cm} (3.54)
where $V$ is velocity of the wave and $f$ is frequency. This equation is valid only for $Q >> 1$.

Kuster and Toksoz (1974) measured the attenuation, using peak to peak amplitude. To calculate the attenuation, they used the amplitude ratio technique to remove effects of geometric spreading and the directional response to the transducers; since seismic wave amplitudes are strongly affected by geometric spreading, reflections, and scattering in addition to intrinsic damping (Toksoz et al., 1981).

The amplitude of a compressional wave ($A$) propagated in pure fluid (subscript $fl$) and in a fluid captured in a rock matrix (subscript $m$) can be written as a function of frequency ($f$) as:

$$A_{fl}(f) = G_{fl}(x) \cdot T_{fl}(f) \cdot e^{i(2 \pi f k x)},$$

$$A_{m}(f) = G_{m}(x) \cdot T_{m}(f) \cdot e^{i(2 \pi f k x)},$$

where $x$ is the distance between source and receiver and $k$ is the wave number. $G$ is a spreading factor depending upon geometry, $T$ is the frequency response of the recording system, and $\gamma(f)$ is the attenuation factor. If the geometry and recording system are kept the same for measurements in the suspension and in pure matrix, we would obtain:

$$\ln \left( \frac{A_{m}}{A_{fl}} \right) = x \left( \gamma_{fl}(f) - \gamma_{m}(f) \right).$$

For a wide variety of materials, $\gamma$ is approximately a linear function of frequency (Knopoff, 1964) which will give us: $\gamma(f) = \gamma_f + \delta$ where $\gamma$ and $\delta$ are independent of frequency. $\gamma$ is referred to as the “attenuation coefficient” which is related to the “quality factor” as: $\gamma = \pi / QV$ where $V$ is velocity. Combining the equations we will have:

$$\ln \left( \frac{A_{m}}{A_{fl}} \right) = \left( \gamma_{fl} - \gamma_{m} \right) f x + \left( \delta_{fl} - \delta_{m} \right) x,$$
The $Q$ of a rock matrix to the $Q$ of pure fluid is measured from the slope of a line fitted to the logarithm of the amplitude ratios versus frequency (spectral ratio method).

### 3.2.6 – Scattering and Dispersion of Seismic Waves

In heterogeneous media, wave scattering could be a cause of wave attenuation. Scattering also is one of the mechanisms that causes velocity dispersion in media, which includes the effect of fluid interaction such as fluid induced flow within the rock frame and pore size. Since some of the transmission ultrasonic tests that I did, included the effect of the rock frame which was simulated using glass beads (providing the scattering media dependent on the wavelength of the transmitted signal), attenuation resulting from scattering will be discussed. Velocity dispersion will also be investigated since it could occur as a result of wave scattering.

#### 3.2.6.1 – Seismic Scattering

A particle oscillating under the influence of an ultrasonic wave can generate a compressional wave. A proportion of the transmitted acoustic energy redistributed from the primary wave to the induced compressional wave which does not reach the receiver but is dispersed within the matrix is often referred to as “scattering losses”. The theory of scattering losses in a concentrated dispersive system should account for multiple scattering effects and hydrodynamic interactions between particles. However, these two factors are considered insignificant for suspensions of rigid solid particles (Dukhin and Goetz, 1996).

Mavko et al. (1998) categorized the attenuation caused by scattering into three domains: Rayleigh, stochastic Mie, and diffusion scattering which is a function of the ratio of wavelength to diameter of the scatterers. If the wavelength is much larger than a scatterer’s diameter the heterogeneous media behaves like an effective homogeneous media and scattering effects could be neglected. At very small wavelengths, the heterogeneous media is more like a piecewise homogeneous medium and scattering is
small. Scattering becomes complex when heterogeneity scales are comparable with the wavelength and when path lengths are long (Mavko et al., 1998).

Scattering effects on attenuation in terms of both Rayleigh and diffusion scattering, as predicted by Blair (1990). Melha (1983) and Berryman (1992) provided theoretical estimations of scattering effects on velocity and attenuation, which mostly assume that the concentration of scatterers is small and thus only single scattering was considered.

Yamakawa (1962) presented a relationship to calculate the attenuation of P-wave energy caused by a low concentration of small spherical inclusions. This relationship was later modified by Kuster and Toksöz (1974) based on the case of elastic spheres in a linear viscous fluid which was formerly studied by Epstein (1941) and Epstein and Carhart (1953). Aki and Richards (1980) classified the scattering effect in terms of two dimensionless numbers $ka$ and $kL$, where $k = \frac{2\pi}{\lambda}$ is the wave-number, $a$ is the characteristic scale of the heterogeneity and $L$ is the ray-path length of the incident wave in a heterogeneous media. Later Hudson (1981) gave the attenuation coefficient for elastic waves in cracked media using aligned penny-shaped ellipsoidal cracks. These theories will be expanded in Chapter 6, where calculations made from the model data are presented.

3.2.6.2 – Seismic Dispersion

Many physical mechanisms have been proposed and modelled to explain velocity dispersion and attenuation in rocks, such as: scattering, viscous and inertial fluid effects, hysteresis related to surface forces, thermo-elastic effects, phase change and so forth (Mavko et al., 1998). While scattering and surface forces appear dominant in dry or nearly dry conditions, viscous fluid mechanisms are dominant in the case of poroelasticity described by Biot (1956), local flow or squirt mechanisms; when a fluid’s effect is more important (Mavko et al., 1998). In fully saturated rock and at ultrasonic frequencies, the microscopic squirt flow induced between the stiff and soft parts of the
pore space by an elastic wave is responsible for velocity-frequency dispersion and attenuation (Dvorkin et al., 2003) whereas it is believed that for frequencies below 1 KHz, the most important cause of elastic wave attenuation is the wave induced flow between mesoscopic inhomogeneities (Müller and Gurevich, 2005). This attenuation and dispersion due to mesoscopic flow can be modelled using Biot’s equations of poroelasticity with spatially varying coefficients.

Surface waves in layered media are dispersive and their velocity is a function of frequency (Mavko et al., 1998). Thus for an impulsive time function at the source, surface waves at some distances are formed as wave trains, different frequencies arriving at different times, and their arriving time, amplitudes and phases are frequency dependent (Udías, 1999). Velocity dispersion studies are based on phase velocity changes. Figure 3.10 shows the schematic representation of phase velocity ($V$), which travels at different rates to group velocity ($U$) which is the energy at which a wave propagates. If the phase velocity is the same for all frequencies in a pulse, the pulse shape will not change and the wave is non-dispersive, whereas if the velocity varies with the frequency, the different frequency components travel with different speeds and the pulse changes shape and media- the medium and wave of which is called “dispersive” (Sheriff and Geldart, 1982). Velocity dispersion due to scattering in heterogeneous media can be explained by studying the ratio of wavelength to scatterer’s size, $\frac{\lambda}{d_s}$ (Mavko et al., 1998). At very long wavelengths ($\lambda \gg d_s$) the phase velocity is non-dispersive and is close to the static effective medium result. At very short wavelengths ($\lambda \ll d_s$) velocity is again non-dispersive and can be calculated by ray theory. But, when the wavelength is comparable to the size of the scatterers, the medium is strongly dispersive and a large change in phase velocity can be expected.

Various dispersion mechanisms and their mathematical description are dependent on their calculated velocity (or modulus) in three frequency domains (Mavko et al., 1998):
Figure 3.10 (a). Presentation of group and phase velocity (b). Phase velocity could be different from group velocity. If phase velocity at each frequency is different, velocity dispersion exists (after Sheriff and Geldart, 1982).
1- The low-frequency in which the velocity \( (V_o) \) is referred to as the “relaxed” state.
2- The high-frequency in which the velocity \( (V_\infty) \) is referred to as the “un relaxed” state.
3- The characteristic frequency, \( f_c \), which separates high frequency behaviour from low frequency and relates to the range of frequencies in which velocity is changing more rapidly.

The velocity of the wave in most of the rocks generally increases with frequency (Mavko et al., 1998) which gives us \( M_\infty > M_0 \) while \( M \) is the bulk moduli of the rock. As can be seen in Figure 3.11 the total amount of dispersion between very low and very high frequency is referred to as the modulus effect where:

\[
\frac{M_\infty - M_0}{M} = \frac{\Delta M}{M} \quad \text{where} \quad M = \sqrt{M_0 M_\infty} \tag{3.58}
\]

Depending on how large this difference is, one or two mechanisms of dispersion can be eliminated when modelling it (Mavko et al., 1998). The characteristic frequency could be estimated simply for different models, but the parameters specified for grains and micro-geometry of them, such as crack aspect ratio and rock permeability, is poorly understood.

Table 3.2 shows the characteristic frequency calculated for different mechanisms which cause dispersion. As can be seen in Figure 3.12 normalized velocity versus frequency calculated by Biot, patchy saturation, and scattering models have been compared. Based on this figure, velocity is non-dispersive at the low and high-frequency limits, whereas it changes rapidly in the transitional zone of \( f \approx f_c \). All of these curves are similar to the dispersion predicted by the standard linear solid (Mavko et al., 1998):
Figure 3.11. Schematic of the standard linear solid in the frequency domain. Elastic moduli is higher in the high frequency area (after Mavko et al., 1998).
Figure 3.12. Normalised velocity at different frequency ranges calculated using Biot, patchy saturation, and the scattering model. The wave is the most dispersive when frequency is equal to characteristic frequency ($f_c$) (after Mavko et al., 1998).
### Table 3.2. High and low frequency limits and characteristic frequency of dispersion mechanisms (after Mavko et al., 1998).

<table>
<thead>
<tr>
<th>Mechanism</th>
<th>Low-frequency limit</th>
<th>High-frequency limit</th>
<th>Characteristic frequency ((f_c))</th>
</tr>
</thead>
<tbody>
<tr>
<td>Biot [6.1]</td>
<td>Gassmann’s relations [6.3]</td>
<td>Biot’s high-frequency formula [6.1]</td>
<td>(f_{\text{biot}} \approx \phi \eta / 2\pi \rho_h \kappa)</td>
</tr>
<tr>
<td>Squirt [6.7–6.10]</td>
<td>Gassman’s relations [6.3]</td>
<td>Mavko-Jizba relations [6.7]</td>
<td>(f_{\text{squin}} \approx K_0 \alpha^3 / \eta)</td>
</tr>
<tr>
<td></td>
<td>Kuster–Toksöz dry relations [4.7] (\rightarrow) Gassmann</td>
<td>Kuster–Toksöz saturated relations</td>
<td>(\text{“})</td>
</tr>
<tr>
<td></td>
<td>DEM dry relations [4.9] (\rightarrow) Gassmann</td>
<td>DEM saturated relations</td>
<td>(\text{“})</td>
</tr>
<tr>
<td></td>
<td>Self-consistent dry relations [4.8] (\rightarrow) Gassmann</td>
<td>Self-consistent saturated relations</td>
<td>(\text{“})</td>
</tr>
<tr>
<td></td>
<td>Hudson’s dry relations [4.10] (\rightarrow) Brown and Kornig [6.5]</td>
<td>Hudson’s saturated relations</td>
<td>(\text{“})</td>
</tr>
<tr>
<td>Patchy saturation [6.11]</td>
<td>Gassman’s relations [6.3]</td>
<td>Hill equation [4.5]</td>
<td>(f_{\text{patchy}} \approx \kappa / L^2 \eta (\beta_p + \beta_h))</td>
</tr>
<tr>
<td></td>
<td>Gassman’s relations [6.3]</td>
<td>White high-frequency formula [6.13]</td>
<td>(f_{\text{patchy}} \approx \kappa K_s / \pi L^2 \eta)</td>
</tr>
<tr>
<td>Viscous shear Scattering [3.8–3.12]</td>
<td>Effective medium theory [4.1–4.12]</td>
<td>Ray theory [3.9–3.11]</td>
<td>(f_{\text{visc} \cdot \text{crack}} \approx \alpha \mu / 2\pi \eta)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Time average [3.9–3.11]</td>
<td>(f_{\text{scatter}} \approx V / 2\pi a)</td>
</tr>
</tbody>
</table>

\(^{a}\)Numbers in brackets [ ] refer to the handbook of rock physics (Mavko et al., 1998)

\(^{b}\)Inputs are measured dry rock modulus; no allowance for pore geometry.

\(^{c}\)Pore space is modeled as idealized ellipsoidal cracks.

\(^{d}\)Anisotropic rock modeled as idealized penny-shaped cracks with preferred orientations.

\(^{e}\)Dry rock is homogeneous; saturation has arbitrarily shaped patches.

\(^{f}\)Dry rock is homogeneous; saturation is in spherical patches. Same limits as \(^{c}\).

\(^{g}\)Dry rock can be heterogeneous; saturation is in spherical patches.

\(^{h}\)Mechanism modeled by Walsh (1969) is related to shearing of penny-shaped cracks with viscous fluids. Interested only for extremely high viscosity or extremely high frequency.

\(^{i}\)General heterogeneous three-dimensional medium.

\(^{j}\)Normal incidence propagation through layered medium.

\[\begin{align*}
\phi &= \text{porosity} \\
K_0, \mu &= \text{bulk and shear moduli of mineral} \\
K_s &= \text{saturated rock modulus} \\
\beta_n &= \text{compressibility of the pore fluid} \\
\beta_p &= \text{compressibility of the pore space} \\
a &= \text{characteristic size (or correlation length)} \\
L &= \text{characteristic size (or correlation length)} \\
\rho_n &= \text{density of the pore fluid} \\
\eta &= \text{viscosity of the pore fluid} \\
\alpha &= \text{pore aspect ratio} \\
\kappa &= \text{rock permeability} \\
V &= \text{wave velocity at } f_c
\end{align*}\]
Figure 3.13. Based on the pressure-temperature-phase diagram, at pressure and temperatures above critical point (deep reservoirs), CO$_2$ could be in the vapour phase which is a state between gas and liquid (after figure from Carbon Dioxide website, 2006).
\[ \text{Re}(M(\omega)) = \frac{M_0M_\infty \left[ 1 + \left( \frac{f}{fC} \right)^2 \right]}{M_\infty + \left( \frac{f}{fC} \right)^2 M_0} \]  

(3.59)

### 3.3 – CO₂ Phase Study

CO₂ can exist in different phases as a function of its surrounding pressure and temperature. Based on the physical properties of CO₂ in each phase, its effect on seismic waves may also be different. For accurate interpretation of seismic attributes affected by CO₂ in its different phase forms, it is necessary to know during the injection process, which phase of CO₂ we are dealing with, which changes in rock properties we are expecting to see, and which seismic attributes are sensitive to allow monitoring of the CO₂ movement underground within the reservoir. Knowledge of the seismic response to CO₂ in all phases is necessary, since CO₂ within the subsurface reservoir may change phase due to pressure and temperature variations. For example, a fluid moving vertically may experience a reduction in pressure and temperature as it travels towards the surface.

#### 3.3.1 – CO₂ phase definition

Figure 3.13 shows the pressure-temperature-phase diagram for CO₂. As can be seen in this figure, the carbon dioxide phase diagram has three phases: gaseous, liquid and solid. At normal atmospheric pressure and temperature, the stable carbon dioxide phase is gas. This means that the final product of pressure reduction to room temperature is gaseous carbon dioxide and this final state is independent of the initial phase.

The triple point (pressure 5.1 atmosphere, temperature - 56.7°C) is defined as the temperature and pressure where three phases of pure CO₂ (gas, liquid and solid) can co-exist simultaneously in thermodynamic equilibrium. Above the critical point (pressure 72.8 atm., temperature 31.1°C) the liquid and gas phase cannot exist as separate phases. This region, known as the super-fluid or supercritical phase, has properties indistinguishable from the gas phase.
In deep reservoirs in which the temperature usually is much higher than 32°C and pressure is also above 73 atm, CO$_2$ can only be in its supercritical phase, while in shallow reservoirs it could be injected in its liquid form. If the CO$_2$ migrates to upper formations, it may change phase from gaseous (supercritical) or a liquid to free gas, as has been seen in Sleipner field in the North Sea (Arts et al., 2004; McKenna, 2004).

In shallow reservoirs (low pressure and temperature) the phase of CO$_2$ (fluid or gas) is distinct from reservoir brine as well as in deep reservoirs; this can result in a large difference in properties of CO$_2$ and brine which makes the seismic method suitable for monitoring purposes (figure from Carbon Dioxide website, 2006).

CO$_2$ can be dissolved in the formation water (brine) as a function of temperature and pressure and subsequently form a weak acid. This dissolved mixture is a very important trapping mechanism of CO$_2$ over long periods of time and is the focus of this thesis. In this thesis, I have used different seismic attributes such as phase, frequency and attenuation of signal for quantification of dissolved CO$_2$ zones in brine, with respect to dissolved CO$_2$ importance as a trapping mechanism for CO$_2$ after injection into a water filled reservoir.

### 3.3.2 – Seismic Properties of CO$_2$ in Different Phases

The seismic response to pore fluids is controlled by the bulk modulus (K) and density ($\rho$) of the fluid. McKenna (2004) used isothermal density and compressibility of brine, CO$_2$, and brine fully saturated with CO$_2$ to predict the seismic response to CO$_2$ in its different phases. Based on his work, for almost all pressures and temperatures in the shallow subsurface, there is a strong contrast in the bulk modulus and density of CO$_2$ (which mostly would be in its gas form) and brine which causes considerable change in seismic velocity (Figure 3.14). CO$_2$ storage in sedimentary basins having high geothermal gradients will also be beneficial to most seismic monitoring programs, since CO$_2$ in the vapour phase (pressure and temperature above the supercritical point) will also cause a large change in density contrast which make it a good target for seismic monitoring.
Despite the considerable difference in bulk modulus and density of CO₂ in its vapour or gas form compared with brine, the contrast in density between brine saturated with CO₂ and brine containing no dissolved CO₂ is extremely small. This small contrast has shown that quantitative analysis of CO₂ volumes in solution using seismic velocity parameters would be a difficult task (McKenna, 2004). This was the reason I looked at different attributes rather than velocity for quantification purposes.

3.3.3 - Need for examining seismic attributes
Based on the previous work of McKenna (2004), seismic velocity is not a good attribute to monitor dissolved CO₂ mixtures because the difference in density and bulk modulus of dissolved CO₂ in brine and brine without CO₂ dissolved in it, is not large enough to cause considerable changes in seismic velocity. So, it became necessary to use other seismic attributes such as amplitude and frequency content. Applying techniques such as amplitude versus offset (AVO) may also provide an identification of in-situ state of phase of CO₂ in a reservoir (McKenna, 2004).

3.3.4 – Henry’s Law and Dissolved CO₂
Solubility measurements for carbon dioxide in water date back at least as far as Henry’s Law in which he illustrated his now-established law of solubility (Henry, 1803) From that time many attempts within the disciplines of pure chemistry, geochemistry, chemical engineering, physics and biology have been made on measuring CO₂ solubility in water, and which were eventually compiled by Carroll et al. (1991). These experiments were carried out with different methods of measuring the solubility and at different levels of accuracy under different temperature and pressure conditions, which could be categorized in two groups based on pressure. Since much literature studied the solubility of CO₂ at low pressure due to different applications up to 1 MPa, and in my ultrasonic experiments I used a low pressure cell to investigate the effect of dissolved CO₂ content on different seismic attributes, I studied the solubility of CO₂ in two categories, below and above 1MPa.
Figure 3.14. Bulk moduli and density of (a) CO$_2$, (b) brine (35,000 ppm) fully saturated with CO$_2$ and (c) pure brine (35,000 ppm). The contrast between brine and free CO$_2$ will have a big impact on the seismic response while seismic response of dissolved CO$_2$ is limited by the very small contrast between brine and brine saturated with CO$_2$ (after McKenna, 2004).
Figure 3.15. Solubility of CO₂ as a function of temperature and pressure below 1MPa is demonstrated by Henry’s Law (after Carroll et al., 1991).
3.3.4.1 – At Low Pressure (less than 1 MPa)
Carroll et al. (1991) presented a simple extension of Henry’s Law to accurately model the solubility of CO$_2$ in water. Their model assumed that (1) the activity coefficients for both CO$_2$ and water are equal, and (2) the Poynting correction (effect of pressure on the reference fugacity of CO$_2$) was negligible. It then used the equilibrium relationship between CO$_2$ and water vapour to calculate the Henry’s constant and then the solubility of CO$_2$. In this model the total pressure is limited to about 1MPa, because it neglects liquid phase non-ideal forms which at pressures greater than 1MPa begin to become important and results in significant errors. Also above 1MPa the Poynting correction becomes important (Carroll et al., 1991). Figure 3.15 shows the effect of temperature and pressure on solubility of CO$_2$ in eight isotherms from 0 to 80 °C.

During the ultrasonic experiments that I did the temperature was room temperature, so I used the values presented for the 20 °C isotherm. In my experiments, by knowing how much CO$_2$ was dissolved in water to produce a dissolved CO$_2$ fluid, and using this graph, the bubble point pressure for each amount of CO$_2$ dissolved in water could be calculated. I investigated the bubble point predicted by Henry’s law using ultrasonic waves and the attenuation caused by CO$_2$ bubble formation.

3.3.4.2 - At High Pressure (greater than 1 MPa)
For most reservoirs, especially deep reservoirs, temperatures up to 100 °C and pressures up to 100MPa are relevant. In these conditions the solubility of CO$_2$ will be different from the low pressure criteria. Diamond and Akinfiev (2003) established a model for predicting CO$_2$ dissolution at high pressure and temperature with a small error of 2% based on calculation and calibration of their model with existing literature on CO$_2$ solubility measurements. They used a thermodynamic model based on Henry’s Law and on recent high-accuracy equations of state assuming that the activity coefficients of aqueous CO$_2$ were equal to unity. This was found to be valid up to solubilities of approximately 2 mol% (Diamond and Akinfiev, 2003).
Figure 3.16. Phase diagram of the CO$_2$–H$_2$O system in the pressure–temperature region relevant to different studies, with water stable. LCEP denotes the lower critical end-point of the system. Quadruple point Q$_1$ involves equilibrium between H$_2$O-ice, CO$_2$-clathrate-hydrate, CO$_2$-bearing water and CO$_2$-vapour (Ice–Cla–Laq–V). The quadruple point Q$_2$ marks equilibrium between CO$_2$-clathrate-hydrate, CO$_2$-bearing water, CO$_2$-liquid and CO$_2$-vapour (Cla–Laq–L CO$_2$–V). Phase boundaries for the pure H$_2$O end-member have been omitted for clarity. In the presence of excess H$_2$O the continuation of the Laq–L CO$_2$–V curve at temperatures below Q$_2$ is metastable (dashed). The P–T area covered by Diamond and Akinfiev (2003) is shown relative to previous low-pressure compilations by Carroll et al. and Crovetto (after Diamond and Akinfiev, 2003).
Figure 3.17. P–T diagram showing selected solubility isopleths calculated between 0.25 and 4 mol%. (a) Entire P–T region of model validity; (b) details of low-P, low-T region (after Diamond and Akinfiev, 2003).
After evaluating the performance of the basic model, they made an empirical correction to arrive at a precise description of the accepted data. Similar to CO₂ solubility predictions at low pressure and temperatures (Carrol et al., 1991), the basis of the calculations was through solving the equilibrium equations of vapour CO₂ and water, in order to derive Henry’s constant.

Figure 3.16 shows the work which has been done to calculate the solubility of CO₂ in terms of pressure and temperature. Figure 3.17 shows isotherm curves for CO₂ solubility at low and high pressure (Diamond and Akinfiev, 2003). As can be seen in this figure, solubility of CO₂ increases when pressure increases and when temperature reduces.

It should be noted that solubility of CO₂ in Figures 3.15 and 3.17 is shown in mol% which needs to be scaled in terms of total mass for volumetric studies.

3.4 – Seismic Imaging and sensing of geology

Using the basic concepts of elastic wave propagation and rock physics relationships, conventional seismic interpretation’s goal is to invert recorded seismic signals into quantitative rock properties. Different seismic attributes (beside the seismic velocity and amplitude as main attributes) which I used to image and quantify the changes in the elastic properties of a matrix filled with different dissolved CO₂ fluids, is discussed in this section.

3.4.1 – Complex Trace Attributes

Complex trace analysis (Taner and Sheriff, 1977) treats seismic amplitudes as analytic signals and extracts various attributes to aid feature identification and interpretation. Computations of these attributes are carried out at each sample of the seismic trace and so have also been called instantaneous attributes.

As it can be seen in Figure 3.18, each complex seismic trace can be expressed as:

\[ F(t) = f(t) + if^{*}(t) = A^{\theta} \]
where $f(t)$ is the real part and $f^*(t)$ is the imaginary part of the complex trace. $A$ is the instantaneous amplitude and $\theta$ is the instantaneous phase. The real part is the recorded signal and may be thought of as a measure of kinetic energy (Taner and Sheriff, 1977); and the imaginary part is known as the quadrature trace and is calculated by applying a 90° phase shift to the real component using the Hilbert transform (Bracewell, 1965). This may be thought of as a measure of potential energy.

After obtaining the complex trace, it is possible to compute instantaneous attributes of the seismic signal. These attributes are sensitive to noise and multiples while the amplitude and frequency content of seismic signal must be preserved in each processing step (Yilmaz, 2001).

### 3.4.1.1 – Instantaneous amplitude (Reflection Strength)

Instantaneous amplitude which is also known as “reflection strength” or “amplitude envelope” is given by (Yilmaz, 2001) as:

$$A(t) = \sqrt{f^2(t) + f^2(t^*)}.$$  

Reflection strength is the amplitude of the envelope of a trace and is an indicator of reflectivity, and hence impedance contrast. It is independent of phase and may have its maximum at points other than peaks or troughs of the real trace, especially when a reflection is the composite of several reflections (Taner et al., 1979).

Relative change in instantaneous amplitude may be used in the interpretation of seismic data, in which the rate of change is of particular importance. High instantaneous amplitude values may indicate major changes in lithology, such as unconformities, while rapid local changes in instantaneous amplitude may indicate faulting or gas accumulations, and gradual changes may be caused by tuning effects from thinning beds, and may be of value in determining bed thickness (Taner et al., 1979).
Figure 3.18. Isometric representation of a complex trace in its real and imaginary components. The actual seismic trace is the projection of this vector onto the real plane (after Taner and Sheriff, 1977).
3.4.1.2 – Instantaneous Phase

The instantaneous phase attribute is the angle between the trace and its Hilbert transform at an instant in time (Barnes, 1994). It is an amplitude-independent estimate of character of the trace and is always plotted between -180° and +180°. Instantaneous phase is given by (Yilmaz, 2001):

\[ \theta(t) = \tan^{-1}\left[ \frac{f^*(t)}{f(t)} \right] . \]

Instantaneous phase emphasizes the continuity of events since it is independent of reflection strength and often makes weak coherent events clearer.

3.4.1.3 – Instantaneous Frequency

The notions of complex signals associated with real signals and of instantaneous frequency (Flandrin, 1993; Cohen, 1995) have been used in seismic analysis by Taner et al. (1979). Given a real signal \( x(t) \), we can build the corresponding analytic signal (or complex trace):

\[ X(t) = x(t) + jH\{x(t)\} \]

where \( H\{x(t)\} \) is the quadrature signal corresponding to \( X(t) \) obtained using the Hilbert transform:

\[ H\{x(t)\} = \frac{1}{\pi} \text{pv} \int_{-\infty}^{+\infty} \frac{x(\tau)}{t-\tau} d\tau , \]

in which

\[ \text{pv} \int_{-\infty}^{+\infty} \]

is the Cauchy principle value of the integral.

The analytic signal can also be obtained by:

(1) taking the Fourier transform of \( X(t) \);

(2) zeroing the amplitude for negative frequencies and doubling the amplitude for positive frequencies; and
(3) taking the inverse Fourier transform.

The complex trace can be expressed as:

$$X(t) = A(t)e^{i\theta(t)},$$

where $A(t)$ is the instantaneous amplitude (or reflection strength) and $\theta(t)$ is the instantaneous phase. We can then define the instantaneous frequency $\omega(t)$ as the time derivate of the instantaneous phase,

$$\omega(t) = \frac{d\theta(t)}{dt}.$$

Instantaneous frequency is very sensitive to variations in the thickness of thin beds. The superposition of individual reflections from the top and bottom of a thin bed may produce a frequency pattern which characterises the composite reflection (Taner et al., 1979). Any changes in the thickness or lithology of the thin layer will result in changes in the character of the frequency response. It should be mentioned that the reliability of the instantaneous frequency is dependent on the signal-to-noise ratio of the data.

Instantaneous frequency is also useful in the search for hydrocarbons. A shift towards low frequencies is often observed from reflectors below hydrocarbon bearing reservoirs. Low frequency shadows often only occur immediately below the petroliferous zone, reflections from deeper reflectors appearing normal. Fractured zones in brittle rocks are also sometimes associated with low frequency shadows (Robertson and Nogami, 1984). Two explanations for the shift towards lower frequencies below gas sands have been suggested; the first implies that gas sands filter out the higher frequencies due to frequency dependent absorption or natural resonance; and the second suggests that the travel time through the gas sand increases due to the lower velocities which result in poor summing of underlying reflectors.

**3.4.2 – Seismic Reflectivity**

As seismic waves travel through the Earth and encounter layer boundaries, the energy of the incident wave is partitioned at each boundary. Part of the energy is reflected back into
the first medium and part is transmitted or reflected into the second medium. When a
down-going compressional wave (P-wave) arrives at a boundary, part of the incident
energy is mode-converted to shear wave energy. Both the P- and S-wave energy is partly
transmitted through this interface, as shown in Figure 3.19 (Hiltermann, 2001). For a P-wave at normal incidence in which there are no tangential stresses and displacements and
energy impinges normal to the interface, the reflection coefficient \( R \) is given by:

\[
R = \frac{\rho_2 V_2 - \rho_1 V_1}{\rho_2 V_2 + \rho_1 V_1} = \frac{Z_2 - Z_1}{Z_2 + Z_1} = \frac{1}{2} \frac{\Delta AI}{AI},
\]  

(3.60)

where \( Z \) is the impedance of a layer and is calculated as the product of velocity \( (V) \) and
density \( (\rho) \), whilst \( AI \) is acoustic impedance or zero offset reflectivity. If the angle of
incidence for an approaching ray is greater than \( 0^\circ \), the directions of all reflected and
transmitted waves are defined by Snell’s law:

\[
\frac{\sin \theta_1}{V_{p1}} = \frac{\sin \theta_2}{V_{p2}} = \frac{\sin \phi_1}{V_{s1}} = \frac{\sin \phi_2}{V_{s2}}.
\]  

(3.61)

The complete solution for the amplitude of transmitted and reflected P- and S-waves as a
function of angle are defined by Zoeprittz (1919) equations. Zoeprittz equations can be
expressed in matrix form as (Yilmaz, 2001):

\[
\begin{bmatrix}
R_p \\
R_s \\
T_p \\
T_s
\end{bmatrix} = 
\begin{bmatrix}
\sin \theta_1 & -\cos \phi_1 & \sin \theta_2 & \cos \phi_2 \\
\cos \theta_1 & -\sin \phi_1 & \cos \theta_2 & -\sin \phi_2 \\
\sin 2\theta_1 & \frac{\alpha_1}{\beta_1} \cos 2\phi_1 & \frac{\rho_2 \beta_2^2 \alpha_1}{\rho_1 \beta_1^2} \sin 2\theta_2 & \frac{\rho_2 \beta_2^2 \beta_1^2}{\rho_1 \alpha_1} \cos 2\phi_2 \\
-\cos 2\phi_1 & \frac{\beta_1}{\alpha_1} \sin 2\phi_1 & \frac{\rho_2 \alpha_2 \alpha_1}{\rho_1 \alpha_1} \cos 2\phi_2 & \frac{\rho_2 \beta_2 \alpha_2}{\rho_1 \alpha_1} \cos 2\phi_2
\end{bmatrix}
\begin{bmatrix}
\sin \theta_1 \\
\cos \theta_1 \\
\sin 2\theta_1 \\
\cos 2\phi_1
\end{bmatrix},
\]  

(3.62)

where

\[
R_p = \text{Reflected P-wave}
\]
\[ R_s = \text{Reflected S-wave} \]
\[ T_p = \text{Transmitted P-wave} \]
\[ T_s = \text{Transmitted S-wave} \]
\[ \theta_1 = \text{Incident P-wave angle} \]
\[ \theta_2 = \text{Transmitted P-wave angle} \]
\[ \phi_1 = \text{Reflected S-wave angle} \]
\[ \phi_2 = \text{Transmitted S-wave angle}. \]

3.5 – CO\textsubscript{2} Sequestration Steps in the Reservoir

Injection of CO\textsubscript{2} into porous sandstones saturated with brine will cause a significant change to the reservoir matrix over time. Injection of large volumes of CO\textsubscript{2} at high injection rates may cause storage reservoirs to become over-pressured if there is not sufficient hydrodynamic flow to accommodate the influx of CO\textsubscript{2}. Over-pressurizing the CO\textsubscript{2} may have the largest effect on seismic wave propagation, since it will reduce the effective pressure and lower the effective bulk moduli of the rock.

During the injection, CO\textsubscript{2} will be trapped by different mechanisms including: structural and stratigraphic trapping, residual CO\textsubscript{2} trapping, solubility trapping, and mineral trapping. As Figure 3.20 shows, these mechanisms have a different contribution over time. Stratigraphic trapping which is the major trapping mechanism at the beginning of injection, becomes less influential, whereas the effect of solubility increases its contribution to the storage process over time. Mineral trapping is important after a long period of time, when mineralization from the interaction of dissolved CO\textsubscript{2} with the chemical composition of rocks is enhanced. Residual CO\textsubscript{2} also has a major contribution to trapping during all periods of time; however its contribution is less at the start of injection becoming greater during the long time period after initial injection.
Figure 3.19. Notation for the principal ray path for the P and S-waves (after Hiltermann, 2001).
3.5.1 – Structural Trapping of CO₂

After the injection of CO₂ into a reservoir, the lower density of CO₂ compared with the reservoir brine causes buoyancy and it migrates to the top of the reservoir, becoming trapped below the impermeable seal, and thereafter becomes controlled by seal topography (McKenna, 2004). Structural trapping of CO₂ is the primary mechanism.

3.5.2 – Solubility Trapping of CO₂

Based on Figure 3.20, after CO₂ stayed trapped for some time, its dissolution in water increases. This process of solubility trapping makes the brine heavier, so it descends in the aquifer. An increase in the lateral extent of immiscible plume of CO₂ will significantly increase the volume extent of plume-aquifer interaction and enhances CO₂ dissolution (McKenna, 2004).

Some reservoir simulations predict that approximately 15 to 20% of the injected CO₂ mass will dissolve into the formation brine during the first 10 years of CO₂ injection (Johnson et al., 2001).

When CO₂ dissolves into the formation water, there will be a balance between the aqueous-phase pH within the immiscible CO₂ plume between the two opposing processes (Johnson et al., 2001). The first is the equilibrium aqueous solubility of influx CO₂ which dramatically decreases pH, as:

\[
CO_2(g) + H_2O \Leftrightarrow CO_2(aq) + H_2O \Leftrightarrow HCO_3^- + H^+ .
\]

The second is dissolution of silicate minerals due to a decrease in pH which increases the value of pH. Dissolution of minerals causes a slight increase in porosity of the aquifer, but dolomite mineral precipitation as a result of interaction of dissolved CO₂ fluid with rock components will increase pH and reduce porosity (Johnson et al., 2001).
Figure 3.20. Schematic representation of the change of dominant trapping mechanisms and increasing CO₂ storage with time (after Benson and Cook, 2005).
3.5.3 – CO₂ Mineral Trapping

The chemical interaction of dissolved CO₂ fluid with the formation rock is called mineral trapping. Mineral trapping causes CO₂ participate in the mineral structures. While CO₂ may be captured in the mineral structure it is chemically stable and cannot be free. This is the ultimate aim of CO₂ sequestration process.

The low pH of the aquifer as a result of CO₂ dissolution catalyze dissolution of minerals such as silicates in the rock then releases carbonate forming fluids into the CO₂⁻ saturated brine (Johnson et al., 2001). This will cause Mg²⁺ to be available exclusively from silicate minerals (Phlogopite).

\[ KMg_3AlSi_3O_{10}(OH)_2 + 2H^+ \rightleftharpoons KAlSi_3O_8 + 3Mg^{2+} + 4OH^- . \]

\[ Ca^{2+} \text{ becomes available from Albite:} \]
\[ CaAl_2Si_2O_8 + 2Na^+ + 2CO_{2(aq)} \rightleftharpoons NaAlCO_3(OH)_2 + 2SiO_2 + Ca^{2+} . \]

Carbonate precipitation within the formation will occur primarily in the form of dawsonite according to the formula:
\[ KAlSi_3O_8 + Na^+ + CO_{2(aq)} + H_2O \rightleftharpoons NaAlCO_3(OH)_2 + 3SiO_2 + K^+ . \]

Then it is in the form of the calcite-group carbonates, dolomite or magnesite:
\[ Mg_2 + CO_{2(aq)} + H_2O \rightleftharpoons MgCO_3 + 2H^+ . \]

And calcite:
\[ Ca^{2+} + CO_{2(aq)} + H_2O \rightleftharpoons CaCO_3 + 2H^+ . \]

Dawsonite will participate in the entire CO₂ plume, whilst dolomite and calcite precipitates as a carbonate rind along the outer shell of the CO₂ plume (Johnson et al., 2001).
Both dissolving the silicate minerals and formation of new minerals (such as Dolomite) will have an impact on the seismic response, because both mechanisms will cause a porosity reduction in the rock frame which will affect the elastic moduli of the rock strongly. So, repeating seismic surveys over a long period of time should allow monitoring of the zones of mineral trapping, with regard to changes in the rock structure (porosity reduction).

The structural trapping mechanism is the main target of seismic studies because of its importance in trapping CO$_2$ over short periods of time. This mechanism could be investigated using conventional seismic when CO$_2$ is in its gas form and causes a large contrast between physical properties of the reservoir and the seal. CO$_2$ in supercritical or in its liquid phase may also cause considerable change in rock properties and can be studied by conventional seismic methods.

CO$_2$ solubility trapping effects on seismic events have only been studied on attributes of seismic velocity and have not been exclusively investigated with other seismic attributes. This would be the main subject of this thesis in order to find a seismic attribute which may be used in monitoring and quantification of dissolved CO$_2$ in water.

3.6 - Summary of This Chapter

This chapter has reviewed some of the fundamental seismic theories that may be used in my future analysis of the data I collect using the transmission of ultrasonic energy through CO$_2$ in solution as a fluid (with and without different rock frames). Also, different trapping mechanisms for CO$_2$ injected into geological targets of CO$_2$ sequestration were discussed in this chapter. The following chapter will provide background information on the physical models used in my experiments, the fluid and rock frame preparation, generated and recorded signal parameters, and the recording system and instrumentation used in the experiments.
4.1 – Methodology of the Experiments

In order to study the effects of dissolved gas content and CO$_2$ phase change on the seismic characteristics, sets of direct transmission and reflection experiments were performed using a pressure cell filled with a controlled amount of dissolved CO$_2$ in water with and without a simulated rock matrix.

CO$_2$ phase changes as a function of pressure and temperature. For different dissolved CO$_2$ content at the same temperature, the pressure at which CO$_2$ changes from a dissolved to a free gas form can be predicted (Carroll et al., 1991). Diamond and Akinfiev (2003) predicted CO$_2$ solubility with pressure and temperature at realistic reservoir pressures and temperatures (up to 100 MPa and up to 100°C). CO$_2$ phase can change at a pressure as a function of the dissolved CO$_2$ content, so I decided to perform my experiments at low pressure with a low dissolved CO$_2$ content in order to study the CO$_2$ phase change effect on seismic wave propagation; so a low pressure cell was constructed. Direct transmitted tests were performed while I reduced the pore pressure from 40 to 0 psi and the transmitted ultrasonic signal allowed me to compute seismic attributes such as velocity, and attenuation as well as complex attributes such as amplitude envelope, instantaneous frequency and phase of recorded data.

The quantification of the changes in seismic attributes- particularly seismic transmission amplitude- as a function of the amount of dissolved CO$_2$ content under variations in pressure was also studied during constant temperature and volume conditions using a medium pressure cell. This would then provide a laboratory basis to explain empirical work produced by using field data. The medium pressure cell could hold a pressure of up to 400 psi (2.76 MPa) which allowed the dissolution of a greater proportion of CO$_2$ in water. Seismic transmission was the main focus of interest, which could potentially be used to quantify the dissolved CO$_2$ content. However, later complementary reflection tests were performed to explain some phenomena which were observed during the transmission tests.
4.2 – Pressure Cells

To simulate a reservoir’s variable pore pressure condition, a cell is needed to hold the confining pressure and provide the pore pressure when injecting fluid into it. I used either a synthetic sandstone core or packed glass beads as a rock matrix within the cell and the dissolved CO$_2$ fluids were made ready before injection into this simulated matrix, the same procedure as Domenico (1974) chose for his set of seismic transmission experiments.

It was important for me to study the dissolved CO$_2$ phase change to the free gas form as well as the single CO$_2$ phase (dissolved) effect on seismic attributes. It was necessary by optical observation to determine the pressure of CO$_2$ bubble formation so the body of the cell was constructed of acrylic which was sufficiently transparent to see bubbles forming and growing within the cell. On the other hand, using acrylic limited the confining pressure which could be applied to the cell. However, since the acoustic properties of the acrylic was known to be close to that of glass beads, it would cause less reflection from acoustic interfaces (surfaces) and more transmission energy, which was my main seismic focus for quantification purposes.

4.2.1 – Design of the Experiments

Using such pressure cells, seismic monitoring of the CO$_2$ phase change would be possible when a measured quantity of CO$_2$ dissolved in water under pressure, was injected into a pressure cell. A controlled decrease in pore pressure can cause CO$_2$ bubbles to emanate from a dissolved CO$_2$ fluid while transmission or reflection seismic tests are being continuously performed.

The first step of such experiments designed for the low pressure cell were to investigate which seismic attribute would be sensitive for monitoring the dissolved CO$_2$ phase change.

The next step in my experiments included tests using the medium pressure cell to study the effect of dissolved CO$_2$ phase change at greater pore pressure (well below typical oil
field reservoir pore pressures but within the range of shallow aquifer reservoirs). During this experimental stage it was also possible for me to dissolve more CO₂ in water compared with the previous stage.

It had been hoped that the final stage would be to perform the tests in a large pressure chamber, but this did not eventuate due to the requirement that a complex model must be constructed with the presence of a complex recording system and this was not available during the limited time available for this thesis research (Keshavarz and Evans, 2004).

4.2.2 – Low Pressure Cell

CO₂ can undergo a phase change as a function of a reservoir’s pore pressure and temperature, dependent on the amount of dissolved CO₂ content. Consequently, a small transparent cell was constructed from transparent acrylic which could hold pore pressure up to 50 psi (0.34 MPa) and would allow the observation of gas formation during pressure reduction while maintaining the temperature as constant (Figure 4.1(b)). A schematic design of the cell can be seen in Figure 4.1(a). It was initially easy to control the pressure system of the low pressure cell prior to moving to higher pressures, so this cell was chosen for the commencing experiments tests.

A low pressure fluid was introduced into the cell by injecting the fluid under pressure through an injection valve at the side of the cell. Inside the cell, the injected fluid would pass through a perforated acrylic plate which was perforated to provide an even, uniform fluid movement through the simulated rock matrix. The extraction valve on the other side of the cell allowed venting of the fluid for flushing the core or glass beads under pressure. A filter covered the extraction pipe to prevent the loss of any particles particularly in the case where small glass beads filled the cell.

Four stainless steel rods were positioned around the cell, preventing any curvature in shape of the cell under pressure and these rods also assisted in compacting the core or glass beads after the cell had been packed, in order to induce a partial confining pressure.
To provide a constant temperature when experimenting with dissolved CO$_2$ fluid, the cell was put into a water-filled bath at a temperature of 20°C for 1 hour. In addition, a spring-coil fluid injection tube which was submerged in the water increased the length of fluid injection before entering to the cell and caused the fluid to be injected to the cell at the bath’s water temperature (20°C).

### 4.2.3 – Medium Pressure Cell

A medium pressure cell made of transparent acrylic which could hold a pore pressure of up to 400 psi (2.76 MPa) was made for dissolved CO$_2$ quantification tests. Both reflection and transmission seismic experiments could be performed on the cell (Figure 4.2 (b)). A schematic picture of the cell can be seen in Figure 4.2 (a)).

Since I could put a higher pore pressure on this cell, this allowed me to dissolve a greater quantity of CO$_2$ into the water than previously described using the low pressure cell. Seismic tests on this cell filled with different quantities of dissolved CO$_2$, allowed me to quantify the dissolved CO$_2$ content.

As can be seen in Figure 4.3 (a), this cell was made in two parts which were connected together using bolts. Inside the cell there is a cavity (Figure 4.3 (b)) which could be filled with core or glass beads. Here again a filter covered the extraction pipe to prevent losing any particles in the case where glass beads filled the cell. Again, a spring-coil fluid injection tube increased the length of fluid pathway before entering the cell with the injected fluid being at room temperature (20°C). I made a few modifications to the cell to increase the strength of the transmitted signal through it and decrease the edge diffraction patterns which were disturbing the reflection sections, which I will explain in later chapters of this thesis.

Transducers were fixed at the centres of the top and bottom faces of the cell for transmission tests. Single or multiple offset reflection tests could be performed on the upper surface of the cell using a computer controlled movement system, which therefore allowed the ability to perform CMP surveys.
Figure 4.1 (a) Schematic of the low pressure cell. (b) Picture of low pressure cell made of transparent acrylic (to observe CO$_2$ bubbles evolving from dissolved to a free gas form) which can hold a pore pressure of up to 50 psi.
Figure 4.2 (a) Schematic of the medium pressure cell. (b) Picture of medium pressure cell made of transparent acrylic which could hold up to 400 psi pore pressure.
4.3 – Rock Frame Simulations

It was important to have a simulated rock frame for studying the effect of CO₂ phase changes. This frame would capture the CO₂ bubbles caused by reducing pressure within the pressure cell. Then the effect on seismic waves of pre-bubble and post-bubble formation fluids could be studied. Because I wished to study the seismic transmission response using real drill-hole core, and because of the possibility that heterogeneities within the real drill-hole cores could provide anomalous results (apart from the ability to obtain a sandstone drill-hole core with a diameter of 50 cm), I decided to use a synthetic sandstone core (Figure 4.4 (a) and (b)) as well as glass beads (Figure 4.4 (c) and (d)) to simulate a rock frame.

Seismic wave property theories to calculate the dry modulus for packed spherical grains have been provided by Pickett (1963) and Mindlin (1949) which were later modified by Mavko et al., (1998). The theory for cemented granular materials (sands) has also been studied by Dvorkin and Nur (1996). The elastic modulus for a saturated sandstone or glass beads can be estimated using the Gassmann equations (1951).

4.3.1 – Weak Synthetic Sandstone Core (CIPS)

Synthetic sandstones which I used in my experiments were made by CSIRO using a Calcite In-situ Precipitation cementation technique named CIPS (McKenna, 2004). In this method the solution precipitates calcite cement around the quartz grains and at the grain contacts in a manner that mimics the natural cementation processes that occur in nature. The method was suitable to make the synthetic sandstones, because firstly the synthetic core had a high porosity (36%) and secondly the grains were cemented uniformly, so the core was mostly homogeneous (Figure 4.4 (a)).

The core produced in this process was very loose and not suitable for several drying (heating) and flushing (with fluids) processes because it became weaker and looser during the preparation process of each test. This will be discussed later in the sample preparation section in this chapter.
Figure 4.3 (a) Medium pressure cell consisted of two acrylic parts. (b) Glass beads filled the cavity to simulate a rock frame.
Figure 4.4 (a) Soft synthetic sandstone core made by the CIPS process. (b) Compacted synthetic sandstone. (c) Fine glass beads (0.1mm in diameter). (d) Large glass beads (0.7 mm).
4.3.2 - Strong Synthetic Sandstone Core

If during the CIPS process, the calcite is flushed through the sand over a longer period in time, the cementation process would be increased and the result is a dense, stronger synthetic sandstone. The porosity of such synthetic sandstone which I used in most of the experiments was 36%. The density of a strong core I have obtained was also higher than a weak core I had since more cement had been deposited (Figure 4.4 (b)). As a result, the stronger synthetic core was more resistant to heating and vacuuming (in preparation for each experiment) and had less corrosion, so I used it in more experiments than the weaker core. Consequently, the rate of fluid flushing through the stronger core was less than the weak core because it was less permeable and it made it hard for fluid to penetrate through.

4.3.3 – Packed Fine Glass Beads (0.1 mm)

I used very fine spherical glass beads (0.1mm diameter) in the pressure cells to simulate a rock frame (matrix) during the experiments (Figure 4.4 (c)). These beads were excellent as a frame, because they could fit in the cell and be packed well by shaking the cell to also achieve a high porosity (37%). Equal sizes of glass beads also provided a better fit with theories which estimate the bulk modulus and were used to calculate the velocity of the seismic energy transmitted through each model’s matrix. These beads used in the experiments with the low pressure cell.

4.3.4 - Packed Large Glass Beads (0.7 mm)

Larger sized glass beads (0.7 mm diameter) made a more consistent frame for the experiments (Figure 4.4 (d)), since in practice they produced the higher porosity (38% - a larger bead will push against the container sides and cause the porosity to be slightly greater than a smaller bead size or packing could be a bit smaller than fine glass beads). I used these larger beads since their permeability was also higher than small beads, so they could be flushed easier. Another benefit of eventually preferring the use of beads to synthetic sandstone core was that beads provide a confirmed homogeneous medium whereas the simulated cores still had in-homogeneities and high permeability streaks.
within them. These beads provided a suitable matrix for the experiments in both low pressure and medium pressure cells.

As Mindlin (1949) described for un-cemented beads, the effective pressure affects the dry bulk modulus directly and beads need to be compacted to transfer the seismic energy from bead to bead. In the experiments, beads compacted together as a result of tightening the model’s two acrylic halves of the cell together containing the cavity filled with beads.

4.3.5 – Packed Very Large Glass Beads (2 mm)

I used very large beads (2 mm) in the medium pressure cell to compare the results of seismic transmission on this matrix with 0.7mm beads as the matrix. The porosity that pores between spherical grains provide is not a function of grain size since the pore size is related to the size of grains. Larger grains provide larger pore size. I used 2 mm glass beads in the experiments to simulate the rock frame with large pore spaces. Once the transmission experiments had been completed with the smaller set of beads (0.7 mm), they were replaced with the larger beads (2 mm) and repeated. It was to be expected that the recorded transmission waveforms would vary slightly, due to the different sized beads and their attenuation due to scattering and absorption of the transmitted ultrasonic energy. The commencing centre frequency provided a nominal wavelength of 38.5 mm, and this would be passing through beads of 2 mm and 0.7 mm diameter.

4.3.6 – Sample Preparation

The sample preparation process was considered as an important procedure which has required to be constant during lots of tests to avoid any systematic error in the experiments. It included a process of drying, setting up packing in the cells, vacuuming and saturating with di-ionized water before CO₂ fluid injection.

I dried both the sandstone cores and the glass beads by heating them for 6 hours at 100°C before packing the cavity of the cell. This removed any moisture from the frame. After the sample cooled down, I filled the cell with the sample. Setting up of the cores in the cell was different from the beads, since I packed the beads by shaking the cell while
hitting the body of the cell continuously, whereas with the cores, I just fitted them within the cell as a tight fit. After closing the cell, I would vacuum it for 20 minutes in order to remove any air trapped within the cores or glass beads. Residual air could appear as air bubbles within the cell and cause attenuation of seismic energy while reducing pressure during the experiments.

The last process before doing the pressure maintenance tests and performing the seismic experiments included saturating the samples with distilled water. The cell was flushed with di-ionized water at 100 psi (0.69 MPa) for 30 minutes while it was vacuumed. This step confirmed the removal of any trapped air in the cell. The next step was to fill the cell with pure water in order to inject CO$_2$ fluids to water under the pressure to prevent CO$_2$ bubbling out of solution.

4.4 – Set up of the Cells

As can be seen in Figure 4.3 (a), the pressure cell consisted of two parts which were bolted together when the cavity of the cell was filled with sand or glass beads. After filling the cavity and connecting the base of the cell to the top part and tightening the bolts, I attached the transmitter and the receiver to the outside of the cell by placing them in a position at the centre of the cell’s top and bottom sides, and then held them in place using transducer holders. In the case where a reflection test was to be performed, there was no need to connect the transducers to the cell since they would move across the top of the cell. The cell was made ready by applying pressured CO$_2$ after connecting the injection and vent tubes, and pressure gauge (Figure 4.2 (a)).

4.4.1 – Pressure Maintenance Tests

Before doing any test it was necessary to be sure there were no fluid leaks from the cell. Any leakage from a cell could cause a change in volume and decrease in pore pressure which could also cause fluid phase change due to a pressure reduction. So, I did a pressure maintenance test by flushing the cell and injecting de-ionized water at 350 psi pressure to the cells. After closing the valves, I would then observe a pressure change over time (24 hours). If the pressure had decreased, I tightened the components of the cell
firmed together and re-ran the test again unless the cell’s pressure and other conditions were maintained.

De-ionized water was flushed through two fully open valves (injection and extraction valves) at the highest possible rate at 100 psi (initial pore pressure which could be provided by a compressed air source to drive the fluids), then I could block the vent tube and close the injection valve. I then compressed the fluid captured in tubing by closing the fluid extraction valve (which was a needle valve). Since water is not compressible and the volume was fixed, this induced a pressure increase from 100 psi to 400 psi.

It became common practice not to open the cells between tests when different concentration of dissolved CO$_2$ were used, because it could disturb and move any glass beads from their original compressed location and consequently change the geometry of the frame.

4.4.2 – Recording System

To generate and record the seismic arrivals, I used a PC. The computer generated signal had its amplitude spectrum shaped using a computer code programmed in Labview. The program provided different types of signals. The input and output sampling frequency could be chosen up to 2 and 4 MHz, respectively. The centre frequency of the amplitude spectrum, the initial phase of the signal and length of signal envelope could be defined by the user. Figure 4.5 shows the screen of the computer program during one of the experiments.

The transducers had a natural frequency range of 1 MHz and I set the sampling frequency as 2 MHz which provided a 0.5 microsecond accuracy in the signal timing sequence. I chose 500 KHz frequency as the central frequency which was the dominant frequency of the transducers. I used a Tone Burst signal as the source, with 1 cycle transmission to the source transducer and its initial phase set at 0 degree to avoid having a time delay.

4.4.3 – Recording System Electronics
The electronic system included a signal generator (computer), an amplifier, a pair of similar transducers as transmitter and receiver, and an electronic receiver box which was connected to each other as shown in Figure 4.6.

In order to generate a controlled signal and record the received signal as digital files, I used a PC equipped with an A/D converter. Routines had been written using LabView to generate a signal and receive the transmitted or reflected signal as binary files. The received signal could be seen on the screen and saved as a file.

A Krohn-Hite amplifier was used to amplify the computer generated source signal. The range of source output voltages induced at the transducer was up to 110 Volts peak; I chose 25 Volts which gave a good received signal during initial tests and also avoided any damage to the transducers.

To receive the transmitted or reflected wave, I used a Panametrics electronic receiver/pulser box which was connected to the PC via the analog to digital converter for digitally recording the received signal.

4.4.4 – Transducers

The seismic waves used in both reflection and transmission tests were Videoscan type 103 ultrasonic contact transducers made by Panametrics (Figure 4.7(a)). These transducers generate a longitudinal wave with a non-semicolonircular radiation pattern (Figure 4.8), that is intended for direct contact with a test piece and are equipped with a WC5 wear face that offers superior wear resistance and probe life (Panametrics Ref Manufacturers tech sheet, 2006). Videoscan transducers provide a heavily damped broadband performance (Figure 4.7 (b)). They are the best choice in applications where good axial or distance resolution is necessary or in tests that require improved signal-to-noise in attenuating or scattering materials.
As stated earlier, the natural frequency range of the transducers was 0-1 MHz with a centre frequency of 500 KHz. Figure 4.7 (c) shows the signal generated with these transducers in a test on a distilled water-filled cell at 55 psi pressure.

In direct transmission tests, the transducers were fixed to the cell, whereas in reflection tests, the transducers were fixed to the moveable holders which were controlled by the computer in order to perform other forms of surveys.

4.5 – Dissolved CO₂ Preparation
The amount of CO₂ which could be dissolved in water was limited by the temperature and pressure. The temperature was set at 20°C (room temperature) and pressure was limited to the pressure which the cell system could tolerate (40 psi for the low pressure cell and 350 psi for the medium pressure cell). Based on the CO₂ solubility graph (Carroll et al., 1991) for these conditions, the maximum amount of CO₂ which could be dissolved in a litre of water was 0.3 mol CO₂. Since I wanted to study the effect of different dissolved CO₂ amounts, I dissolved CO₂ in quantities of 0, 0.1, 0.2 and 0.3 moles in water, in order to inject into the pressure cell to perform the seismic experiments.

To dissolve CO₂ in water, I used a simple mechanism which was adding solid CO₂ (dry ice) to the water in a fixed volume. But before I did this it was necessary to determine the purity of the dry ice which was to be mixed with the water.

4.5.1 – Purity of CO₂
To identify how pure the dry ice (solid CO₂) was, I filled a dry container with 50cc dry ice and covered that with a form of textile filter to let CO₂ gas escape, but the water and impurities be captured on the textile. Total weight of dry ice and the filter was 130g. Putting the container at 20°C over 12 hours, resulted in the solid CO₂ changing phase to a gaseous CO₂ and weighing again (0g in the container) showed that there was no impurity in the CO₂, especially no solid particles in the dry ice.

4.5.2 – Dissolution Process
Dissolved CO₂ fluid mixes (0.1, 0.2 and 0.3 mol) were prepared by adding known amounts of dry ice (solid CO₂) to a known volume of distilled water (Table 4.1) within a fixed volume cylinder. To do this, a simple two valve system attached to a stainless steel cylinder was used. Figure 4.9 (a) shows the schematic picture of the cylinder. Figure 4.9 (b) shows the cylinder which could hold up to 2000 psi pressure.

First I poured 695g distilled water slowly into the dry cylinder. It filled the cylinder to below the bottom valve. Then dry ice (which was weighed previously) was added to the cylinder while the bottom valve was closed and the volume between the two valves was dried by flushing with compressed air. After closing the top valve, I opened the bottom valve which let the dry ice fall into the water.

As dry ice (presumably) floated in the water which was warmer (20°C) than the dry ice (-70°C), it began changing phase to a gas. The fixed volume of the cylinder caused the pressure built up (to 1200 psi) as a result of CO₂ gas accumulation.

Shaking and cooling down the cylinder over a period of time caused the CO₂ gas to dissolve in the water and subsequently, the pressure decreased to low amounts, which then indicated that all of the added CO₂ had dissolved into the water. The density of the dissolved CO₂ was then calculated, allowing for the volume between the two valves. The CO₂ density table is given as Table 4.1.

4.6 – Summary of This Chapter
This Chapter has explained the physical model that I used for my ultrasonic experiments. The Physical description and design of the pressure cells, simulated rock frames, considerations on dissolved CO₂ preparing, and electronic instrumentation used in the experiments was also explained in this chapter. The subsequent chapter will present the methodology developed for transmission and reflection experiments, and will explain the calibration tests which were performed prior to the pressure and fluid dependent tests. Further explanations of the tests to be performed on different rock frames and fluids with some preliminary results, will also be discussed in the next Chapter.
Figure 4.5. An example of the transmitted and recorded signal controls in Labview. The input and output parameters could be set by the user to generate and record the seismic signals.
Figure 4.6. Electronic parts used in the recording system connected to a computer which was used to generate and digitally record the seismic arrivals.
Figure 4.7. (a) Panametrics videoscan contact transducers. (b) Time waveform and spectrum of a generated signal. (c) Waveform and spectral analysis of an arrival in my experiments (after Panametrics Ref book, 2006).
Figure 4.8. Radiation pattern of the ultrasonic transducers used as seismic source and receiver (after Wandler, 2007).
Figure 4.9. By adding the CO$_2$ in its solid phase as dry ice to the water having a fixed volume, and then by shaking the cylinder, the mixing process occurred. The cylinder was cooled in a water-filled ice-bath, to assist the CO$_2$ to dissolve in water contained within the cylinder. (a) Schematic of CO$_2$ dissolution cylinder, and (b) Photograph of the CO$_2$ dissolving cylinder.
Table 4.1. CO₂ dissolved content (mol) versus density. Different CO₂ amounts as dry ice were mixed with water to produce different dissolved CO₂ content (0.3, 0.2 and 0.1 mol dissolved CO₂). The GWR (Gas Water Ratio) demonstrates the ratio of gaseous CO₂ volume to water volume.

<table>
<thead>
<tr>
<th>Dissolved CO₂ content (mol/Litre)</th>
<th>Water (g)</th>
<th>Dry ice (g)</th>
<th>GWR (Lit/Lit)</th>
<th>Density (g/cm³)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.3</td>
<td>695</td>
<td>9.174</td>
<td>6.72</td>
<td>1.0114</td>
</tr>
<tr>
<td>0.2</td>
<td>695</td>
<td>6.116</td>
<td>4.48</td>
<td>1.007</td>
</tr>
<tr>
<td>0.1</td>
<td>695</td>
<td>3.058</td>
<td>2.24</td>
<td>1.0026</td>
</tr>
<tr>
<td>0</td>
<td>695</td>
<td>0</td>
<td>0</td>
<td>0.9982</td>
</tr>
</tbody>
</table>

Table 4.1. CO₂ dissolved content (mol) versus density. Different CO₂ amounts as dry ice were mixed with water to produce different dissolved CO₂ content (0.3, 0.2 and 0.1 mol dissolved CO₂). The GWR (Gas Water Ratio) demonstrates the ratio of gaseous CO₂ volume to water volume.
CHAPTER 5 – LABORATORY TESTS AND ANALYSIS

5.1 – Calibrations
To perform analysis of both reflected and transmitted signals which have been affected by porous media filled with dissolved CO₂ fluids, it was necessary to cancel off the effect of wave attenuation within the acrylic parts of the cells. I also cancel off the effect of changes in reflectivity of the acrylic-porous media, caused by different fluids injected to the cell by calculating how much reduction in energy they would cause.

An accurate measurement of intrinsic attenuation is a difficult task and seriously limits the utilization of elastic properties. Both in the laboratory and in the field, seismic amplitudes are strongly affected by geometric spreading, reflections, and scattering in addition to intrinsic damping. Thus, to obtain the true attenuation, it is necessary to correct for these other effects (Gardner et al., 1964).

In the laboratory attenuation is usually measured by one of several techniques such as the resonant bar method (Birch and Bancroft, 1938; Born, 1941; Gardner et al, 1964; Spetzler and Anderson, 1968); stress strain cycling (Jackson, 1969); amplitude decay of multiple reflections (Peselnick and Zietz, 1959); or a pulse transmission decay where amplitude decay of seismic signals travelling through a sample is measured (Kuster and Toksoz, 1974; Tittmann et al, 1974; Watson and Wuenschel, 1973).

The pulse transmission method is the most suited for pressure vessels with jacketed and saturated samples, provided corrections can be made for geometric factors such as beam spreading and reflections (Gardner et al., 1964). This method is used to measure attenuation relative to a reference sample which has very low attenuation.

Moreover these corrections, to calculate the correct velocity of the wave in media, require that the inherent temporal delay in transducers need to be corrected.
5.1.1 – Attenuation measurements

The amplitudes of a pulsed seismic wave acting as a reference and transmitting through a sample can be expressed as:

\[ A_1(f) = G_1(x) e^{-\alpha_1(x) f} e^{i(2\pi ft-k_1 x)} \]  
\[ A_2(f) = G_2(x) e^{-\alpha_1(x) f} e^{i(2\pi ft-K_2 x)} \]

where \( A = \) amplitude, \( f = \) frequency, \( x = \) distance, \( k = \frac{2\pi f}{v} = \frac{2\pi}{\lambda} = \) wave number, \( v = \) velocity, \( G(x) = \) geometrical factor which includes spreading, reflections, etc., and \( \alpha(f) \) is the frequency dependent attenuation coefficient (Toksöz and Johnston, 1981). Equations (5.1) and (5.2) refer to the reference and sample respectively. From available data it is reasonable to assume that over the frequency range of the measurements, 0.01 to 1 MHz, \( \alpha \) is a linear function of frequency although the method itself tests the assumption (Knopoff, 1964; Jackson and Anderson, 1970; McDonald et al, 1958). Thus one can write

\[ \alpha(f) = \gamma f , \]  
\[ Q = \pi f / \alpha v , \]

where \( \gamma \) is constant and related to the quality factor \( Q \) by

\[ Q = \pi f / \alpha v , \]

where the same geometry is used by both sample and standard (i.e., same sample dimensions, transducers, and geometry), \( G_1 \) and \( G_2 \) are frequency independent scale factors. The ratio of the Fourier amplitude is

\[ A_1/A_2 = G_1/G_2 e^{-(\gamma_1-\gamma_2)f x} \text{ or} \]
\[ \ln \left( A_1/A_2 \right) = (\gamma_2-\gamma_1)xf + \ln \left( G_1/G_2 \right) \]

where \( x \) is sample length.

When \( G_1/G_2 \) is independent of frequency, \( (\gamma_1-\gamma_2) \) can be found from the slope of the line fitted to \( \ln(A_1/A_2) \) versus frequency. If the \( Q \) of a standard reference is known, \( \gamma_2 \) of a
sample can be determined. When the Q of the standard is very high (i.e., $Q_1 \approx \infty$), then $\gamma_1 = 0$ and $\gamma_2$ of the rock sample can be determined directly from the slope (Gardner et al., 1964).

5.1.2 – Acrylic Attenuation Tests

In order to analyse any change in amplitude in terms of attenuation caused by fluids and the rock frame, it was necessary to measure the attenuation within the acrylic parts of the cell. Based on Figure 4.2, two acrylic pieces surrounded the cavity of the cell which contained the sample saturated with fluids. I measured the attenuation of acrylic pieces and removed their effect from the energy of the transmitted wave in order to study the seismic wave attenuation of the sample, caused by glass beads and dissolved CO$_2$ fluid between the grains.

To measure the attenuation in acrylic, I performed direct transmission tests through acrylic bodies with different lengths and measured the amplitude envelope of the transmitted wave (Figure 5.1 (a)). The exponential relationship of the amplitude and length of the acrylic piece was used to compare with wave attenuation decay equation ($A = G_0 e^{-\alpha x}$) and consequently the attenuation coefficient ($\alpha$) was calculated (Figure 5.1 (b)). $G_0$ relates to geometric spreading, reflections, and scattering in addition to intrinsic damping (Gardner et al., 1964). The expression $e^{-\alpha x}$ was later removed from the recorded signal to calculate the real amplitude change due to different dissolved CO$_2$ content.

5.1.3 – Attenuation test in Water

The attenuation in distilled water should be measured as a base line value in order to calculate the attenuation of dissolved CO$_2$ fluid. To measure the attenuation in water, I filled a tank with de-ionized water and recorded transmission signals using two 1MHz transducers at different distances within the water. Figure 5.2 (a) shows the seismic waves transmitted through water at different distances, as a gather. The attenuation coefficient ($\alpha$) was measured by fitting a logarithmic relationship on the amplitude of the transmitted wave versus distance (Figure 5.2 (b)).
Figure 5.1. (a) P-wave attenuation in acrylic due to distance. (b) Attenuation coefficient in acrylic calculated from attenuation test.
Figure 5.2. (a) P-wave attenuation in water due to distance. (b) Attenuation coefficient in water calculated from attenuation test.
5.1.4 – Transducer delay corrections
To calculate the delay caused by the small distance between a piezoelectric crystal and the surface of the transducers, I performed some simple transmission tests using acrylic pieces with different thicknesses (transmission distances) distances. By fitting a linear trend between the first break time and distance of transducers, I calculated the time delay within the transducers to be 0.2 microsecond (Figure 5.3 (a)). This calculated time had to be removed from all first break measurements to calculate the correct velocity of the transmitted waves. The velocity of the p-wave propagated through the acrylic body measured around 2732 m/s (Figure 5.3 (b)).

5.2 – Low Pressure Tests and Analysis
I designed a low pressure cell to perform tests with CO$_2$ dissolved in water at low pressure to firstly observe at which pressure CO$_2$ became liberated from its dissolved phase to form gas bubbles, and secondly to investigate which seismic attribute would be affected by different CO$_2$ content, and also when it might change phase (pre- and post-bubble formation).

I was expecting to see no major velocity change due to different dissolved CO$_2$ content (McKenna, 2004) whereas the presence of CO$_2$ as a gas through changes in pressure was expected to be measurable (Han and Batzle, 2002; Domenico, 1964, 1974, 1976). Another attribute I was expecting to change due to CO$_2$ gas generation was seismic amplitude (attenuation affected by bubbles) decay (Leighton et al., 1997; Zhang et al., 2002; Tuffin et al., 2001; Domenico, 1982; Hay and Burling, 1982; Hay and Mercer, 1985). Frequency content of the transmitted signal which was observed through seismic surveys (McKenna, 2004) was another potential attribute to study. I studied the instantaneous frequency of the signal at different pressures as well as instantaneous phase (pre- and post-bubble formation) to see any frequency changes caused by CO$_2$ bubbles.

5.2.1 – Reflection Tests
The recording system allows me to record the reflected wave. In this setting, a transmitter produced a seismic pulse and the same transmitter received the reflected wave from the
Figure 5.3 (a) Time delay in transducers can be calculated by transmission tests on acrylic bodies with different thickness. (b) Velocity of p-wave can also be calculated by linear the relationship of travel time versus travel path.
other side of the cell. The results showed that the reflected wave wasn’t strong enough to be analysed in terms of reflection time or amplitude, because of its high attenuation along the long ray path; so I did not do any further analysis of the reflection data recorded in the low pressure cell.

5.2.2– Transmission Tests
Seismic transmission tests were performed through the cell, which was either filled with glass beads of different sizes or cores having different porosities. These were the physical model types used to study the effects of dissolved CO$_2$ content and CO$_2$ phase change. In each test, the geometry of the cell was the same; both transmitter and receiver were fixed to the same end points and there was no change in matrix while pressure reduced from 40 to 0 psi. Different model matrices (synthetic sandstone core or glass beads) were used in the cell with different CO$_2$ content.

5.2.2.1– Initial Tests
The cell was initially filled with a coarse synthetic sandstone core, which had high permeability (based on observations) and porosity (36%) to provide a frame into which the solution could be adjusted. Then the cell was placed within a tank of water to fix the temperature to room temperature (21°C). Different quantities of CO$_2$ (2, 1.5, 1 and 0.5 gr dry ice) were then dissolved in one litre of distilled water and flushed through the cell at 40 psi pressure. The pore pressure could then be reduced from 40 psi to 0 psi by releasing minor amounts of fluid through a valve.

In this experiment I fixed large spherical 50 KHz transducers on the external sides of the cell; one transducer transmitted the seismic wave (p-wave) and the other transducer received the wave propagating through the fluid saturated sandstone, resulting in a transmission test. These tests were performed with the cell in a horizontal position and it was noted that the free CO$_2$ gas moved upward under buoyancy away from the ray path when the pressure reduced.
The loose and porous nature of the core allowed the capture of small CO\textsubscript{2} bubbles in the sand pore spaces which affected the transmitted signal. This action could be seen through the clear cell walls. Subsequently, the main experiments were performed with the cell in a vertical position, to let the free CO\textsubscript{2} remain in the transmission wave path.

Initially at room temperature, there appeared to be no velocity change for measurements of p-wave velocity when pressures were decreased from 40 to 0 psi. Using unconsolidated sand as the matrix, gas was observed to form as expected at lower pressures (10 psi) but this CO\textsubscript{2} phase changing from dissolved to a free gas didn’t cause any considerable change in p-wave velocity (Figure 5.4), whereas a significant change in seismic amplitude occurred. There was an expectation that the arrival velocity should reduce when gas was formed, according to the work of both Domenico (1974) and (Geertsma, 1958) amongst others.

Domenico (1974) indicated that where a core was saturated with water (a pore volume fill of 100%), then the velocity would change markedly when the water-filled pore space was replaced by up to 10% gas- after which it did not vary much. The lack of change in arrival time and hence velocity, was considered to be due to the small amount of CO\textsubscript{2} gas liberated at low pressure. This small amount of CO\textsubscript{2} gas was insufficient to form a gas layer in order to affect the velocity, since the gas was in single bubble form.

Using Henry’s Law (Carroll et al., 1991), I computed that the pore volume of free gas at 0 psi after flushing with 2gr CO\textsubscript{2} would amount to little more than 0.75% (Table 5.1) and as such, this was inadequate to observe a minor velocity change in practice.

5.2.2.2 – Tests on Loose Synthetic Sand

The first full dataset recorded in the low pressure cell was a set of direct transmission tests on loose synthetic sandstone (Figure 5.5 (a)). We chose the most porous synthetic sandstone which we could to let more fluid be captured in its cavities. More fluid content could show the effect of fluid phase change on the transmission test better than when less porous core was used. The results showed a change of seismic amplitude can help us to
Figure 5.4. No velocity change observed while reducing pressure in initial test on the low pressure cell, but seismic amplitude decreased as pressure reduced below the bubble point.
model changes of CO$_2$ content (Figure 5.5). The error in amplitude measurements (reading the error amplitude envelope) was 1.35% which was adequately low for analysis.

A regular change of the seismic amplitude at each pressure suggested that amplitude could be expressed in terms of pressure change and be related to the amount of dissolved and free CO$_2$ within the cell.

As can be seen in Figure 5.5 (b), by decreasing the dissolved CO$_2$ content, the transmission amplitude decreased regularly. Based on these results, it was observed that pressure reduction before bubble forming didn’t cause large changes in amplitude, whereas after bubble formation, signals attenuated and amplitude decreased dramatically.

5.2.2.3 – Tests on Compact Synthetic Sandstone

The results of the tests on the loose sandstone indicated that improvements in measurement accuracy were needed in the experiments, such as more intensive sampling during high amplitude slope changes. Also the loose synthetic core which was used often did not fit completely into the cell due to the corrosive nature of core preparation prior to testing (each time the core was prepared for new tests it was removed from the cell, vacuumed and heat dried resulting in weakening of the cement bond thereby making the external core sample smaller before the next set of tests). It was recognized that the synthetic core needed to be replaced with a more compact synthetic sandstone and this would also reduce fluid fingering along the holder walls when a core did not completely fit to the cell walls. A more compact core could provide a better fit to the cell and was more tolerant of the vacuuming and heat drying processes ready for the next series of experiments (with different dissolved CO$_2$ contents).

A new core was then used which was a better fit to the cell, but it had a smaller grain size and less porosity and permeability causing it to have higher density. It was apparent that gas bubbles didn’t stay in their smaller cavities (pore pressure didn’t equilibrate in the pores and bubbles didn’t form) which resulted in ineffective amplitude experiments.
Table 5.1. The pore volumes occupied with CO₂ gas at different pressures at 20º C.
Figure 5.5 (a) No velocity change observed while reducing pressure in loose sandstone, whereas the amplitudes decreased as pressure was reduced below the bubble point. (b) Using loose synthetic sandstone, the amplitude changed regularly as dissolved CO$_2$ content changed.
Changes of amplitude at each pressure appeared to be related to changes of CO$_2$ content, upon which a relationship was then developed.

The results of tests with this compact synthetic core were more irregular than the results of the tests on the replaced high permeable synthetic core because of a difference in porosity, particle size, cementation and density of the cores and different permeability path ways which resulted in variable curves.

5.2.2.4 – Tests on Fine Glass-Beads

It was thought that having variable porosity and permeability and a denser core was causing less fluid content to form within the cavities and phase changes didn’t happen as consistently as was required for stable observation of the cell. By contrast, changing the synthetic sandstone core to using fine glass beads with more porosity and permeability and less density, could provide a better medium for capturing the free gas caused by dissolved CO$_2$ during a phase change. This may allow the pore pressure to be distributed more uniformly. Having loose glass beads could provide better contact within the cell wall interior, which would also have the benefit of not having to vacuum or heat the core frame.

After replacing the core with fine glass beads (having an average size of 0.1 mm in diameter), no change in p-wave velocity during pressure reduction was observed and the amplitude changes were more stable, not following the sometimes erratic trend experience during the first transmission tests.

5.2.2.4.1 – 200 KHz Transducers

During these experiments, sudden irregular signals were received which had not appeared on the experiments before. These unknown signals appeared to be radio pickup signals. I checked the radio signal transmission in the vicinity of the laboratory and also checked the electronics units, but none of them were causing the erratic transmission readings. Eventually I found that one of the 50 KHz transducers had broken wiring so I replaced the 50 KHz transmitter with a new transmitter operating at 200 KHz and the transmitted
signal appeared stable again. I replaced both of transducers with two 200 KHz ultrasonic transducers, because there was a mis-match between 50KHz and 200 KHz transducers, and the amplitude changes due to pressure reduction became smoother, and sensibly related to dissolved CO₂ content changes.

5.2.2.5 – Tests on Large Glass-Beads
It was thought that CO₂ forming as a free gas had caused a decrease in amplitude when pressure reduced. A larger grain size would provide bigger cavities between the grains and provide a larger CO₂ phase change from dissolved to a free gas. This encouraged me to observe the effect of using larger glass beads instead of fine glass beads. Also the low permeability of fine glass beads suggested a replacement of fine glass beads (0.1 mm diameter) with the larger one (0.7 mm in diameter). Paper reviews proved such a range of grain size had been used in previous studies (Domenico, 1974). Modifying the CO₂ dissolving cylinder to provide a higher capacity of CO₂ fluid would allow an increased ability to flush the cell more continuously. Increasing the flushing provided greater confidence in removing all gas bubbles within the cell.

A uniform fluid flushing tool benefited fluid flushing through the high permeability large glass beads, and the process became more efficient and faster than previously. Also I used glands to hold the transducers against the cell tight enough to give a consistently good coupling.

The result of these changes produced the best measurements with this cell, having higher accuracy and repeatability than previously (Figure 5.6). As can be seen in the Figure 5.6, there was an interesting change in amplitude value related to the amount of dissolved CO₂ in water. A higher dissolved CO₂ content appeared to produce a higher transmission amplitude.

5.2.2.6 – Velocity and Amplitude Analysis
Two main seismic attributes I looked at were first break time (P- wave velocity) and the peak amplitude envelope of the transmitted signal. As can be seen in Figure 5.7 (a),
decreasing the pressure from 40 to 0 psi caused no change in first break time whereas the amplitude decreased dramatically. Repeating the pressure-dependent tests for different dissolved CO₂ content showed no change in wave velocity (at the same pressure) when different dissolved CO₂ content was injected into the cell (Figure 5.7 (b)). Different CO₂ content did not cause a change in velocity whereas it did cause a change in transmitted wave amplitude.

5.3 – Medium Pressure Tests and Analysis

It was thought that there being no change in velocity of the wave due to different dissolved CO₂ content could be because of the low amount of CO₂ which dissolved in water (2 gm CO₂ in 1 litre of water), so I decided to use a pressure vessel which could hold a greater value of pressure to allow me to dissolve more CO₂ into the water and observe its effect on the transmitted seismic attributes. Also doing the experiments at higher pressure made the conditions a little closer to reservoir pressures.

To achieve these aims, a new cell which could hold pressure up to 350 psi was made (Figure 4.2). Using this cell it was possible to study the reflection events as well as direct transmission, to allow the development of an AVO analysis for quantification purposes of CO₂ phase change (Wandler and Evans, 2007).

5.3.1 – Reflection Tests

Following the direct transmission test, a zero offset reflection test was recorded to produce a form of 2-D seismic section. Then a CMP survey was performed to provide reflection data from top and base of the cavity (filled by a glass bead sample) for AVO analysis (Wandler and Evans, 2007).

The main idea of the reflection tests was to use reflection events from top and bottom of the glass bead filled cavity to study AVO effects of different fluids. Results of the transmission tests on this medium pressure cell showed that the transmitted amplitudes decreased when reducing the pressure, corroborating the previous low pressure cell results.
Figure 5.6. High permeability in glass bead (0.7 mm diameter) tests caused more consistent results showing amplitude changing more regularly with regard to the amount of dissolved CO$_2$ content. Error bars show 95% confidence limits.
Figure 5.7 (a) No velocity change but a major amplitude change occurred (a) due to different pressures in the tests through the 0.7 mm glass beads. (b) Due to different dissolved CO$_2$ content, while amplitude changed in the tests through the 0.7 mm glass beads.
I wanted to investigate if the density changed due to a CO\textsubscript{2} phase change was responsible for this amplitude reduction or not (since it could cause change in reflection coefficient and consequently change in transmission amplitude); so I used the pressure dependence reflection tests through the cell while different dissolved CO\textsubscript{2} amounts were injected into the cell.

5.3.1.1 – Initial Tests
Previously it was thought that a change in density when CO\textsubscript{2} is changing from dissolved to a free gas was causing a decrease in transmission amplitude; and reflection amplitude should consequently increase.

Zero offset reflection tests on the medium-pressure cell containing 0.1mol dissolved CO\textsubscript{2} fluid within 0.7 mm glass beads showed a decrease in reflection amplitude when pressure reduced from 300 to 100 psi (Figure 5.8). In the initial test, the base of the cell was made from steel.

5.3.1.2 – Modifications to the cell
As can be seen in Figure 5.9 (a) (where 0.1 mol dissolved CO\textsubscript{2} was flushed through and filled the cell at 100 psi), multiple events were produced which were thought to have been caused by the steel base, while some noisy parts in the sections could be related to the rough surface of the steel and edge diffraction patterns in the centre of the section. It was then decided to repeat the tests after polishing the metal base to see any improvement in the zero offset reflection section.

5.3.1.3 – Tests with Modified cell
After the process of polishing the steel base and re-installing it in the cell, I produced a new seismic section in which the diffraction events from the sides of the cell were decreased and now more flat events in the centre of the section could be seen.

The seismic section produced with these new recordings (Figure 5.9 (b)), showed improvement in reducing the noise and lowering the multiple reflection events.
Figure 5.8. Reflection tests on 0.7 mm glass beads showed a decrease in reflection amplitude when pressure reduced from 300 to 100 psi.
Figure 5.9 (a) Side diffraction and multiple reflection events using a constant offset reflection test. (b) Less side diffractions and stronger flat reflection events occurred after polishing the steel base.
A decrease in reflection amplitude with pressure reduction from 300 to 0 psi in the cell with the polished steel base can be seen in the grey scale zero-offset reflection section (Figure 5.10). A decrease in reflection amplitude as well as transmission amplitude suggested that a change in density of fluid during pressure reduction could not explain the phenomena. It seemed that the wave was attenuated more when pressure decreased and this caused a reduction in both transmitted and reflected amplitudes.

To remove the multiples from the seismic section, it was considered that using a material with less impedance difference may give better results, so I changed the steel base to an acrylic which would decrease the p-wave velocity difference (impedance) between a cavity filled with glass beads and the base, which would also be expected to decrease diffractions and the strength of any multiples.

Initially, the idea was tested using a thin acrylic sheet as the steel base. Reduction in the side diffraction patterns occurred suggesting continuing the experiments with the modified model (replacing the thin acrylic layer with a thick acrylic layer which could tolerate pressure). The section using a new thick acrylic base showed very good results in that diffractions almost disappeared (Figure 5.11 (a)) and the decrease in reflection amplitude (as an indication of wave attenuation within the CO₂ and glass bead filled cavity of the cell) could be studied better however, multiple events still existed. It was noticed that the reflection from the base of the cell did not appear. It was believed that the coupling between the two acrylic parts was inadequate, so I used wide washers to provide good coupling between top and base parts of the cell (Figure 5.11 (b)).

The most interesting result of these tests was the effect of the gas accumulation at the top of the cell at 0 psi pressure which increased reflection amplitude from the top (Figure 5.12). The gas cap did not let ultrasonic energy propagate below the top and has resulted in a large reflection, similar to a bright spot. This suggests that a study of the top reflection event would be as useful as those from the base of the cell. Figure 5.13 shows the reflection amplitude reduction due to pressure reduction in the modified cell.
Figure 5.10. Decrease in reflection amplitude due to change in pressure using the polished steel base.
Figure 5.11. (a) Diffraction almost disappeared by replacing the polished steel base with a thick acrylic slab. (b) Using wide washers for better coupling of acrylic parts caused a reduced diffraction and more reflection events. The reflection from the base of the cell which couldn’t be seen in previous tests, can now be seen.
Figure 5.12. The reflection amplitude reduced due to pressure reduction in the modified cell.
Figure 5.13. Graph of reflection amplitude as pressure reduced in the modified cell.
The modified cell was thereafter used for transmission tests. Improvements in seismic reflection made the best recording equipment for transmission tests also.

### 5.3.1.4 – Velocity and Amplitude Analysis

The results of the reflection tests showed that there was no measurable change in velocity of transmitted waves during pressure reduction, while the amplitudes of transmitted energy continued to decrease with a decrease in pressure from 300 to 0 psi.

### 5.3.2 – Transmission Tests

Using the medium pressure cell, I repeated the transmission tests using greater dissolved CO$_2$ amounts because the cell could now be pressurized more than the previous low pressure cell. The results of these tests continued the work on velocity and amplitude but at higher pressures and provided an accurate acoustic methodology of studying CO$_2$ phase change from dissolved to a free gas. Also I could quantify the changes of transmission amplitude related to changes in dissolved CO$_2$ content.

### 5.3.2.1 – Tests without a matrix

I performed a full set (different CO$_2$ content) of transmission experiments on the cell containing only fluid (no glass beads or rock frame), the intent being to measure the velocity change and attenuation caused by fluids with different densities due to different dissolved CO$_2$ content as a base line for the next tests which would be using glass beads as a simulated rock frame.

After flushing the water-filled cell (to maintain the pressure and stop bubbling up when CO$_2$ bearing fluid filled the cell) with 500 cc of different CO$_2$ fluid mixes at 100 psi pressure injection, the cell was pressurized to 180 psi. Then pressure was decreased in 20 psi steps (with 15 minutes delay after each pressure reduction to let the pressure be stabilized within the cell) while recording the transmitted signal through the cell.
5.3.2.2 - Velocity and Amplitude Analysis

Figure 5.14 (a) shows the recorded traces from transmission tests through the cell without the beads for 0.1 mol dissolved CO₂ (4.4 gr CO₂ dissolved in a litre of distilled water) at pressures from 180 to 80 psi as a gather. As can be seen in the figure, the first break event time (velocity) didn’t change during the pressure reduction while minor amplitude changes are observed. No major change in amplitude happened because no bubbles could be formed even at the lowest pressure maintained (80 psi). Table 5.2 shows the values for velocity and transmission amplitudes for different dissolved CO₂ tests.

I also compared the traces related to different CO₂ content tests (at the same pressure). In Figure 5.14 (b) it could be seen clearly that the velocity of the wave with 0.3 mol dissolved CO₂ tests was 25 m/s faster than the velocity for other dissolved CO₂ tests. This meant that the density of the 0.3 mol fluid was higher than 0.1 and 0.2 mol CO₂ fluids, sufficient to cause a change in velocity, since no other factors were changed in all sets of these tests.

I also compared the amplitudes of the transmitted wave due to different CO₂ content (at the same pressures). As can be seen in Figure 5.15, the amplitude seems to be related to the CO₂ content of fluids (the higher the dissolved CO₂ content, the higher the transmission amplitude). The error bars show 5% errors. Within these error margins, we can relate several amplitude fluctuations to errors in the signal generation and recording process (for example, not sampling at the absolute waveform peak).

5.3.2.3 – Tests with matrix

The same methodology which had been applied to the tests without beads, was used on the cell filled with 0.7 mm glass beads. I designed the tests to compare the results to the tests with fluids so that I could calculate the attenuation caused by the beads. Since on this occasion I also wanted to study the CO₂ phase change effect on the seismic wave, I reduced the pressure from 180 to 0 psi, and in analysis I compared the seismic records in the same pressure range to the tests not using the beads.
Figure 5.14 (a) Gather of transmission tests (0.3 mol dissolved CO$_2$) in the medium pressure cell without glass beads. (b) Change in amplitude and velocity of P-wave due to different CO$_2$ content observed in transmission tests in the cell without beads.
<table>
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<tr>
<th>Pressure (Psi)</th>
<th>Dissolved CO2 (mol)</th>
<th>P-wave velocity (m/s)</th>
<th>Amplitude (Volt)</th>
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Table 5.2. P-wave velocity and amplitude from tests in the cell (without glass beads) at different pressures.
Figure 5.15. Change in amplitude of transmitted wave has a relationship with dissolved CO₂ amount. Error bars represent 95% confidence limits.
5.3.2.4 - Velocity and amplitude analysis

Similar to the tests without a matrix, the velocity of the wave for each different CO$_2$ amount didn’t change due to a reduction in pressure (Figure 5.16 (a)), but the amplitude of the transmitted wave did decrease. The biggest reduction in amplitude was related to the low pressure values when bubbles were forming. When bubbles were very small and couldn’t be observed through the acrylic, they attenuated the seismic wave and caused big decreases in amplitude. This observation was used to modify the Henry’s relationship of CO$_2$ solubility versus the pore pressure.

Based on these observations (Figure 5.16 (b)), the velocity of the transmitted wave didn’t change due to different dissolved CO$_2$ content, which means that irrespective of the tests without the frame fill, the bulk density hadn’t changed enough to cause any change in velocity. As a result of these tests, I could relate the amplitude (despite velocity which did not changed during the experiments with different fluids) to the amount of dissolved CO$_2$ in water..

In the tests with a matrix within the cell, amplitude differences due to differently dissolved CO$_2$ were smaller than the tests without a matrix in the cell. This indicated that the frame had reduced the effect of the fluid on seismic amplitude (more porosity, more fluid, more amplitude change). This also can be seen on the constant velocity in the tests with a matrix, whereas the velocity changed in the tests without beads as the matrix in the cell.

5.3.2.5 – Tests with Very Large Glass Beads (2 mm)

A series of ultrasonic transmission experiments were conducted on the medium pressure cell with the aim of understanding the transmission response to variations in pore size and pressure.

I performed the same procedure used for transmission experiments over the cell containing 0.7 mm glass beads as the frame. The cell’s cavity was filled with 2 mm glass beads which allowed variations in pore volume compared to 0.7 mm beads per individual
pore and length along the seismic raypath, while allowing changes in pore pressure (so porosity was constant while permeability was higher with the larger glass beads).

In the tests with 2 mm beads I injected water and 0.3 mol CO$_2$ dissolved in a litre of water into the pore matrix at pressures up to 250 psi and reduced the pressure while transmitted signal through the media was recorded. The aim was to observe the seismic transmission response to changes in pore size, pressure, and mix of dissolved CO$_2$ in water at constant temperature.

5.4 – Summary of This Chapter

In this Chapter, the calibration tests and measurements which were needed to provide corrections on the results of the ultrasonic tests were explained. This information made the tests reliable for dissolved CO$_2$ quantification. I explained the methodology of the experiments with differently pressured cells and modifications made on them to make their design suitable for final tests. I then reviewed the initial transmission tests which lead to find the seismic attributes which were sensitive to dissolved CO$_2$ content. Preliminary results of the tests using the two pressure cells and some discussions was also included in this chapter.

The complementary analysis of recorded data using different pressures, fluids, and rock matrices will be discussed in the next chapter. Also an investigation into the accuracy of Henry’s Law, which explains CO$_2$ phase change from dissolved to a free gas at different temperature and pressure conditions, will be discussed.
Figure 5.16 (a) P-wave amplitude decreased dramatically during the transmission tests in the medium pressure cell as a result of CO$_2$ bubbles forming at low pressures. (b) No velocity change was observed due to different dissolved CO$_2$ content in transmission tests on the medium pressure cell.
CHAPTER 6 – SEISMIC RESPONSE TO DISSOLVED CO₂

In this Chapter, I analyse variations in different seismic transmission attributes using the low pressure and medium pressure cells. The low pressure cell provided an initial insight into the seismic attributes of greatest importance at low pressures surrounding the bubble point from 40 psi down at ambient room pressure, and these results guided me into the study of those and other attributes when the pressure increased to values closer to 200 psi using the medium pressure cell.

6.1 – Low Pressure Results and Analysis

I analysed the results of ultrasonic transmission experiments performed on a low pressure cell (below 40 psi). The analysis involved studying the changes in p-wave velocity and amplitude as pressure was varied in the cell which contained different amounts of dissolved CO₂ content. Reductions in transmission energy at the transmitter’s centre frequency were studied as changes occurred due to the changing quality factor when CO₂ content was changed.

Since I did the experiments on this cell to obtain an idea about which attributes were most sensitive to the presence of dissolved CO₂ in fluids, and the tests were performed only below 40 psi pressure, no further analysis than that presented here was performed using the recorded data since it was considered that the maximum information was achieved using the low pressure cell. Later tests were performed at higher pressure using a different medium pressure cell to understand how these and other attributes responded at higher pressure, and these will be discussed later.

6.1.1 – Pressure Effect on Velocity and Amplitude

Velocity is the first seismic attribute which is commonly used when studying transmitted arrivals. It was shown in Figure 5.7 that the first breaks show there was no change in arrival time when pressure was reduced from 40 to 4 psi. However, there was a change in amplitude during the decreasing pressure steps. Amplitude dropped dramatically at low pressures below 20 psi when dissolved CO₂ changed phase from being dissolved in water
to becoming a free gas - bubbles filled the cell and the transmitted signal almost disappeared. A synthetic sandstone core was now placed within the cell. Figure 6.1(a) shows the change in velocity and amplitude in initial tests (using 50 KHz transducers) through the sandstone core filled with 0.075mol dissolved CO$_2$ fluid. As can be seen in this figure, amplitude decreased from 0.65 to 0.07 Volts whereas there was no change in velocity as pressure was reduced from 40 to 4 psi.

Replacing the sandstone core having a low permeability with glass beads with higher permeability provides an alternative indication. Figure 6.1 (b) shows the change of amplitude and p-wave velocity on the cell filled with 0.7mm glass beads and 0.075 mol of CO$_2$ dissolved in water. As it can be seen in this figure, again velocity did not change during the pressure reduction while the transmitted amplitude decreased from 2 Volts at 40 psi pressure to almost zero volts at 8 psi pressure and the signal was completely attenuated. The values extracted from Figure 5.10 can be seen in Table 6.1. It is assumed that the random nature in the pores of the cemented sandstone resulted in smaller bubbles being formed than in the case of the less-random matrix of glass beads.

The results of the tests showed that amplitude is more sensitive to pressure variations at low pressure (particularly around the bubble point) than p-wave velocity which is commonly used. Also, the transmission amplitude has been affected by CO$_2$ bubble right after its nucleation and even very small bubbles attenuated the ultrasonic wave and affected the transmission amplitude.

At the commencement of this research, there was an expectation that the arrival velocity should reduce when gas was formed, according to the work of both Domenico (1974) and others (Geertsma, 1958) amongst others. However, this was not the case at these low pressures; but the observation of P-wave velocity during the tests at varied pressure, showed that measuring the velocity below bubble point can’t be done properly due to substantial energy attenuation by the CO$_2$ bubbles.
Figure 6.1(a) Amplitude decreased when 0.075 mol CO$_2$ fluid changed phase to a free gas by reduction in pressure, whereas no change in velocity was observed. This test used 50 KHz transducers and was performed on the low pressure cell filled with synthetic sandstone core. (b) Amplitude decreased dramatically when 0.1 mol CO$_2$ fluid changed phase to a free gas by a reduction in pressure, whereas there was no change in observed velocity. This test used 250 KHz transducers and was performed on the low pressure cell filled with glass beads as the rock frame.
<table>
<thead>
<tr>
<th>Pressure (psi)</th>
<th>Synthetic sandstone</th>
<th>Glass beads</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Amplitude (V)</td>
<td>Velocity (m/s)</td>
</tr>
<tr>
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<td>0.65</td>
<td>2251</td>
</tr>
<tr>
<td>35</td>
<td>2251</td>
<td>1.9706</td>
</tr>
<tr>
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<td>2251</td>
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<td>2251</td>
</tr>
<tr>
<td>8</td>
<td>0.2</td>
<td>2251</td>
</tr>
</tbody>
</table>

Table 6.1. Amplitude and velocity of transmitted wave through low pressure cell filled with synthetic sandstone core and glass beads at different pressures.
Similar to the results of Han and Batzle (2002), the lack of change in arrival time and hence velocity, was considered to be due to the small amount of CO₂ and the low pressure of which CO₂ was being dissolved. Domenico (1974) indicated that where a core was saturated with water (a pore volume fill of 100%), then the velocity would change markedly when the water-filled pore space was replaced by up to 10% with gas- after which it did not vary much. Using Henry’s Law (Carroll et al., 1991), it was computed that the pore volume of free gas at 0 psi after flushing with 10 mol% of CO₂ would amount to little more than 0.75% and as such, this was inadequate to observe a minor velocity change in practice.

While the velocities for homogeneous and patchy-vertical gas saturations can be explained with existing theories, the source of the strong extensional wave attenuation resulting from partial gas saturation is presently not well understood (Liu et al., 2002). However there are uncertainties in using amplitude as a fluid detector. Amplitude analysis using time-lapse 3-D data over a reservoir has successfully shown the fluid movement (Landro, 2003). These analyses are more qualitative than quantitative. Seismic amplitude may be sufficiently sensitive to the fluid content of rocks to be used for fluid detection (Arts et al., 2004; Xue et al., 2005; Wandler, 2006), but it has not been a common parameter to date, since seismic velocities based on using Gassmann equations have been preferred in poro-elastic media studies to date (Bourbié et al., 1987).

Much has been done to study the gas/liquid effect on seismic velocity (Arts et al., 2004; Sinartio, 2002; Lei and Xue, 2006; Siggins, 2006; Wandler, 2006) and wave attenuation (Cadoret et al., 1998; Liu et al., 2002 and Yin et al., 1992), but dissolved CO₂ quantification has not been mentioned in any papers. No models currently exist to correctly quantify the amount of liquids and gas during this phase change from dissolved gas to free gas, probably because the phase-change effect can be transient in the laboratory, and hard to simulate under real field conditions. The effects however, are clearly apparent on field data (Keshavarz and Evans, 2005).
6.1.2 – Effect of varying quantity of dissolved CO₂ Content on Velocity and Amplitude

Both during the initial and the last set of the experiments performed using the low pressure cell, regardless of the amount of CO₂ in solution, by decreasing the pressure, as McKenna (2002) predicted, no change in p-wave velocity occurred- however the transmission amplitude decreased rapidly and the signal disappeared after CO₂ bubbles evolved from the dissolved CO₂ fluid.

In each set of the tests I dissolved different quantities of CO₂ into distilled water (0.025, 0.05, 0.075 and 1 mol CO₂ in 1000 gm water) and recorded the transmitted wave through the cell with these fluids at pressures from 40 to 0 psi. Initial experiments showed that the higher amount of dissolved CO₂ caused consistently higher amplitude of the transmitted arrivals (Figures 5.8 and 5.9) and the curves followed the decreasing trend (attenuation caused by CO₂ bubbles) as pressure moved lower.

Regular increases in amplitude (at all pressures) when more CO₂ was dissolved in water suggested that there is relationship between the changes of transmission amplitude to the change of quantities of CO₂ dissolved in water. As can be seen on Figure 6.2 (a), there is a good fit using the linear relationship of amplitude and CO₂ content at each pressure in the cell using 200 KHz transducers. The change of amplitude versus the changes in CO₂ content is also related (Figure 6.2 (b)). This result suggests that it may be quite possible to calculate the change of dissolved CO₂ in water during a CO₂ injection procedure using time-lapse observation of directly transmitted seismic waves (that is, VSP surveys).
Figure 6.2. (a) Amplitude related to dissolved CO$_2$ content. (b) It is possible to predict changes of dissolved CO$_2$ content by observing the change in transmission amplitude at these low pore pressures.
6.1.3 – Dissolved CO₂ Content Effect on Q Values

To study the effect of differently dissolved CO₂ content at different pressures on the quality of transmission, I used equations 5.4 and 5.5:

\[ Q = \frac{\pi}{(\gamma \cdot v)} \rightarrow Q = \frac{\pi f}{\alpha v} , \]

\[ \frac{A_1}{A_2} = \frac{G_1}{G_2} e^{-(\gamma_1-\gamma_2)fx} \text{ or} \]

\[ \ln \left( \frac{A_1}{A_2} \right) = (\gamma_2-\gamma_1)xf + \ln \left( \frac{G_1}{G_2} \right) , \]

where \( A_1 \) is amplitude of transmitted signal through a water filled cell and \( A_2 \) is amplitude of the signal passed through the cell filled with dissolved CO₂. The parameter \( Q \) is calculated just for the centre frequency (20 KHz) with a sample length= 0.195 m.

Calculated values for \( Q \) (ranging between 0 and 6) for an uncemented glass bead-filled cell, shows strong attenuation and seems reasonable with respect to the published data of Toksöz et al. (1978). The results can be seen in Figure 6.3. As can be seen in this figure, more dissolved CO₂ content causes a higher value for \( Q \). That is, dissolving CO₂ in water thereby increasing its density making it easier for waves to propagate through. Therefore an increase in density when CO₂ mixes with water in its “dissolved phase” can explain these observations.

6.2 – Medium pressure Results and Analysis

Both transmission and reflection data recorded using a medium pressure cell were processed and analysed in terms of p-wave velocity, amplitude, frequency and phase changes due to pressure and dissolved CO₂ content changes. I analysed the amplitude spectrum in order to study the wave attenuation and Q-factor, as well as the effect of CO₂ phase changing from a dissolved to a free gas using theories which explain absorption of energy of transmitted wave as a result of bubble reverberation.
Figure 6.3. Values of signal transmission quality through the water saturated sample increased by dissolving CO$_2$ into the water (which made the water a denser and stiffer liquid).
6.2.1 – Reflection Test
As mentioned previously in Chapter 5, reflection tests were used as a calibration for transmission tests and gave me the ability to improve the design of the cell in order to remove any edge diffraction patterns caused by the geometry of the cell, and to reduce any internal multiple events caused by 3-D scattering of the signal.

I also analysed the velocity and amplitude reflection data to observe the effect of both higher pressure and higher CO₂ content on CO₂ phase change from dissolved to a free gas form.

6.2.1.1 - Pressure Effect on Velocity and Amplitude on Reflection Events
As Figure 5.12 shows, in reflection tests through the medium pressure cell filled with dissolved CO₂ content, there was no velocity change during pressure reduction. The P-wave velocity in the acrylic parts of the cell was 2732.1 m/s (after performing other transmission tests through acrylic bodies) and a time delay of 0.2 microseconds was apparent in the transducer, so the velocity of the wave in the cavity of the medium pressure cell was calculated as 1775 m/s (Figure 6.3). I should mention that in Figure 5.12 at 0 psi, the reflected signal from the base of the cavity couldn’t be observed because the large amplitude reflection caused by the gas accumulation at the top of the cavity attenuated the wave completely not allowing energy to pass through the cell and also strong multiples from the top of cavity masked any possible reflection events from the base of the cavity and cell.

As Figure 5.12 shows, the amplitude of reflection from the base of the cell decreased dramatically due to the pressure reduction (whereas no velocity change was observed). The amplitude at 0 psi decreased to 0 Volts because the energy of the wave reflected from the CO₂ gas layer occupying the top of the cavity and there were no reflections from the base of the cell. Despite no reflections from the base, the reflection amplitude from the top of the cavity increased during pressure reduction and reached to its highest value at 0 psi (gas layer formed).
Figure 6.4. Reflection events from top and base of the cavity filled with glass beads and dissolved CO$_2$ fluid. The base of the cell could also be observed on the zero offset reflection section. The velocity of the wave in the cavity was 1775 m/s.
6.2.2 – Transmission Tests

I did transmission tests on the medium pressure cell both with and without glass beads using differently dissolved CO\textsubscript{2} content, and I studied both the effect of fluid on the transmitted signal’s attributes and energy scattering of the transmitted wave. Based on these two sets of tests, the effect of porosity was also studied since tests without beads (100% porosity) and the tests with glass beads (39% porosity) had been performed.

6.2.2.1 - Pressure Effect on Velocity and Amplitude

Figure 5.14 (a) showed the gather of transmitted signals through the medium pressure cell filled with dissolved CO\textsubscript{2} fluid and without glass beads at different pressure levels. As can be seen in that figure, no change in velocity has been observed and only a minor change in amplitude of the transmitted wave as pressure reduced from 200 to 80 psi. Figure 5.15 showed the amplitude change due to the pressure change as a graph. In the tests without beads I didn’t decrease the pressure below the bubble point since I was studying just the effect of dissolved fluid and also there was no frame to capture the evolved CO\textsubscript{2} bubbles in order to study the effect of free gas as well. Based on Figure 6.5 (a), which compares amplitude change due to pressure with velocity change for the 0.3 mol test, no major changes in transmission amplitude or velocity were observed.

The tests performed on the cell without glass beads and filled with distilled water showed that the measured velocity in distilled water at different pressures was very close to the velocities calculated by Batzle and Wang (1992) (only 0.8% different). However, there was no reference for velocity of different dissolved CO\textsubscript{2} fluid (Table 6.2).

Transmitted signals through the cell filled with glass beads and dissolved CO\textsubscript{2} as a fluid can be seen on Figure 5.16 (a). Figure 6.5 (b) now shows the change in amplitude and velocity due to a pressure reduction using 0.3 mol dissolved CO\textsubscript{2}. As can be seen in this figure no change in velocity was observed when pressure reduced from 210 to 70 psi whereas amplitude changed slightly from 210 to 100 psi and then decreased dramatically from 80 to 60 psi. The large amplitude reduction was a result of CO\textsubscript{2} gas evolving from dissolved CO\textsubscript{2} fluid at low pressure (80 psi) and attenuating the signal.
Figure 6.5 (a) 0.3 mol test without glass beads showed no change in either velocity or transmission amplitude. (b) Despite no change in velocity, a major reduction in transmission amplitude occurred through the cell filled with glass beads as CO₂ phase changed from dissolved to its free gas form.
Table 6.2. The velocities measured in the transmission tests through distilled water were very close to the velocities of Batzle and Wang (1992). The difference was only 0.8%.

<table>
<thead>
<tr>
<th>Pressure (psi)</th>
<th>Density (gr/cm³)</th>
<th>Batzle and Wang Velocity (m/s)</th>
<th>Measured Velocity (m/s)</th>
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</thead>
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</tr>
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</tr>
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<tr>
<td>0</td>
<td>0.9971</td>
<td>1482.35</td>
<td>1497.05</td>
</tr>
</tbody>
</table>
Figure 6.6 (a) shows the velocity comparison in the cell with 0.3 mol dissolved CO$_2$ with and without glass beads. It can be seen that adding glass beads to the cell caused the velocity to be higher (1984.4 m/s) than when I filled the cell just with fluid without glass beads (1527.1 m/s). Amplitude in the 0.3 mol tests with and without glass beads are compared in Figure 6.6 (b). At high pressure energy scattering by the glass beads has caused a lower transmission amplitude to be recorded than for the test using CO$_2$ dissolved in water alone. Changes in amplitude due to scattering, is discussed later in this chapter. At low pressures CO$_2$ bubbles have been liberated from dissolved CO$_2$, resulting in signal attenuation which disappeared at low pressure.

6.2.2.2 - Dissolved CO$_2$ Content Effect on Velocity and Amplitude

In Chapter 5, direct transmission tests through the cell filled with just fluids (CO$_2$ mixtures) demonstrated that there was a velocity increase when 0.3 mol dissolved CO$_2$ in 1000 gm of water was injected into the cell (Figure 5.14 (b)). This change in velocity seems to have been related to an increase in fluid density when more CO$_2$ was dissolved in water whereas no other parameters in the cell changed. Figure 5.16 (b) also shows that there was no velocity change due to differently dissolved CO$_2$ content in the tests performed on the cell filled with glass beads as the rock frame. This could be explained by referring to the bulk density of the matrix and fluids in the cavity of the cell.

The presence of glass beads caused less fluid to occupy the cavity and hence there was a relatively lesser density increase when dissolving more CO$_2$ in water (density of the glass beads didn’t change in the experiments). Table 6.3 (a) presents the density of different dissolved CO$_2$ content fluids (the tests without glass beads) and Table 6.3 (b) shows the bulk density due to different amount of CO$_2$ in solution.

Figure 6.7 shows the change in velocity due to differently dissolved CO$_2$ content, which was only observed using 0.3 mol CO$_2$ dissolved in 1000 gm of water. It was assumed that for the lesser quantities of CO$_2$ mix (0.2, 0.1 and 0 mol), the density change wasn’t adequate to cause a velocity change.
Figure 6.6 (a) Velocity of transmitted wave increased after adding glass beads to 0.3 mol fluid filled cell. (b) Transmission amplitude decreased due to scattering caused by glass beads (at high pressure) resulting in reducing signal as CO$_2$ bubbles formed (at lower pressure).
Table 6.3 (a) Measured density of fluids due to different quantities of dissolved CO₂ showed 1.305% increase with 0.3 mol/100 gm water. (b) Calculated bulk density changed only 0.27% due to the increase in dissolved CO₂.
Figure 6.7. Velocity of the transmitted wave increased by adding the glass beads into the cell (earlier arrivals) with increasingly higher values of dissolved CO$_2$. 
Figure 5.15 shows the amplitude change due to differently dissolved quantities at different pressures for the tests on the medium pressure cell without glass beads. As can be seen in the figure, amplitudes at different dissolved CO₂ content don’t change with pressure (within 5% error margins). In the tests without glass beads, the effect of the CO₂ bubble forming due to pressure reduction was not observed since the pressure didn’t decrease below 100 psi (bubble forming point for 0.3 mol CO₂ dissolved in 1000 gm water). The figure shows that more dissolved CO₂ in water caused higher transmission amplitudes.

In Figure 6.8 (a) it can be seen that when the pressure cell contained 0.7 mm glass beads, transmission amplitudes due to differently dissolved CO₂ also decreased with pressure reduction. Again, the error bars have a 5% error margin. Figure 6.8 (b) shows a close-up of this graph. It can be seen that at each pressure, amplitude increased when more CO₂ content was dissolved in 1000 gm of water. This regular change in amplitude made it possible to relate the amplitude change to dissolved CO₂ change which could be used as a good attribute to quantify the amount of dissolved CO₂. The pressure that CO₂ started coming out of solution can be seen in this figure as a dramatic reduction in amplitude, whereas at higher pressures (where the CO₂ existed in its dissolved form), only a slight change in amplitude could be observed. Amplitude reduction due to signal attenuation caused by CO₂ bubbles confirmed the bubble point pressures that Henry’s Law predicted (which is based on equilibrium relationship between CO₂ and water vapour which was calibrated with many optical observation of CO₂ bubbles) for each CO₂ quantity dissolved in a litre of water. This provided proof of accuracy of the preparation method used for dissolving CO₂ in pure water.

The slight decrease in amplitude might be explained to be as a result of the existence of residual gas which was trapped between the glass bead cavities (pores). These trapped gas (air or CO₂) bubbles grow larger when pressure reduces and attenuate the transmitted energy. This same phenomenon can be seen during the tests with distilled water as the test fluid.
Figure 6.8 (a) CO₂ bubbles attenuate the signal causing a major decrease in amplitude. (b) Close-up of the figure shows a regular change in amplitude due to the dissolved CO₂ content (within a 5% error margin).
Comparing transmission amplitudes with and without glass beads in the cell containing dissolved CO$_2$ showed that there was strong attenuation (scattering) due to the presence of the beads which caused a reduced amplitude to be recorded (this is discussed later in this chapter).

6.2.2.3 - Density Effect on Velocity and Amplitude

In the experiments with and without glass beads in the cell, the only parameters which were changed in the tests of differently dissolved CO$_2$ were fluid density and bulk modulus. Using the experimental data, density (refer to Table 6.2 (a) and (b)) has an approximate linear relationship with transmission amplitude (Figure 6.9 (a) and (b)). This relationship therefore has suggested the possibility of quantifying the density of fluid (amount of CO$_2$ dissolved in water), and using this knowledge, the volumetric content of injected CO$_2$ may be calculated more precisely.

There is no published reference for fluid bulk modulus changes due to the differently dissolved CO$_2$ content but since velocity changed in the tests without glass beads, I have provided the calculation of bulk modulus in Table 6.4. The calculated value for 0.3 mol dissolved CO$_2$ in 1000 gm water (2.358 GPa) was higher than the other values tested of 0, 0.1 and 0.2 mol CO$_2$ dissolved in 1000 gm water, since the recorded velocity was faster for the 0.3 mol solution.

6.2.2.4 - Dissolved CO$_2$ Content Effect on Scattering

The attenuation coefficient, \( \gamma_s = \frac{\pi f Q}{V} \) (where Q is the quality factor, V is the seismic velocity, and f is the frequency) that results from elastic scattering depends on the ratio of seismic wavelength, \( \lambda \), to the diameter, \( d_s \), of the scattering heterogeneity. Roughly speaking there are three domains for wave energy scattering (Mavko et al., 1998):

1- Rayleigh scattering, where \( \lambda > d_s \) and \( \gamma_s \propto d_s^3 f^4 \)

2- Stochastic/Mie scattering, where \( \lambda \approx d_s \) and \( \gamma_s \propto d_s f^2 \)

3- Diffusion scattering, where \( \lambda \) and \( \gamma_s \propto 1/d_s \)
Figure 6.9 (a) Density of fluid (test without glass beads as frame) has a general linear relationship with amplitude of differently dissolved CO₂ content. (b) Bulk density (glass beads filled with dissolved CO₂ fluid) also shows relationship with transmitted amplitude.
### Table 6.4

(a) Bulk modulus of dissolved CO$_2$ calculated from observed velocity of transmission test. (b) Attenuation coefficient of fluids with different dissolved CO$_2$ based on Kuster and Toksöz (1974). Higher dissolved CO$_2$ gives less attenuation coefficient which results in more transmission energy and higher amplitude.
When $\lambda \gg d$, the heterogeneous medium behaves like an effective homogeneous medium and the scattering effects may be negligible. At the other limit, when $\lambda << d$, the heterogeneous medium may be treated as a piecewise homogeneous medium. Scattering becomes complex when heterogeneity scales are comparable with the wavelength and when the path lengths are long (Mavko et al., 1998).

In my experiments, the medium consisted of randomly distributed inclusions with a specified size. The waves incident on such inhomogeneous media undergo multiple scattering (inclusions scatterers). Consequently, the wave energy is redistributed thereby reducing the amplitude of the coherent wave. The coherent or average wave was assumed to be a plane wave propagating in a homogeneous continuum characterized by a “bulk” complex wave number which is a function of frequency, physical properties and concentration of discrete scatterers, causing the effective medium to be dispersive (Varadan et al., 1989). The reasoning for assuming the wave as plane wave was that at such a high frequency (1MHz) the wave could be considered as a plane wave.

The effect of multiple scattering on the coherent wave is of great practical importance, in particular the dependence on concentration at wavelengths comparable to scatterer size. At very low concentrations (<1% by volume) multiple scattering effects can be neglected among weak scatterers and each scatterer can be treated as independent (Ishimaru, 1978). However, in many practical situations such as pores in welded segments, glass-filled polymer solids, porous rocks, etc. the volume fraction can vary from 1% to 40%, where one cannot ignore the multiple scattering effects.

In the experiments I did on the medium pressure cell with glass beads (scatterers), the centre frequency was 500 KHz and the diameter of scatterers (glass beads) was 0.7mm. Based on the measured velocity for the cell filled with water and dissolved CO$_2$ content (1984.4 m/s), I calculated $\lambda = \frac{V}{f} = 0.00397m$ and the ratio of $\frac{\lambda}{d_s}$ calculated as 5.67 which lies in the Mie scattering area.
Many theoretical estimates of the effects of scattering on velocity and attenuation have appeared (Mehta, 1983; Berryman, 1992). Most are in the long wavelength limit, and most assume that the concentration of the scatterers is small. Thus only single scattering (scattering from single scatterers without having contact) is considered.

The attenuation of P-waves caused by a low concentration of small spherical inclusions (the tests with glass beads) could be estimated by measuring seismic amplitude at each pressure, and is given by (Yamakawa, 1962; Kuster and Toksöz, 1974)

\[
\gamma_{\text{sph}} = c \frac{3\omega}{4V_p} \left( \frac{\omega}{V_p} a \right) \left[ 2B_0^2 + \frac{2}{3}(1+2\zeta^2)B_1^2 + \frac{(2+3\zeta^2)^2}{5}B_2^2 \right]
\]

(6.1)

Where

\[
B_0 = \frac{K-K'}{3K'+4\mu}
\]
\[
B_1 = \frac{\rho-\rho'}{3\rho}
\]
\[
B_2 = \frac{20}{3} \frac{\mu(\mu'-\mu)}{6\mu'(K+2\mu)+\mu(9K+8\mu)}
\]
\[
\zeta = \frac{V_p}{V_s}.
\]

The terms \(V_p\) and \(V_s\) are the P and S velocities of the host medium, respectively, \(c\) is the volume concentration of the spheres, \(a\) is their radius, \(\omega\) is the frequency, \(\rho\) is the density, \(K\) is the bulk modulus, and \(\mu\) is shear modulus. The unprimed moduli refers to the background host medium and the primed moduli refers to the inclusions.

In the case of elastic spheres in a linear viscous fluid with viscosity \(\eta\), the attenuation is given by Epstein (1941); Epstein and Carhart, (1953); Kuster and Toksöz, (1974):

\[
\gamma_{\text{sph}} = c \frac{\omega}{2V_p} (\rho-\rho') \text{Re} \left[ \frac{i+b_0-i\frac{b_0^2}{3}}{\rho(1-ib_0)-(\rho+2\rho')\frac{b_0^2}{9}} \right]
\]

(6.2)

where
This theory describes attenuation for a mono-dispersive system of spherical particles and is valid only for dilute systems. In the theory, all particles are assumed to have the same diameter and each particle can be adequately represented as a sphere. Formulas for fractional energy loss in random heterogeneous media are limited to weak scattering. The term “dilute” indicates that there is no considerations of particle-particle interaction. These normally restrict the application of the resultant theory to dispersion with a volume fraction of less than a few volume percent. However in some cases, correct interpretations of experimental data have been done even for volume fractions as large as 30% (McClements, 1994). The complicated mathematical formula cannot be generalized for particle-particle interactions and works only when particle-particle interaction is insignificant. Therefore “thermal losses” are much less sensitive to the particle - particle interaction than “viscous losses” for almost all known liquids (Dukhin and Guetz, 2002).

Based on Hay and Burling (1982), the amplitude and phase of scattered waves are affected by thermal and viscous loss mechanisms operating near the surface of each scatterer which results in both the absorption of energy from the incident wave and a reduction in total scattering energy. They showed that attenuation by a mobile elastic sphere in a viscous heat conducting fluid is formulated in terms of phase angles for spatial scattered waves which is expressed by the ratio of thermal to viscous absorption. They illustrated that this ratio is small for particles with bulk densities greater than 1.3-1.7 g/cm³ suspended in water, depending upon the thermal and elastic properties of the solid. The relative importance of viscous and thermal absorption in suspensions of solid particles in water could be estimated Hay and Burling (1982) by

\[
R_{01} = 10^{-3} \left[ \frac{2 \rho'_0 + \rho_0}{\rho'_0 - \rho_0} \right]^3.
\]

Which \( \rho_0 \) and \( \rho'_0 \) are the densities of fluid and solid particles respectively. Based on this equation, thermal losses would be negligible under these conditions for particles with

\[
b_0 = (1 + i) a \sqrt{\frac{\pi f \rho}{\eta}}.
\]
densities greater than about 1.6 g/cm³. So, since the density of the glass beads I used in the experiments were 2.55 gm/cm³, the thermal loss is negligible whereas viscous absorption of the wave in the cell containing glass beads filled with water and dissolved CO₂ fluids was a major cause of energy scattering.

In the experiments with glass beads in the cavity of the medium pressure cell, attenuation of energy of the wave was a result of scattering from the glass bead- bead contact interaction and wave-induced fluid flow. Wave-induced fluid flow means that a passing wave creates local pressure gradients within the fluid phase and the resulting fluid flow is accompanied with internal friction until the pore pressure is equilibrated again. In fact this attenuation mechanism is a dissipative process, where seismic wave energy is converted to heat (Muller and Gurevich, 2004) which has been expressed as a thermal and viscous loss of energy (Dukhin and Guetz, 2002).

However the approaches described in this section have their limitations such as a lack of calculations for particle-particle interactions in scattering, since there did not appear to be a better equation for calculating the scattering, I used equation (6.2) to calculate the attenuation coefficient for differently dissolved CO₂. Table 6.5 shows the results of the calculation. As can be seen in the table, higher dissolved CO₂ content gives a smaller attenuation coefficient which means higher energy can be transmitted through the cell and higher transmission amplitude is then received.

A more qualitative approach that explains the scattering caused by heterogeneities in a heterogeneous medium has been discussed by Aki and Richards (1980). They classified scattering phenomena in terms of two dimensionless numbers $ka$ and $kL$, where $k=2\pi/\lambda$ is the wave number, $a$ is the characteristic scale of the heterogeneity, and $L$ is the path length of the primary incident wave in the heterogeneous medium. Figure 6.10 shows the strength of scattering due to values of $ka$ and $kL$ (Mavko et al., 1998). As can be seen in this figure, scattering effects are not very important for very small or very large $ka$, and they become increasingly important with increasing $kL$. For the parameters of my experiments, $ka$ and $kL$ were calculated as 1.018 and 60.3, respectively.
Scattering problems also may be classified on the basis of the fractional energy loss caused by scattering, \( \Delta E/E \), and the wave parameter \( D \) defined by \( D = 4L/ka^2 \) which was calculated as 196.5 in my experiments. The wave parameter is the ratio of the first Fresnel zone to the scale length of the heterogeneity. Ray theory is applicable when \( D < 1 \). In this case the inhomogeneities are smooth enough to be treated as piecewise homogenous. Effective medium theories are appropriate when \( ka \) and \( \Delta E/E \) are small (Mavko et al., 1998). Based on \( ka \) and \( kL \) values calculated for my experiments, there is strong scattering in the pressure cell caused by glass bead scatterers (Figure 6.10).

### 6.2.2.5 - Dissolved CO\(_2\) Content Effect on Dispersion

Dispersion of body waves is a consequence of most theories which have been proposed to account for absorption (Sheriff and Geldart, 1982). In the experiments with dissolved CO\(_2\), to study whether dispersion happens by adding scatterers (glass beads) to different dissolved CO\(_2\) fluids, I calculated the phase velocity (velocity of the wave as a function of frequency), which is not necessarily the same as the group velocity which I calculated using signal first break times. Based on Sheriff and Geldart (1982), dispersion is more important in connection with surface and channel waves and it is expected that dispersion for body waves would be small.

To calculate the velocity as a function of frequency, I compared a reference signal- the transmitted signal through the cell containing fluids only- with the transmitted signal through the cell filled with beads and fluids. I then removed the difference in time-break of the signals by shifting the signal to align it with the reference. I transformed both signal and reference to the frequency domain so that the phase velocity could be computed from the frequency spectra of the measured signal and reference (Liu et al., 2002) by:

\[
c = \frac{\omega \Delta l}{\Delta \varphi}
\]  

(6.4)
Figure 6.10. Classification of scattering phenomena (after Mavko et al., 1998). Scattering could be interpreted in terms of wave number ($k$), path length in the sample ($L$), and characteristic scale of the scatterers ($a$). The measured parameters for the tests performed on the medium pressure cell are shown by the symbol $\oplus$ on the graph.
where \( \Delta l \) is the wave travel distance in the sample and \( \Delta \varphi \) is the phase difference between the reference and signal. Figure 6.11 (a) shows the phase difference between signal and reference for the tests with and without glass beads with water and Figure 6.11 (b) shows the calculated phase velocity for the same situation. The phase difference and phase velocity related to the tests performed with 0.1, 0.2, and 0.3 mol dissolved CO\(_2\) as the fluid is shown in Figures 6.12 to 6.14, respectively. As can be seen in Figure 6.11 (a), there is a difference in phase between signal and reference at 0-300 KHz and less difference for 900-1000 KHz, whereas in the middle of the frequency range, the phase difference is almost negligible. Comparing these to Figure 6.15 which shows the spectrum of signals and references at different dissolved CO\(_2\) content, at the frequencies that phase difference is highest, the amplitude is smallest, unaffected by random noise.

Ignoring these minor points, there was no major phase difference observed between the signal and reference. This caused no major change in phase velocity calculated for signal transmitted through glass beads to be observed (Figure 6.11 (b)).

No change in phase velocity indicated that there was no dispersion due to the presence of the glass beads.

6.2.2.6 – Wave Energy Absorption Effect of CO\(_2\) bubbles

As mentioned previously in the experiments with dissolved CO\(_2\), decreasing the pressure caused CO\(_2\) to evolve from a dissolved solution to a free gas form and CO\(_2\) bubbles start to nucleate and grow in size. The amplitude of transmission decreased as a result of the growing bubbles’ size (increasing fractional CO\(_2\) bubble volume at lower pressures). The increasing attenuation of acoustic energy could be explained by two phenomena, which are: energy scattering by the bubble surface, and energy absorption as a result of bubble reverberations.
Figure 6.11. (a) Phase difference in tests with distilled water with and without glass beads  
(b) Phase velocity of the test using the cell filled with water and glass beads as scatterers.
Figure 6.12. (a) Phase difference of tests using 0.1 mol dissolved CO$_2$ with and without glass beads. (b) Phase velocity of the test using the cell filled with 0.1 mol dissolved CO$_2$ with glass beads as scatterers.
Figure 6.13. (a) Phase difference of tests using 0.2 mol dissolved CO$_2$ with and without glass beads. (b) Phase velocity of the test using the cell filled with 0.2 mol dissolved CO$_2$ with glass beads as scatterers.
Figure 6.14. (a) Phase difference of tests using 0.3 mol dissolved CO$_2$ with and without glass beads. (b) Phase velocity of the test using the cell filled with 0.3 mol dissolved CO$_2$ with glass beads as scatterers.
Figure 6.15. Comparison using a computer modelled signal, of transmitted wave through the cell without beads, and transmitted wave through the cell with glass beads at 100 psi in differently dissolved CO₂.
Dukhin and Goetz (2002) have suggested that bubbles will affect sonic attenuation and velocity, and acoustic spectra is affected by bubbles. and also that the level of contribution to a bubbles sound velocity and its attenuation depends on the bubble size and sound frequency. Their analysis of air bubbles showed that dispersive systems operate at frequencies above 1 MHz and usually up to 100 MHz, at which point the size of the bubbles must be well below 10 micron in order to affect the frequency range. Bubbles with sizes below 10 micron are very unstable as is also known from general colloidal chemistry and the theory of flotation. According to Schaaffs (1967), “colloid-sized gas bubbles have astonishingly short lifetimes, normally between 1 µs and 1 ms.”

Absorption of acoustic wave energy by bubbles on formation and deformation and the relationship between frequency and radius of bubbles was initially introduced by Minnaert (1933) and reviewed by Leighton (1994). The reverberation (resonant) frequency of a bubble depends upon the bubble radius, the thermal properties of the gas, the dynamic shear modulus, the bulk density of the sediment and the ambient hydrostatic pressure. Minnaert (1933) represented this as:

\[
 f_r = \frac{1}{2\pi r} \left( \frac{3\mu \gamma P_0 \rho}{\rho} \right)^{\frac{1}{2}} 
\]

where \( r \) is bubble radius (ft), \( \mu \) is the polytropic factor \( (\frac{1}{\gamma} \leq \mu \leq 1) \), \( \eta \) is the coefficient of water viscosity (lb sec/ft²), \( \gamma \) is an adiabatic exponent, \( P_0 \) is the ambient pressure (lbs/ft²) and \( \rho \) is the density of gas-water mixture (lb/ft³).

Figure 6.16 shows the amplitude spectrum for 0.3 mol dissolved CO₂ tests at different pressures. At lower pressure, CO₂ came out of solution as bubbles and attenuated the wave at almost all frequencies of the spectrum. However Figure 6.17 shows that at some frequencies (close to centre frequency) energy attenuation was greater than at others. To investigate whether the reduction in amplitude spectrum was caused by bubble energy absorption, I applied the equation (6.4) for CO₂ bubbles (\( \rho = 1.0114 \text{ g/cm}^3 \), \( \gamma = 1.3 \), \( \mu = 0.85 \), and \( P_0 = 96.7 \text{ psi} \)) for frequencies of 350, 450, 550, 650, and 750 KHz.
The result predicted that bubbles with sizes of 3.748, 2.915, 2.385, 2.018, and 1.749 micrometres, could absorb the energy at those frequencies, reverberating and attenuating the energy also in the form of heat losses. However my observations during the pressure reduction process showed that the seismic wave was sensitive to both bubble size and distribution. The amplitude of the transmitted wave decreased when bubbles could be physically observed which were bigger than the micron scale in size, and bubbles of course were formed along the transmission ray path. Such small size in bubbles could be related to bubble nucleation whereas the received energy was not affected by bubbles forming at the higher pressures. As a result, it is considered that reverberation did not cause the huge attenuation resulting in major signal losses, and it seems that bubble resonance caused the major effect on acoustic propagation through the aqueous fluids. Solid scatterers would not have such an effect since such scattering (caused by inclusions) could be greater than attenuation caused by bubble resonance attenuation (specially at high frequencies).

It should also be mentioned that equation (6.4) only considers the resonance of a spherical single bubble and further work is needed to constrain the model input parameters, especially bubble size distribution, total gas volume and sediment shear strength.

6.2.2.7 - Dissolved CO$_2$ Content Effect on Attenuation

Based on Dukhin and Goetz (2002) the most well known acoustic theory for heterogeneous systems was developed by Epstein and Carhart (1953), and Allegra and Hawley (1972). Their theory takes into account the four most important mechanisms known as viscous, thermal, scattering and intrinsic attenuation. However such theory was valid only for dilute systems and describes attenuation for mono-dispersive systems of spherical particles in which there was no consideration of particle-particle interaction. “Thermal losses” are much less sensitive to the particle - particle interaction than “viscous losses” for almost all known liquids (Dukhin and Goetz, 2002).
Figure 6.16. Tests performed on 0.3 mol dissolved CO$_2$ showed that amplitudes across the spectrum decreased while CO$_2$ changed phase from dissolved to its free gas form. Below 100 psi nucleating bubbles attenuated the signal strongly, whereas resonant absorption in small bubbles (during nucleation) at high pressure (100psi) did not cause significant attenuation.
Figure 6.17. As pressure reduction occurred at 0.3 mol dissolved CO$_2$, an increase in CO$_2$ bubble size caused significant attenuation at 300 to 700 KHz frequencies.
In my experiments, with regard to equation (6.3), thermal losses would be negligible under conditions where particles had densities greater than about 1.6 g/cm³. However, viscous absorption of a wave passing through glass beads filled with water and dissolved CO₂ fluids would be a major cause of energy scattering.

Intrinsic attenuation in porous fluid-saturated structures with micro- and mesoscopic heterogeneities is the attenuation mechanism of wave caused by induced fluid flow. Wave-induced fluid flow means that a passing wave creates a local pressure gradient within the fluid phase and the resulting fluid flow is accompanied by internal friction until the pore pressure is equilibrated again. In fact, this attenuation mechanism is a dissipation process, in which seismic energy is converted to heat (Muller and Gurevich, 2004).

Table 6.3 (b) shows the attenuation coefficient calculated using equation (6.2) which is based on scattering of transmitted energy through the cell filled with glass beads containing differently dissolved CO₂ for the centre frequency of the recorded data. I also calculated the attenuation coefficient for the experiments with differently dissolved CO₂ based on the practical tests rather than theoretical. To calculate the experimental attenuation coefficient, I compensated for the attenuation within the acrylic parts of the cell and also considered the reflection from the acrylic-medium boundaries as well as the acrylic-transducer contacts. This could be expressed as:

\[
\text{Output (Volts)} = \text{Input (Volts)} \times f \times (1 - RC_1^2) \times (1 - RC_2^2) \times A_1 \times A_2 \quad (6.5)
\]

where \( f \) is the transducer efficiency function, \( RC_1 \) is the reflection coefficient of transducer-acrylic contact, \( RC_2 \) is the reflection coefficient of the acrylic-medium contact, \( A_1 \) is the attenuation in the medium, and \( A_2 \) is the attenuation in acrylic.

In all of the experiments, input amplitude was set to be 25 Volts (controlled automatically by computer and manually by the amplifier controls) and the output was the amplitude of the recorded signals.
To calculate $RC_1$, I used the water test and calculated the reflection coefficient obtained during transmission/water tests in water. Regression of the amplitude against transducer distance data was expressed as: Amplitude = $2.9476 e^{-4.905x}$ (Figure 5.2).

Based on the equation: $Output = Input \cdot f \cdot (1-RC)^2 e^{-\alpha(x)}$

At $x=0$, we have $2.9476 = Input \cdot f \cdot e^{-\alpha(0)}$  \rightarrow Input \cdot f = 2.9476

At $x=0.2$, we have $0.99 = Input \cdot (1-RC)^2 \cdot f \cdot e^{-4.905 (0.2)} = 2.9476 \cdot (1-RC)^2$

and $(1-RC)^2 = 0.8958$ \rightarrow $(1-RC) = 0.9465$ \rightarrow RC = 0.05353

Using the transducer/water reflection coefficient, density and velocity in distilled water, I calculated the acoustic impedance for the coating material of the transducer, and then the transducer/acrylic reflection coefficient ($RC_1)$ calculated as 0.3157. $RC_2$ values were obtained using the acoustic impedance of acrylic and the cavity of the cell (Table 6.4 (a)) due to the presence of differently dissolved $CO_2$.

To obtain the value of $f \cdot A^2$, I solved equation (6.5) for the test on the cell without glass beads and only filled with distilled water, which was:

Input= 25 Volts
Output= 3.135 Volts

$(1-RC_1)^2 = (1-0.315681)^2 = 0.4683$
$(1-RC_2)^2 = (1-0.0.3684)^2 = 0.0.39892$
$A_1 = e^{-\alpha x} = e^{-4.905*0.0381} = 0.82954$

So, $f \cdot A_2 = 0.80919$ which would be constant for all of the experiments since the same pair of transducer were used as the transmitter and receiver, and attenuation within the acrylic parts of the cell was also constant for all experiments. By having all parameters of equation (6.5), I solved it to find $A_2$ attenuation values in the cavity of the cell filled with glass beads containing 0.3, 0.2, 0.1, and water as the fluid. Table 6.4 (b) shows the calculated values for attenuation in glass beads filled with differently dissolved $CO_2$. Comparing values of attenuation coefficient based on the geometry of the pressure cell, and calculated values based on the scattering equation (6.2), it can be seen that the
attenuation caused by the high concentration of packed glass beads was higher than those of diluted and unpacked solutions. This difference could be explained by “multiple scattering” of high concentration of dense scatterers (Varadan et al., 1989) as well as the intrinsic attenuation caused by wave-induced fluid flow which was not included in the equation (6.5).

Based on Table 6.5, an increase in transmission amplitude due to an increase in dissolved CO\textsubscript{2} content (increase in density) is the optimum explanation, since the attenuation coefficient decreases with greater CO\textsubscript{2} dissolved in water and more energy (amplitude) can then be transmitted through the medium (higher Q, \( \gamma_s = \frac{\pi f}{QV} \)). Also, values of attenuation (A1) in the table show that a greater decrease in amplitude in the tests with glass beads is expected than the tests without glass beads, when more CO\textsubscript{2} dissolves in water. Figure 6.18 shows a decrease in transmission amplitude from without glass beads to with glass beads (because the beads were scatterers). A greater decrease in amplitude for 0.3 mol dissolved CO\textsubscript{2} in 1000g of water was observed with less of a decrease in tests with water alone.

6.2.2.8 – Pore Size Effect on Seismic Attributes
To investigate the effect of pore sizes on seismic propagation, two sets of transmission experiments using the medium pressure cell containing 0.7 mm glass beads and 2 mm glass beads and distilled water were performed. 0.3 mol of CO\textsubscript{2} was dissolved in a litre of water at 250 psi pressure. During the experiments, pressure was reduced while the transmitted energy (through the cavity of the cell) was recorded.

Figure 6.19 (a) shows the signals transmitted through 0.7 and 2 mm beads filled with distilled water at different pressures. As can be seen in this figure, the waveforms indicate a consistent signal with a steady amplitude, changing very slightly with reducing pressure and no change in velocity of the wave observed during the pressure reduction process. The amplitude spectrum of these two sets of tests also shows a minor amplitude change
Table 6.5 (a) Reflection coefficient of glass bead-filled medium and acrylic contact calculated using an acoustic impedance of two mediums. (b) Loss of transmission energy (attenuation), A1, and attenuation coefficient calculated for all dissolved CO₂ amounts.
Figure 6.18 (a) Amplitude change due to pressure using differently dissolved CO$_2$ content- with and without glass beads. (b) Amplitude difference of the tests with and without glass beads with pressure change for each dissolved CO$_2$ contents.
when pressure (Figure 6.19 (b)) was reduced; so, pressure reduction in this case did not cause any change in amplitude or velocity of the wave. The spectral analysis of Figure 6.19 (b) showed that the 2 mm bead’s peak frequency was 320 kHz, and the maximum frequency was 400 kHz, whereas for 0.7 mm beads, the frequency range was broader (peak frequency of 500 KHz with the maximum frequency at 900 KHz). It can be mentioned that high frequency components of signal were attenuated using 2 mm beads due to scattering attenuation of the larger glass beads. This is expected according to theory, which states that the attenuation and therefore upper frequency content of a waveform is a function of the wavelength/inhomogeneity size ratio. That is, with 3.8/2 size ratio (R=1.9) for the 2 mm beads compared with 3.8/0.7 size ratio (R=5.4), there is greater attenuation of high frequency content with the larger 2 mm beads (Muller et al., 2002; Auriault et al., 2002). The larger the size ratio R, the higher the received frequency range (ie. lower attenuation).

The velocity in the tests using 2 mm beads (2309.1 m/s) is higher than when using 0.7 mm beads (2005 m/s). This could be because the wave travels a greater refracting distance through the glass bead’s material when glass beads are larger.

Figure 6.20 (a) shows the transmitted signal through 0.7 and 2 mm beads filled with 0.3 mol dissolved CO\textsubscript{2} content in water. As can be seen in this figure, in both experiments signal started to attenuate below 100 psi which was calculated to be the bubble point for 0.3 mol dissolved CO\textsubscript{2}. When CO\textsubscript{2} as gas bubbles evolved from the fluid, no velocity change was observed in either set of experiments, but a large reduction in amplitude occurred because of gas bubbles attenuating the signal. Amplitude spectra in this case also showed a decrease in amplitudes at all frequencies (Figure 6.20 (b)).

Figure 6.21 (a) combines all values of peak to peak amplitudes versus pressure. The error bars represent 5% error margins for each data point, and what is immediately apparent is:

1. Transmission amplitudes for the smaller beads are higher than for the larger beads irrespective of water or water containing dissolved CO\textsubscript{2}. 
Figure 6.19. (a) Signals transmitted through 0.7 and 2 mm glass beads filled with distilled water. (b) Minor change in amplitude spectra of signals transmitted through 0.7 and 2 mm glass beads in different pressures could be observed.
2. Amplitudes are greater when the pore space contains dissolved CO₂, because the density is greater. There is a steady difference in amplitude between the two.

Figure 6.21(a) also raises the point that Henry’s Law predicts the bubble point as a point where bubbles are optically observed (100 psi in my experiments). However, the seismic transmission method used in these experiments demonstrates that ultrasonic transmission is sufficiently sensitive to detect any miniscule bubble formation (at 100 psi) before they are optically observed (60 psi).

From Figure 6.21(a), it is apparent that the amplitude values for 0.7 mm beads moved upwards with higher pressure. It is believed that this could be because some trapped gas bubbles expand marginally as pore pressure is reduced (and it also could be possible that the beads were reseating themselves). It was considered that the amplitude values should be flat across the pressure range. There is also the point that below the bubble point, amplitude values were decrease slower at the larger beads than the smaller beads, which indicates that the larger number of small bubbles between smaller beads attenuates (scatters) signal more so than the small number of larger bubbles between the larger beads.

Figure 6.21 (b) shows the peak frequency versus pressure graph for pure water and CO₂ dissolved in water. Clearly the peak frequency is a function of bead size alone and is not influenced by the small amount of CO₂ dissolved during these experiments in the pore water. This suggests that in practice, peak frequency as a seismic attribute cannot be used to establish CO₂ dissolved quantity, whereas previous reflection amplitude data has indicated that it might be an attribute which may be useful in that sense (McKenna, 2004). The attribute to be used is clearly transmission amplitude and spectral ratios.

6.2.3 – Accuracy of Henry’s Law
The amount of CO₂ which can be dissolved in water has been predicted using Henry’s Law as a function of temperature and pressure (Carroll et al., 1991). These predictions are based on correlation of more than one hundred measurements of different CO₂
Figure 6.20. (a) Signals transmitted through 0.7 and 2 mm glass beads filled with 0.3 mol dissolved CO$_2$ fluid. (b) Major change in amplitude spectra of signals transmitted through both 0.7 and 2 mm glass beads could be observed below bubble point (100 psi).
Figure 6.21. (a) Amplitude of transmitted signal through 0.7 and 2 mm beads containing water and 0.3 mol dissolved CO$_2$. (b) Peak frequency is not affected by fluid content (distilled water or dissolved CO$_2$) and only decreased below bubble point (100 psi).
quantities dissolved in water at different pressure and temperatures (less than 1 MPa pressure with temperatures between 0 to 160°C) which has been included in Henry’s constant equation. Since much of these reportings were based on optical observation of CO₂ bubble formation (evolved from dissolved CO₂ mixtures), the experiments I did using ultrasonic transducers could investigate the accuracy of such measurements.

In Figure 6.8(a), different amounts of CO₂ were dissolved in 1000 g of distilled water (0.3, 0.2, and 0.1 mol equal to 13.2, 8.8, and 4.4 gr of pure dry ice, respectively). By decreasing the pressure from 160 psi down to 0 psi, CO₂ evolved from dissolved to its free gas form. The existence of CO₂ bubbles affected the ultrasonic waves and caused the energy to be attenuated; this caused a decrease in transmission amplitude. I used the graphs that predict amounts of CO₂ which could be dissolved in water and compared the results with observations in the experiments. I changed the scale of the amount of CO₂ I used in my experiments from mol to mol%, to allow comparison with the values of predictions based on Henry’s Constant. Table 6.6 shows the re-scaled values for dissolved CO₂ content and also observed and expected pressure for CO₂ phase change from dissolved to a free gas. Based on the table, ultrasonic transmission tests I performed are in good agreement with Henry’s Law predictions. However there is a small difference caused by small size CO₂ bubbles (at high pressure when CO₂ bubbles start nucleating) which didn’t affect the acoustic wave. Errors in reading precise values of amplitude in the experiments can equally result at each 10 psi during the pressure reduction process.

Figure 6.22 compares the predicted (Carroll et al., 1991) versus measured bubble point pressures for 0.1, 0.2 and 0.3 mol dissolved CO₂ in a litre of water. Error bars shows a 5% error margin and also shows that measured values for bubble point (obtained using seismic transmission) are within 5% of predicted values, suggesting that the sonic approach to evaluating bubble point is at least as accurate as physical observation.
Table 6.6. At constant temperature (20°C), dissolved CO$_2$ changed phase to free gas at different pressures as a function of the quantity of dissolved CO$_2$ in water. Results of the experiments confirm the prediction work of Carroll et al., (1991).
Figure 6.22. Good agreement between predicted bubble point based on Henry’s Law and measured bubble point for 0.1, 0.2 and 0.3 mol dissolved CO$_2$ in a litre of water.
6.2.4 – Instantaneous Phase and Frequency Study

Seismic attributes have been commonly classified for use during reflection seismic interpretation, and they are classed in terms of groups of attributes (Rock Solid Images, 2007). ‘Instantaneous phase’ and ‘frequency’ are attributes which are computed sample by sample, and represent instantaneous variations of a number of parameters. Instantaneous values of attributes such as trace ‘envelope’, and its derivatives, may be determined from complex traces using the basic Hilbert Transform (Taner, 2001).

To investigate whether instantaneous frequency and phase are sensitive to fluids in a rock matrix or not, and whether they can be used to quantify differently dissolved CO\(_2\) fluids, I produced a number of attribute maps for the experiments with 0.3, 0.2, 0.1 and 0 mol CO\(_2\) content dissolved in a litre of water. I also produced attribute sections of the test with the glass bead filled cell containing 0.3 mol dissolved CO\(_2\) as a fluid, at different pressures to see whether these attributes were useful for examining CO\(_2\) bubbles at low pressures. Figure 6.23 shows the seismic records of the tests without glass beads at different dissolved CO\(_2\) content and the instantaneous phase and frequency attribute maps. As can be seen in this figure, in the signal area, minor changes in instantaneous frequency were observed due to a decrease in dissolved CO\(_2\) content, whereas almost no change occurred in instantaneous phase. Figure 6.24 shows the same data, but this time for the tests with glass beads as the rock matrix. Again in this Figure, there was no change in these attributes. It seems that these attributes were not sensitive to differently dissolved CO\(_2\) in fluid, which was disappointing.

Figure 6.25 compares the instantaneous frequency and phase maps with seismic records for the tests using 0.3 mol dissolved CO\(_2\) in the cell without glass beads. This Figure shows that in the signal area, there was no change in instantaneous frequency and a minor change in instantaneous phase attributes when there was pressure reduction. This was considered reasonable since no gas bubbles evolved at the high pressures- however these attributes in my experiments were not pressure sensitive either. Figure 6.26 shows the same maps for 0.3 mol dissolved CO\(_2\) on the cell filled with glass beads. Based on this figure, instantaneous frequency showed an increase during pressure reduction (CO\(_2\)
bubble formation). It seems that instantaneous phase events have shifted at lower pressure and appear earlier. Instantaneous frequency instead does not show any change during bubble formation. Again, it seems that these attributes are insensitive to CO$_2$ phase change and cannot be used for quantification of dissolved CO$_2$ mixtures.

With regard to the amplitude spectrum of 0.3 mol tests at different pressures (Figure 6.16), it can be concluded that the dominant frequency of the spectrum reduces during pressure reduction and bubble formation (Figure 6.27 (a)). The centre frequency of the fitted curves on the frequency data also shows a decreasing shift in frequency when CO$_2$ bubbles appear (Figure 6.27 (b)). It shows that high frequencies attenuate more than low frequencies.

6.3 – Summary of the Chapter
Different analyses were performed using ultrasonic transmission measurements recorded with pressure cells. Initial tests showed that both using 50 KHz and 250 KHz transducers, the velocity of the P-wave did not change with respect to the different quantities of CO$_2$ dissolved in water, or to the different pressure, whereas the transmission amplitudes changed due to both dissolved CO$_2$ content and pressure change (it reduced when CO$_2$ bubbles appeared out of their dissolved form by decreasing the pressure below the predicted bubble point). Studying the Q-factor on the tests with 250KHz showed that more CO$_2$ in solution will cause an increase in quality of transmission (and consequently greater transmission amplitude).

The medium pressure cell was used for reflection tests and based on these tests some modifications were made on it which made the design better for transmission tests using 1MHz ultrasonic transducers. Using this cell it was possible to dissolve more CO$_2$ at high pressure than previously and the effect of CO$_2$ on seismic attributes could be studied better when more CO$_2$ was dissolved in water. This time the tests were performed in two steps: tests without any rock frame just with fluids, and the tests performed on the cell when it contained different simulated rock matrices.
Observation of the medium pressure cell filled with just fluid, proved that while there was a maximum change in P-wave velocity (of only 5%) due to the different quantities of CO₂ dissolved in water, there was a greater change in transmission amplitude (17%) which introduced transmission amplitude as a better attribute than velocity for dissolved CO₂ quantification purposes. These tests showed that pressure (which was kept above the predicted bubble point) did not affect either the velocity or amplitude of the transmitted energy. Comparing the accuracy of these measurements with those of experiments on velocity observed by Batzle and Wang (1992), accuracies here were very high (0.9% difference between these two datasets).

Tests on the cell containing 0.7mm glass beads as a rock frame showed no change in velocity whereas changes in transmission amplitude were observed. The change in amplitude was less compared to the tests without glass beads, which indicates that transmission amplitude is very sensitive to the quantity of fluid (less fluid within the pores of the rock frame caused less change in transmission amplitude). Also the P-wave velocity due to a smaller quantity of fluid in the pores, did not show any change due to different amounts of dissolved CO₂ content. A slight change in amplitude above the predicted bubble point suggested that some air was trapped between the grains of the matrix which could not be removed with vacuuming or fluid injections or flushing. However, the amplitude reduction was very large when pressure was decreased below the bubble point and CO₂ changed phase from the dissolved state to a free gas. This large decrease in transmission amplitude observed in my experiments was confirmed using the calculation for CO₂ solubility based on Henry’s Law.

Analysis of both with and without glass beads showed that the density of fluid (which increases when dissolving CO₂ in it) could be related to transmission amplitude and by studying the change in transmission amplitude, it was possible to predict the change in density (and subsequently change in dissolved CO₂ content).

The scattering attenuation, as the main reason for the reduction in transmitted energy through the medium, was also investigated. The calculated attenuation was higher than
what theories predict which could be because the theories had not considered the inter-
granular effect on wave scattering. Attenuation due to CO$_2$ bubbles was also studied and
discussed in this Chapter with regard to the spectral amplitude content.

The effect of a matrix grain’s size on ultrasonic energy when dissolved CO$_2$ and water
filled the cell was also studied and discussed in this Chapter and at the end of the
Chapter, discussions on peak and dominant frequency of the transmitted signal’s spectra
as potential seismic attributes to predict CO$_2$ phase change, were explained. Chapter 7
will now provide recommendations and suggestions for future work on quantifying the
CO$_2$ effect on seismic waves, with respect to the experiments and analysis I did.
Figure 6.23. Instantaneous phase and frequency of the transmission records through the cell filled with different dissolved CO$_2$ content without glass beads.
Figure 6.24. Instantaneous phase and frequency of the transmission records through the cell filled with different dissolved CO₂ content with glass beads as rock frame.
Figure 6.25. Instantaneous phase and frequency of the transmission records through the cell filled with 0.3 mol dissolved CO$_2$ without glass beads at different pressures.
Figure 6.26. Instantaneous phase and frequency of the transmission records through the cell filled with 0.3 mol dissolved CO$_2$ with glass beads at different pressures.
Figure 6.27 (a). Curves fitted on amplitude spectra of transmission signal for 0.3 mol dissolved CO$_2$ at different frequencies (b). Both centre and dominant frequencies show a decrease while CO$_2$ bubbles evolve from dissolved mixture.
CHAPTER 7 – DISCUSSIONS, CONCLUSIONS AND RECOMMENDATIONS

Increasing the use of fossil energy resources (oil, gas and coal) over the next decades will cause more carbon dioxide to be released into the atmosphere and consequently enhance the greenhouse gas effect which causes global warming, which is considered to result in more droughts and floods. A good way to decrease CO\textsubscript{2} emissions to the air, is the sequestration of CO\textsubscript{2} in different geological storage sites. Injecting the CO\textsubscript{2} into the ground will cause most of the CO\textsubscript{2} to replace the formation fluids in areas where it can be directly trapped while in other areas it will become dissolved in the interstitial matrix fluid as a function of the reservoir’s pressure and temperature. Attempts have been made to monitor CO\textsubscript{2} fluid movement underground and in evaluating the quantity which appears as a seismic anomaly (Arts et al., 2004) but little research has studied the dissolved component (which has the potential to resolve volumetric assessment errors) apart from that reported within the body of this thesis.

Using ultrasonic transmission experiments through a physical model of a reservoir in which different amounts of dissolved CO\textsubscript{2} passed through different sizes of matrix grains during changes in pressure, it was possible to monitor and verify the seismic response using any particular model. The objectives of my research were to use seismic transmission data recorded on a physical model to:

- Investigate the sensitivity of different seismic attributes to changes in fluid phase;
- Investigate the sensitivity of different seismic attributes to dissolved CO\textsubscript{2} content;
- Quantify dissolved CO\textsubscript{2} content in the pre-bubble condition;
- Predict scattering, dispersion and attenuation effects due to different fluids and scattering matrices;
- Study the effect of CO\textsubscript{2} as a free gas on seismic wave propagation;
- Investigate the predictions of CO\textsubscript{2} dissolution in water for comparison with Henry’s Law;
- Provide an initial seismic data base for the application of attenuation studies in VSP or amplitude tomography, which thereafter would be a guide to time-lapse amplitude versus dissolved content changes.
The data analysis began using a fluid property simulator of Li and Siggins (2006) to understand the change in elastic moduli of fluids during pressure change. Comparisons and analyses were then made with transmission test observations acquired during pressure reduction within the physical models. Most of these observations were accurate and controlled data had good agreement with numerical modelling but opened up new areas of knowledge (such as in terms of amplitude variation) where numerical modelling is presently non-existent. The major conclusions and recommendations of my research work are summarised below.

7.1 – Rock Physics
The injection of CO$_2$ into a reservoir will cause significant change in the elastic properties of the reservoir over time, since the bulk modulus and density of CO$_2$ could be less than that of ambient brine, and it has been shown by numerical modelling that free CO$_2$ will be easily monitored using repeated seismic reflection measurements (Arts et al., 2004). Results of my thesis show that despite previous seismic work to study dissolved CO$_2$, when sequestered CO$_2$ dissolves into the formation water (distilled water for my experiments), it is not only possible to distinguish the change in seismic amplitude, but also it is possible to quantify the amount of dissolved CO$_2$ (above the bubble point) using arrival amplitude variations, whereas there was very little change in seismic velocity. The change in seismic amplitude was more apparent in tests having no matrix while containing solely fluid, which indicates that transmission amplitude is sensitive to variations in pore fluid properties.

It has been mentioned that during the CO$_2$ sequestration process, silicate and carbonate dissolution in CO$_2$ fluids will cause acidic conditions that could cause an increase in porosity, while the crystallization of carbonates over the time of sequestration (mineral trapping) will cause a decrease in porosity and increase in elastic moduli. Both of these mechanisms may be observable in the seismic data over long periods in time. However, over the short period of this research, I have observed changes which result from variation in pore fluid density, which could have a similar seismic response to
mineralization albeit over an almost immediate time period. This would be similar in terms of seismic imaging to the mineralization “rind” suggested by Johnson et al. (2001) and could explain the reason for the seismic imaging of a possible rind soon after injection commencement at Sleipner.

7.2 – CO\textsubscript{2} State of Phase
A change of CO\textsubscript{2} phase from the state of being dissolved in water to a free gas was investigated by reducing the pressure in order that CO\textsubscript{2} was liberated from its dissolved form. This change in CO\textsubscript{2} phase was displayed as a function of amplitude attenuation using transmission data, whereas in the tests performed using dissolved CO\textsubscript{2}, the signal during the pressure reduction process decreased dramatically due to the significant attenuation by CO\textsubscript{2} bubbles. When gas was liberated from the mixture, it was not possible to measure the first arrivals during transmission through an uncemented pack of glass beads filled with CO\textsubscript{2} gas. From the analysis it is apparent that some attributes such as dominant and centre frequency of the spectra are also sensitive to CO\textsubscript{2} phase change from dissolved to a free gas form. However, they were not the best attributes for quantifying the CO\textsubscript{2} in solution..

The CO\textsubscript{2} pre-bubble phase was studied using the transmission method to provide a basic knowledge of the effect on amplitude variation for quantifying dissolved gas content, which may have an application in Gas-to-Oil-Ratio mapping in oil and gas reservoirs. This could provide a control on drilling and plans for production from fields that are loosing pressure through depletion.

7.3 – Using Seismic attributes to Estimate Dissolved CO\textsubscript{2} Content
In this research the attributes of seismic properties being P-wave velocity, amplitude, peak frequency, and instantaneous frequency and phase were used to determine the amount of dissolved CO\textsubscript{2} in water. Also the scattering effect resulting in changes in attenuation was studied, while the effect of gas bubbles on seismic wave peak frequency was of interest. Additionally, attributes were also studied to monitor CO\textsubscript{2} phase change from a dissolved to a free gas. These will now be discussed.
7.3.1 – Seismic velocity

Velocity modelling software locally developed by Li and Siggins (2006) was based on elastic moduli presented by Batzle and Wang (1992). This provided the velocity through water and CO₂ saturated water at different pressures. Experiments showed that the measured ultrasonic transmission velocity in the pressure cell filled with water at different pressures were in good agreement with the modelling results, but the measured velocity in water containing dissolved CO₂ was slightly faster than numerical modelling results possibly because of the use of ultrasonic frequencies.

There was no methodology or science which predicted seismic velocity changes due to different levels of CO₂ dissolved in water, but my transmission experiments showed that dissolving different amounts of CO₂ in a litre of water will not cause significant change in velocity anyway, especially when porosity was low (in the case where there is a matrix); however in the experiments which did not contain a matrix a small change in velocity was observed. So, seismic velocity was considered too insensitive to be used for quantifying the amount of dissolved CO₂.

As was mentioned previously, seismic velocity was unable to be used in monitoring CO₂ phase change from a dissolved to a free gas due to strong attenuation caused by CO₂ bubbles in a matrix when it consisted of uncemented glass beads at low pressures; however, ultrasonic transmission amplitude was sensitive enough to different sizes of glass beads.

7.3.2 – Seismic Amplitude

The experiments showed that seismic amplitude is an attribute for quantifying the dissolved CO₂ content in water. During the pre-bubble phase of CO₂, amplitude was shown to be related to the amount of dissolved CO₂. This research has demonstrated that amplitude is sensitive to the amount of dissolved CO₂ and can be used as the attribute in time-lapse transmission measurements (especially VSP and tomography which deal with directly arriving seismic waves). Based on the results of data analysis, it has been shown
that seismic amplitude has a close relationship to density changes of the host fluid (which is a function of the quantity of dissolved CO$_2$).

Seismic amplitude was also sensitive to CO$_2$ phase change from dissolved to a free gas, when amplitude attenuation was caused by the presence of gas resulting in a decrease in arrival amplitudes, so the CO$_2$ phase change can be monitored by this attribute.

### 7.3.3 – Other Attributes

Analysis of different CO$_2$ content tests showed that instantaneous phase and frequency were not sensitive to dissolved CO$_2$ content. The tests using the same amount of dissolved CO$_2$ showed that instantaneous frequency was very sensitive to pressure change in the case where gas was liberated. A decrease in the value of instantaneous frequency made it a good indicator of the increasing presence of gas as the bubble point is reached and passed.

Values for seismic attenuation and spectrum peak frequency were also studied in this research. Both peak frequency and the dominant frequency spectrum were very sensitive to fluid phase change. Both decreased considerably when CO$_2$ bubbles formed as a result of a pressure reduction. Peak frequency was not a good indicator of the amount of dissolved CO$_2$ however, since there was little change due to differently dissolved levels of CO$_2$ in fluids. It should be mentioned that peak frequency was sensitive to CO$_2$ phase change from dissolved to its gas form and strongly affected by the presence of CO$_2$ bubbles resulting from pressure reduction within the cell. Results of the tests also showed that the peak frequency was insensitive to different fluids (distilled water and dissolved CO$_2$ fluids), nor was it sensitive to different matrix pore size (0.7 versus 2 mm glass beads). It seems that peak frequency could only be a good indicator of fluid phase change from dissolved to free gas and no more (in contrast to the postulation posed by McKenna, 2004).

My studies show that the greater the amount of dissolved CO$_2$, the less the attenuation of seismic energy. This suggests that the amplitude tomography technique may be used in
quantifying dissolved CO₂ content on the certain conditions which error envolved in operation is less than the change in amplitude, where velocity tomography fails due to inadequate sensitivity to dissolved CO₂.

7.4 – Seismic Sensitivity to Solubility of CO₂
Seismic waves are sensitive to the presence of bubbles. Bubbles cause scattering in seismic waves and as a result reduce the transmission amplitude. This fact was used in my thesis to investigate the accuracy of CO₂ solubility calculations based on Carroll et al. (1991). Using the experimental data, seismic attenuation (due to the presence of CO₂ bubbles) occurred at the same pressure (at room temperature) that Henry’s Law suggested. The experiments confirmed the accuracy of the calculations made and based on Henry’s Law. It also proved that the methodology used for developing dissolved CO₂ was correct.

7.5 – Seismic Sensitivity to Pore Size
Analysis of recorded arrivals showed that ultrasonic velocity was sensitive to a rock frame’s particle size with a water-filled matrix (using 0.7 mm and 2 mm glass beads). Seismic velocity increased from 2005 m/s for 0.7 mm beads to 2309 m/s for 2 mm glass beads. Seismic amplitude was also sensitive to the frame’s particle size, since it decreased when scatterers (beads) were larger in size. Another attribute used for studying the effect of differently sized matrix particles was peak frequency. This attribute decreased when larger glass beads were used in the experiments, so it seems that this attribute is sensitive to the size of the pores (which could be related to permeability of the matrix).

It should be mentioned that based on the experiments, only seismic amplitude was sensitive to the fluid content, and other attributes did not change when changing fluid content.

7.6 – Applications to Direct Wave Arrivals (VSP) and Reflections (Time-Lapse Seismic)
The methodology of experiments in my thesis was mostly based on the transmission seismic method. The results of the research showed that a time-lapse study of transmission signals can be used in not only monitoring the shape and location of CO$_2$ in formation water, but also in verification and quantifying the amount of CO$_2$ dissolved in water. Vertical Seismic Profiling (VSP) and Tomography methods are the seismic methods which deal with directly transmitted signals, so the results of this thesis can be applied to these seismic methods.

7.7 – RECOMMENDATIONS

7.7.1 – Transmission and AVO (Reflection) Data Comparison
This research has shown at the ultrasonic scale that transmission seismic amplitude is the seismic attribute to verify changing dissolved CO$_2$ content in water. To make this new knowledge more related to conventional seismic, reflection tests with the same pressure and temperature conditions are required. Since it is apparent that seismic transmission amplitude responds to pore fluid density changes, in like manner to seismic reflectivity, then interpreting amplitudes in terms of reflection data such as AVO may be useful in indicating fluid density which is then related to different dissolved CO$_2$ content. Thus the comparison between these two data sets (transmission and reflection) offers the potential for a more accurate determination of the change in dissolved CO$_2$ content.

7.7.2 – Complementary High Pressure Tests
The experiments at low pressure provide a base line for defining a seismic attribute which senses changes in dissolved CO$_2$. Repeated tests at high pressures can provide more realistic results in terms of reservoir conditions. High pressure tests can also provide higher seismic sensitivity to CO$_2$ bearing fluids since more CO$_2$ at high pressure dissolves in water. It is also the case that performing transmission tests at high temperature brings the tests to the level of true reservoir conditions and with a larger model, the seismic transmission frequencies have the potential to reduce more towards real-world field values.
7.7.3 – Investigating the Frame size Effect on Seismic Transmission

Analyses of the tests with different sized glass beads were performed, in which the frame porosity was the same but the volumetric proportion of fluid between the grains was different (with regard to larger porosity provided into pore space between grains of the large glass beads). Repeating the controlled transmission tests on the different particle sized frames could provide an indication of volumetric changes due to dissolved CO₂ within a different rock framework. Further examination of current scattering theories may help establish new theories and understanding of both transmission and reflection seismic attributes.

7.7.4 – Unexpected Results Caused by Trapped Air Bubbles

One unexpected phenomenon which occurred during the tests with the glass beads was the reduction in amplitude above the bubble point as pressure was decreased. This decreasing amplitude trend could be seen even during the test with distilled water as the fluid. While a decrease in pore pressure should cause the grains of the beads to have better contact and consequently for the transmission amplitude to increase, observations showed that amplitude decreased which meant that the effect of the gas bubbles was greater than that of greater grain contact. Also, energy loss because of frictional force decrease would be expected to be higher when grains are loosely compacted, whereas there was a higher amplitude recorded in this situation during my experiments. A good explanation may be found in the fact that residual gas bubbles were trapped in small pores and these grew bigger during the decreasing pore pressure, resulting in a slight attenuation increase and consequent reduction in transmission amplitude.

In the experiments I performed, I used flushing and a vacuum to remove trapped gas bubbles. This suggests that in future experiments, a stronger vacuum pump is needed to minimize the effect of trapped gas. It would be needed even more so when experiments are to be performed on cemented frames, such as sandstone. Applying a higher flow rate during the flushing step may also assist a reduction in the amount of trapped gas.
7.8 – FUTURE RESEARCH

7.8.1 – Ultrasonic Frequency Effect on Dissolved CO₂

Ultrasonic signals with different frequencies transmitting through the same media provide different $\frac{\lambda}{d_s}$ values. In terms of rock physics, these values can be related to different scattering mechanisms which affect seismic wave propagation. Changes in the characteristics of seismic wave transmission due to different scattering mechanisms may help in more accurate quantification of CO₂ dissolved in water. So, a recommendation is that the tests be repeated at different frequencies to provide better conditions for explaining scattering theories. Also at different frequencies, amplitude attenuation can continue to be studied as an important attribute, as well as scattering and dispersion effects on the waves with different spectra.

7.8.2 – The Frame Effect

In this research, to simplify the performance of the tests and to ignore any unwanted effects resulting from variabilities of frame particle shape and size distribution, I chose round glass beads. There are limitations in rock physics relations when uncemented and uncompacted grains are used, which is usually not the case for deep reservoirs, so it is suggested that using a frame made of cemented grains (like sintered glass beads, synthetic or naturally cemented but highly porous sandstone) may help to apply the effects of confining pressure change on the frame as well as pore pressure on the fluids. By adding the results from simple to more complex situations, it is recommended that the tests be repeated with a frame that is more realistic of consolidated reservoir rock. Using a cemented frame may also decrease uncertainties and noise caused by inter-granular friction which may affect the signal attenuation and change measured transmission amplitudes.

The tests with different grain sizes may also give an indication of how seismic transmission attributes can be used to study the permeability of the rocks, since using
round glass beads of a consistent size will not cause a change in porosity whereas it results in a change in the permeability of the medium.

7.8.3 – Tests Performed on CO\textsubscript{2} in Different Phases
Successful monitoring and quantification of CO\textsubscript{2} storage in a saline aquifer requires a focused study of the CO\textsubscript{2} in its multi-phase conditions which conventionally occurs during CO\textsubscript{2} sequestration operations. By understanding the effect of dissolved CO\textsubscript{2} on seismic velocity, amplitude and subsequent attenuation, it is necessary to do the tests on the same physical model with CO\textsubscript{2} in different phases (gas, liquid, and supercritical) and compensate the effect of dissolved CO\textsubscript{2} to establish new knowledge of the optimum approach for the interpretation of location and quantification of CO\textsubscript{2} in-place.

7.8.4 – Effective Pressure and Pore Pressure
In conventional seismic analysis, the pressure that affects the elastic properties of the frame is not only pore pressure but confining pressure. In the tests I performed, the confining pressure was provided by compressing the glass beads in one direction and during the experiments the confining pressure was fixed and could not change. That was the reason for all analyses to be based on pore pressure rather than effective pressure. My suggestion for future work is that the pressure cell should provide confining pressure in all directions and be capable of changing this confining pressure independent of the pore pressure to allow the study of pressure in terms of effective pressure which is commonly used in rock physics.

I also recommend applying a higher effective pressure to provide a pressure which is closer to a reservoir’s true pressure in nature.

7.8.5 – Developing Technologies
Numerous technologies are being developed to monitor gas storage in the subsurface. The use of permanently installed seismic sources and receivers will improve the repeatability of time-lapse seismic data (because of fixed geometry and coupling) in VSP and Tomography measurements. Although borehole technologies do not provide large
lateral coverage, new cross-well profiling techniques such as cross-well resistivity profiling are being investigated. Integrating these new technologies with seismic has made them very useful while integration workflows are a developing art form. Further investigation of natural and induced tracers may also be required if CO₂ is to be stored in reservoirs of close proximity to operating oil and gas fields. Lastly, despite the best efforts to monitor CO₂ underground, surface environmental monitoring may be required to ensure storage does not present a health, safety and environmental risk, especially in areas close to a population centre. To ensure that any possible leakage of CO₂ is detected, atmospheric CO₂ detectors are being used, as well as soil gas measurements and vegetation monitoring.

It is hoped that the research results provided in this thesis, will form the basis of the ability to remotely quantify any CO₂ in-place to avoid any seepage from a rock formation, and thereby allow better operational decisions to be made to ensure good management of the sequestration process.
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Carbon dioxide, retrieved from:


REFERENCES


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