A Study of In-situ Stress Magnitudes in the North Sea Basin from Borehole Measurements

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Abstract
The stress field in the crust is a fundamental first order geophysical property that is intimately linked to the dynamic behavior of the Earth. This project has concentrated on crustal stress magnitudes as these are generally less well understood than stress orientations. Leak-off tests have been evaluated as a potential method of stress magnitude estimation, and the most suitable uses of leak-off data to estimate stress magnitudes have been defined. Vertical stresses have been estimated using geophysical logs. The variations of stress with depth, geographic domain, lithology and pore pressure have been has been studied in order to investigate the origins of crustal stress in the North Sea basin.

The most reliable method of stress magnitude determination is the hydraulic fracturing (hydro-frac) method, however, hydro-frac data is rare. By contrast the leak-off test is performed routinely by the oil industry, with several tests in each hole drilled. An extensive 3-dimensional dataset of leak-off pressures therefore exists for the North Sea. Datasets have been obtained for the southern North Sea, and also from onshore boreholes drilled by UK Nirex, where leak-off tests and hydro-fracs have been performed in the same holes. This has enabled the leak-off test to be evaluated as a possible stress determination method. From these datasets, it is concluded that the trends of leak-off pressure with depth reflect changes in the minimum horizontal stress magnitude ($\sigma_h$) with depth. Where leak-off test pressure records are available, it is seen that the shape of many leak-off test pressure/volume plots resemble those of hydro-frac re-opening plots, and that in these leak-off tests, the leak-off pressure is very close to the hydro-frac determined $\sigma_h$. Furthermore, when leak-off tests are conducted carefully, a slightly extended test procedure can yield even better estimates of $\sigma_h$.

Over 3,000 leak-off test results have been obtained from throughout the North Sea. The trends of the leak-off test pressures with depth are investigated and from this data it is inferred that $\sigma_h$ varies with geographic location within the North Sea, with lithology, and with formation pore pressure. These variations are investigated within the context of models of stress with depth applicable to sedimentary basins. It is concluded that although a large part of $\sigma_h$ is due to the weight of the overburden, a component of tectonic stress is also present. This component of tectonic stress is significant at all depths in the southern North Sea but increases with depth in the northern and central North Sea to become significant only at depths of several thousand feet and more. This is consistent with the pattern of borehole breakouts seen from previous basinwide studies in the North Sea.

Where possible, estimates of $\sigma_h$ have been combined with borehole breakout measurements from 4-arm dipmeters. This integrated data has been analysed, within the framework of a suitable rock failure criterion, to estimate a lower bound for the magnitude of $\sigma_h$. Results indicates that the stress regime in the North Sea is predominantly a strike-slip regime.

The trend of $\sigma_h$ with depth established for the central/northern North Sea in this study is compared to trends established by other workers in a variety of geological settings from around the world. It is seen that at shallow depths, $\sigma_h$ in the North Sea (and other relatively young sedimentary basins) is low compared to other regions. However, at
depths of several thousand feet and greater, the level of $\sigma_h$ is equal to or greater than many other regions around the world. The high levels of $\sigma_h$ at depth in the North Sea could be partly due, at least in part, to the effects time dependent of creep process.
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List of Symbols and Abbreviations
Symbols and abbreviations are listed below where they are not in common use or are not defined at the point used.

Symbols

\( \sigma_{ij} \): Deviatoric stress tensor
\( p_{ij} \): Hydrostatic stress tensor
\( \sigma_m \): Mean stress
\( \sigma_d \): Differential stress
\( \sigma_1 \): Maximum principal stress
\( \sigma_2 \): Intermediate principal stress
\( \sigma_3 \): Minimum principal stress
\( \sigma_v \): Vertical stress \( Sv \)
\( \sigma_h \): Maximum horizontal stress \( Sh_{max} \)
\( \sigma_h' \): Minimum horizontal stress \( Sh_{min} \)
\( \sigma' \): Effective stress
\( p \): Pore pressure
\( \alpha \): Biot's poroelastic coefficient
\( \xi \): A poroelastic constant
\( K \): Bulk modulus (incompressibility). Bulk modulus (incompressibility) of the rock framework in the context of a poroelastic material
\( K_s \): Bulk modulus of the solid grains of rock in the context of a poroelastic material
\( G \): Shear modulus (Rigidity)
\( \nu \): Poisson's ratio in isotropic materials or Poisson's ratio measured in the plane of isotropy in transverse isotropic materials
\( \nu' \): Poisson's ratio measured normal to the plane of isotropy in transverse isotropic materials
\( E \): Young's modulus in isotropic materials or Young's modulus in the plane of isotropy in transverse isotropic materials
\( e_h \): Minimum horizontal strain
\( e_h' \): Maximum horizontal strain
\( \rho \): Density
\( \rho_m \): Density of the mantle
\( \rho_c \): Density of the continental crust

\( z \): Vertical depth or crustal thickness

\( \phi \): (Internal) friction angle

\( \mu \): Coefficient of friction (tan(\( \phi \))

\( C_u \): Cohesive strength

\( \tau \): Shear stress

\( \sigma_i \): Tectonic stress component

\( V_p \): P wave velocity

\( V_s \): S (shear) wave velocity

\( \sigma_r \): Radial stress

\( \sigma_\theta \): Circumferential stress

\( \tau_\theta \): Tangential shear stress

\( \sigma_z \): Stress parallel to the wellbore axis

\( \sigma_a \): A uniaxial stress

\( \Delta p \): Increase in pore pressure above the original pressure (during hydrofrac pressurisation of an interval with a permeable wellbore wall)

\( r \): Distance from the centre of the wellbore (the origin)

\( a \): Radius of the hole or wellbore

\( \theta \): Angle measured from the direction of \( \sigma_a \)

\( p_b \): The breakdown pressure

\( p_p \): Fracture propagation pressure

\( p_{sip} \): Instantaneous shut in pressure

\( p_r \): Fracture re-opening pressure

\( p_c \): The closure pressure

\( p_W \): Fluid pressure applied within the wellbore.

\( T \): Tensile strength

\( K_i \): Stress intensity

\( K_{ic} \): Critical stress intensity or fracture toughness.

\( \phi_p \): Porosity

\( \phi_{pf} \): Porosity at sea floor

\( \alpha_c \): Coefficient of the tensile strength (see section 6.2.3.2)
Abbreviations

EMW: Equivalent mud weight
MD: Measured depth below rotary table
BSF: (vertical depth) Below sea floor
TVD: True vertical depth
TVDSS: True vertical depth below sea surface
TVDGL: True vertical depth below ground level
LOP: Leak-off pressure
LTP: Limit test pressure or leak-off to pressure
LT: Limit test or leak-off to
LCP: Lost circulation pressure
LOLTP: Leak-off and limit test pressures
psi: Pounds per square inch
Δ1st LOP: 1st LOP minus σₙ
ΔLOP: 1st LOP minus 2nd LOP
Δ2nd LOP: 2nd LOP - σₙ
BVG: Borrowdale Volcanic Group
BB: Brockram Breccia
SBS: St. Bees Sandstone
CS: Calder Sandstone
ΔLOLTP: Difference between the LOLTP measured at the depth of interest and the average LOLTP at that depth (ΔLOLTP = LOLTP measured - LOLTP average at that depth)

ΔP: The difference between the measured pore pressure and the hydrostatic pore pressure at the depth of interest (ΔP = P measured - P hydrostatic at that depth)

Δpₘw: The difference between pressure due to the mud column (pₘw) in the wellbore and hydrostatic pressure of formation pore fluid at that depth (Δpₘw = pₘw - P hydrostatic at that depth)

ΔLOLTP (norm.): ΔLOLTP/depth
ΔP(norm.): ΔP/depth
Δpₘw (norm): Δpₘw/depth
Shmin: minimum horizontal stress or $\sigma_n$
Shmax: maximum horizontal stress or $\sigma_H$
Sv: vertical stress or $\sigma_v$
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Chapter One - Introduction

1.1 Preamble: Crustal Stress as a Multi-disciplinary Issue.

The state of stress in the earth’s lithosphere is becoming an increasingly important issue in geophysics and in industry. The magnitudes and orientations of the principal stresses in the crust (the crustal stress tensor) control the deformation of the lithosphere through folding and faulting, giving rise to the mountain belts, sedimentary basins and other dynamic, topographic features of the earth. Crustal stresses are mapped and studied (within the framework of the World Stress Map for example) in an attempt to understand the relationship between the sources of stress and the motion and deformation of the plates. By understanding the interaction of stress and rock deformation within a plate tectonic context today, it is possible to better understand the processes which have shaped the Earth in the past.

Possible sources of stress in the lithosphere are many, and include tectonic forces, gravitational loading forces, and forces arising from thermal perturbations as well as numerous other smaller scale forces. The exact nature by which these forces are transmitted through the lithosphere, and the interaction (over a range of length scales) of these forces through time to contribute to the total present day stress state remain poorly understood.

On a practical level, the stress state in the upper crust is of great importance in a variety of industrial and civil engineering applications. Hence, stress measurements have been made underground for several decades for the purposes of designing underground excavations such as tunnels and mines. More recently, crustal stress has been studied for the purposes of geothermal energy extraction from hot fractured rock masses, through which the movement of fluids is very strongly controlled by the in-situ stress state.

The importance of crustal stress in several hydrocarbon exploration and production activities is becoming increasingly acknowledged, as oilwell drilling becomes more and more challenging and reservoirs become increasingly difficult to produce commercially.
Specifically, there are several activities within exploration and production which are influenced by crustal stress. The orientation and containment of reservoir stimulation hydraulic fractures are controlled by stress orientations and lithological stress contrasts. Anisotropy in the stress field can cause anisotropy in rock properties such as permeability and wave speed, which can clearly have a profound influence within several areas of exploration and production. Throughout the lifetime of a reservoir, as the fluids are extracted, the evolution of the stress state controls the poroelastic deformations and sometimes failure of the reservoir rocks, and should be monitored for effective reservoir management.

The area where stress information and analysis is probably most commonly sought is in oilwell drilling. As a larger number of non-vertical wells are drilled in order to maximise production, geoscientists and drillers are becoming increasingly concerned about wellbore stability problems. Both the tensile and compressive failure of boreholes are largely controlled by the in-situ stresses in the rock being drilled. Failure of the wellbore wall can lead to drilling problems such as lost circulation or stuck drill pipe. Such problems can be time consuming and thus (especially, as is becoming increasingly common, where drilling is taking place in deep water) extremely costly to the operators.

This economic reality is such that some of the larger oil field service companies, just within the last year or two, have begun rapidly expanding services to operators which include in-situ stress analysis coupled with wellbore trajectory planning and reservoir stress management. It is therefore predicted that the amount and quality of data available (of the type discussed in this study) from sedimentary basins around the world will increase greatly over the next few years, offering excellent opportunities for further collaboration between academia and industry in an attempt to better understand the nature and origin of crustal stress.

The variety of objectives of stress measurement has lead to stresses being sampled on several scales, and there is at present no easily available, absolute method for the measurement of in-situ stress against which other methods can be compared. Stress measurements which are made on small samples of recovered core can sample stresses
which are present only at these scales, such as intergranular residual stresses. Hydrofracture type stress measurements can sample stresses on the scale of one bed, where the tests are specifically conducted for that reason, or several beds, where the hydro-frac is performed for commercial purposes. Stress determinations based on earthquake focal mechanisms on the other hand, can sample stresses on the scale approaching the thickness of the crust. The relationship between stresses sampled at different scales and the effect of the stress history of the rock's present day stress state, remain some of the most challenging issues in this field.

1.2 Origin of Research

In view of the general lack of understanding of the origins of crustal stress, touched upon in section 1.1, it is not surprising that much work on crustal stress measurement has been undertaken throughout the world. A variety of measurement techniques are described in section 4.2, and some of the compilations of stress magnitude measurements are summarised in section 7.5.

The current state of knowledge on crustal stress in and around the North Sea region has been summarised by the World Stress Map Project (section 3.3.7), the European part of which is shown in Figure 2.28. It can be seen that in the North Sea, with the exception of a relatively small area west of Norway, only information on stress orientations has been obtained. The data for the North Sea shown in the World Stress Map has been augmented by a basinwide study of stress orientations from borehole breakout analysis performed in over 115 boreholes. This study was undertaken at UCL by Cowgill et al. (1993) and is also reported in Cowgill et al. (1995) and Cowgill (1994). Stress magnitude data from the North Sea however, has remained conspicuously absent.

This study has focused on stress magnitudes in the North Sea. The origins of stress magnitudes are rather less tangible than those of stress orientations, and generally stress magnitudes are more difficult to measure.

Man made hydraulic fracturing (hydro-fracing) was first recognised as a reservoir stimulation technique within the petroleum industry of the US in the late 1940s and
1950s. More recently, hydraulic fracturing has become recognised as a stress
determination technique. Since the first workshop on hydraulic fracturing as a stress
determination technique in California, 1981, and even more since the international
symposium on rock stress and rock stress measurement in Stockholm, 1986, hydraulic
fracturing has become regarded as the most reliable stress magnitude determination
technique at depths of more than several tens of meters from the surface. Very little
hydro-frac data exists from the North Sea. Leak-off test data on the other hand is
plentiful, as leak-off tests are performed routinely during oilwell drilling. Although leak-
off tests are not performed for the purpose of measuring stress magnitudes, their
similarity to the first part of hydro-fracs invites an interpretation of leak-off pressures in
terms of stress magnitudes. Moreover, previous studies of leak-off test pressures in
regions where hydro-frac data has also been obtained, shows a close empirical
correspondence between leak-off pressures and hydro-frac determined minimum
horizontal stress magnitudes ($\sigma_h$).

It was recognised at the outset of this study that a single leak-off test would not be able
to provide such an accurate measure of stress as a hydro-frac. However, it is considered
that in view of the regional variations generally seen in stress measurement data, a very
large dataset of less accurate data will give a better understanding of the nature and
distribution of stress on a basinwide scale than a very small dataset of more accurate
data.

Leak-off pressures have been used in previous studies in a rather ad-hoc way as
estimates which of $\sigma_h$. In this study, an attempt has been made to use leak-off test data
in a more rigorous way than has generally been the case previously, as it can easily be
seen that simply equating the leak-off pressure with $\sigma_h$ can lead to significant errors. For
this purpose, the leak-off pressure, and in particular the form of the leak-off test
pressure/volume plot, has been considered theoretically, in conjunction with a
comparison of leak-off pressures and hydro-frac data from the same boreholes, as well
as mechanical properties of the rocks in which the tests were performed.
The datasets obtained in this study, which contain both hydro-frac and leak-off test data, come from the southern North Sea: Dowell Schlumberger hydro-frac reports and drilling records from BP and Hamilton Brothers (BHP), and from Sellafield, onshore UK: Hydro-frac and drilling reports from UK Nirex. The datasets obtained for this study which contain just leak-off test data come from throughout the North Sea. The data is described in detail in Chapter 5.

Having established the most suitable methods of using leak-off test data for estimating minimum horizontal stress magnitudes, and having determined values for the vertical stress in the North Sea using geophysical log data, variations of stress magnitude in the North Sea are investigated.

1.3 The Structure of the Thesis
This thesis firstly outlines a geological setting and a theoretical framework in which the data in this study can be interpreted in terms of stress magnitudes. Having outlined such a framework, and described the datasets, the leak-off test and hydro-frac data is examined in order to define some methodologies for using the leak-off data to investigate stress magnitudes. Having established the nature and distribution of stress in the North Sea, the origin of these stresses is considered.

Chapter Two introduces the geological history of the North Sea. In particular, the structural evolution of the crust is considered. This is an integral part of the study, as it will be seen how the numerous complex tectonic events that the North Sea has experienced have given rise to the present day tectono-stratigraphic framework in the in-situ stresses arise. The main structural elements of the North Sea are discussed and the basinwide crustal scale structure is investigated from deep seismic profiles and gravity studies. Recent seismicity of the North Sea and surrounding regions is also reviewed in this section.

Chapter Three introduces some of the ideas and theories regarding the nature of stress in the lithosphere, its origins and distribution. Models of stress distribution with depth,
which have been proposed for other sedimentary basins around the world, are presented in detail in this chapter.

Chapter Four covers stress measurement techniques. Firstly, a broad overview of crustal stress measurement techniques is given. The majority of this chapter is concerned with the details of stress measurement techniques considered in this study. In particular, this chapter presents the theoretical framework in which hydraulic fracturing and leak-off test data can be interpreted. The history of hydraulic fracturing is briefly described and the interpretation of hydro-frac pressure/time plots is discussed. The breakdown of the borehole wall in terms of the stress around the borehole and the condition of the wellbore during pressurisation is given particular attention as it is this process which is most pertinent to the interpretation of leak-off pressures in terms of stress magnitudes. The leak-off test procedure and previous interpretations of leak-off pressures are discussed. Leak-off test pressure/volume plots are described and the theoretical interpretation of leak-off pressures is discussed in view of the information interpreted from the pressure/volume plots. In the final part of Chapter four, the procedure for calculating vertical stress from geophysical logs is outlined.

Chapter Five summarises all the various types of data and the sources of the data which have been used in this study. As the data used comes from a variety of sources, and is in several different formats, this Chapter is crucial as it provides a bridge between the theory outlined in Chapter four and the results and interpretation of Chapters six and seven.

Chapter Six presents the results and some interpretation of the data outlined in Chapter five. There are two stages in the presentation and interpretation of the results:

(i) The first stage (section 6.2) compares the leak-off test and hydro-frac data in order to analyse how (in view of the theory outlined in Chapter four) leak-off test data can be best used to determine $\sigma_h$. It is concluded that where just individual leak-off pressures are available with no additional information, the single leak-off pressure can not be used as a reliable estimate of $\sigma_h$. However, where the leak-off test pressure/volume plots are
available, in many cases, an estimate of $\sigma_h$ can be made from an individual test. It is also concluded that the trends of leak-off pressure with depth (where pressure/volume plots are not available but there is data from many tests) do reflect the trends of $\sigma_h$ with depth and specifically the lower bound to such a dataset will be close to the actual trend of $\sigma_h$ with depth.

(ii) The second stage of Chapter six (sections 6.3 and 6.4) and much of Chapter seven are largely based on the conclusions of section 6.2. Thus, section 6.3 presents the trends of leak-off test pressures with depth in the North Sea, where it is assumed that the individual data points do not necessarily represent good estimates of $\sigma_h$ but that their trend with depth does reflect the trend with depth of $\sigma_h$. Section 6.4, on the other hand, presents actual values of $\sigma_h$ which are the results of leak-off test pressure/volume plot interpretation from the subset of data for which such plots are available. The calculated vertical stresses are also presented in this section.

Chapter Seven consists largely of interpretation of the leak-off test data. Section 7.2 compares various trends of $\sigma_h$ with depth derived from the leak-off test data from the North Sea. The most appropriate trend is chosen for comparison with data from the rest of the world later in this Chapter.

Using the values of $\sigma_h$ and the trends of $\sigma_h$ with depth which have been established above and using the vertical stress calculations at the points of the leak-off tests, the nature of the distribution of stress in the North Sea is discussed (section 7.3) and the possible origins of the stress are investigated in terms of the stress models which have been outlined in section 3.4.

Section 7.4 discusses some results of a simple method to place a lower bound on the magnitude of $\sigma_H$ using breakout and leak-off data, and thus to determine the stress regime.
In Section 7.5 the trend of $\sigma_b$ with depth in the North Sea is compared to other trends of $\sigma_b$ with depth from around the world. In particular, the trend from the North Sea is compared to the trend at the KTB drillsite.

Chapter Eight presents the conclusions of this study. Some suggestions for future work are also discussed, and some recommendations concerning leak-off test procedure and data management are made.
Chapter Two - The North Sea

2.1 Introduction.
The North Sea sits on continental crust which forms part of the North West European continental shelf. The North Sea Mesozoic rift system, which includes the Viking, Central, Moray Firth-Witch Ground Graben and the Horda-Egersund half graben, is continuous for over 1000 km and forms an integral part of the Arctic-North Atlantic rift system (Ziegler, 1990).

Exploration in the North Sea was really initiated on 20 May 1964 when the German Consortium spudded their first offshore well Nordsee B-1. The area has now developed into one of the most prolific hydrocarbon provinces in the world (Brennand et al., 1990). The acquisition of thousands of miles of seismic data together with the geological information obtained from the large number of exploration wells that have been drilled means that the structure and stratigraphy of much of the region is well known. In the central and northern North Sea the data tends to be concentrated along the axes of the Mesozoic rifts, where most of the hydrocarbons are found. In the southern North Sea, the data is mainly distributed throughout the Southern Permian Gas Basin.

The contours to the depth of the top of the Chalk beneath the North Sea (Figure 2.1) indicate a simple pattern of subsidence centered along a north-south line roughly down the middle of the area. In contrast to this relative simplicity, the construction of a fault map of an older Mesozoic or Late Paleozoic seismic reflector at depth, reveals that the structure of some parts of the North Sea is far from simple (Glennie, 1990). Some of the structural complexities of the pre-chalk rocks are illustrated in Figures 2.2 and 2.3.

The North Sea region has a long and complex geological history. Most of the rifting which is generally thought of as characterising the tectonic style of the North Sea took place in the Mesozoic. However, the tectonic history of the pre-Mesozoic rocks stretches back long before this, and crustal features inherited from earlier events may well have a strong influence on the style of more recent deformation (Ziegler, 1990, Bartholomew et al., 1993). This chapter aims to summarise the most important
geological events in the development of the North Sea and is based largely on the work of Cowgill (1994).

Figure 2.1 Top Chalk structural contours. (After Lovell, 1990).
Figure 2.2 The main prospective basins of the North Sea. (After Brennand et al., 1990)
2.2 Major Structural Features of the North Sea Area

The North Sea contains the site of an axis of considerable Tertiary subsidence. This area of subsidence is flanked by the positive areas of the British Isles to the west and Scandinavia (including the Danish peninsula) to the east. The southern margin is marked by the terrestrial limit of the alpine foreland (The Netherlands, Germany). The northern limit of the North Sea follows the edge of the north-northeast trending Atlantic continental margin just beyond the Shetland Isles.

Beneath the Mesozoic deposits the area is dominated by two east-west trending basins, the larger Southern and smaller Northern Permian Basin which are separated by the Mid North Sea-Ringkøbing-Fyn system of highs and are surrounded by positive areas of older deformed rocks (Glennie, 1990).

Beneath the Southern Permian Basin lies an older basin which contains numerous carboniferous coal seams. This coal is important to the hydrocarbon industry as it is the source of most of the gas that is extracted from the southern North Sea (Glennie, 1990).

Overprinted on these Carboniferous and Permian Basins and highs is the dominant structural feature of the North Sea Basin, the zig-zag north-south trending Viking and Central Graben system. The Viking Graben lies to the north of the north Permian Basin and separates the Shetland Platform to the west and the Fennoscandian High to the east. The Central Graben cuts both the Permian Basins and the intervening high.

The other prominent graben of the North Sea is the Moray Firth-Witch Ground Graben which runs roughly east-west and intersects the southern end of the Viking Graben and the northern end of the Central Graben so that the area common to all three of these graben is often referred to as a triple point. These main structural features are summarised in Figure 2.3.

Some of the other major structural features in and around the North Sea area are shown in both Figure 2.2 and Figure 2.3. The Midland Valley Graben crosses Southern Scotland obliquely and extends beneath the North Sea where it dies out before the central Graben is reached. The Southern Uplands forms an area of positive relief and is
made of strongly folded lower Paleozoic strata which, as a structural unit, seems to be continuous with the north-western part of the Mid North Sea High.
Figure 2.4 Cartoons of relative plate movements that affected the North Sea Area during the late Proterozoic and the Phanerozoic. The scales are not constant. (After Glennie, 1990).

2.3 The Tectono-Stratigraphic History of the North Sea Basin.
An outline of the most important events in the geological evolution of the North Sea area has been given very briefly in a series of cartoons in Figure 2.4.

- The combined Late Cambrian to Late Silurian Athollian and Caledonian Orogenies (Figure 2.4C). Prior to these events, the North Sea area comprised widely separated continental fragments in, and marginal to, different parts of the Early Paleozoic Iapetus Ocean and Tornquist Sea (Figure 2.4A, 2.4B; Glennie, 1990).

- Rifting during the Devonian and Carboniferous possibly as a result of adjustments between and along the margins of the formerly separate Laurentian and Baltic cratons. This is exemplified by Devonian movement along the Great Glen Fault (Figure 2.4D) and by Early Carboniferous structural relief (Glennie, 1990).

- The Late Carboniferous Variscan Orogeny marked the closure of the southern Proto-Tethys Ocean and the creation of the supercontinent Pangaea (Figure 2.4E; Glennie, 1990).

- During the Early Permian, subsidence of the Moray Firth and the E-W trending Northern and Southern Permian Basins was followed by the initiation of the N-S trending Viking and Central Graben system (Figure 2.4F). The grabens began to subside rapidly during the Triassic and reached their maximum structural development by the beginning of the Cretaceous due to Mid-Jurassic domal uplift and widespread erosion centred over the axis of the Central Graben.

- Late Jurassic to earliest Cretaceous strike-slip movements and fault-block rotations within and adjacent to the Viking and Central Grabens is thought to have coincided with the opening of the Central Atlantic and rifting between Iberia and Newfoundland (Glennie, 1990; Figure 2.4G). Thus, the duration of an active Viking-Central Graben system coincided with the slow break-up of Pangaea into North America, Europe and Africa.
• The end of the North Sea Graben development was linked to the onset of sea-floor spreading in the North Atlantic Ocean (Figure 2.4H).

2.3.1 The Pre-Devonian (400Ma and earlier)

The pre-Devonian history of the North Sea area is a complex story of movement of different crustal blocks involving oceanic opening and closing, terraining of crustal blocks over large distances, transpression and transtension. As mentioned above (section 2.3) the final coming together of these crustal blocks occurred in the Late Cambrian to Late Silurian combined Athollian and Caledonian Orogeny. This orogenic episode was associated with the closure of the Iapetus Ocean and its eastern arm, the Tornquist sea, to form the supercontinent Laurasia (Glennie, 1990).

Several authors have documented a number of basement lineaments predating the dominant Mesozoic basin formation of the North Sea (Bartholomew et al., 1993). Three structural trends have been documented in the crystalline Caledonian basement rocks of the North Sea by Ziegler (1990):

- The Scottish-Norwegian Caledonides, which cross the northern North Sea and are characterised by a NE-SW striking structural grain.
- The North German-Polish Caledonides (the approximate site of the Tornquist Sea; see Figure 2.4B) which branch off from the Scottish-Norwegian Caledonides in the central North Sea are characterised by a N-S striking structural grain.
- In the south-central and southern North Sea, the North German-Polish Caledonides exhibit an ESE-WNW structural grain.

2.3.2 The Devonian (400Ma - 360Ma)

The closure of the Iapetus Ocean resulted in the Creation of the supercontinent Laurasia and in the uplift of the Appalachian mountain range which stretched from the USA through eastern Canada, Northern Britain and the northern end of the then united Greenland and Scandinavia (Glennie, 1990). Erosion of this mountain belt and deposition in a continental environment of the erosional products produced the Devonian Old Red Sandstone.
In the northern North Sea area basins such as the Orcadian Basin and the Midland Valley Basin (Figure 2.5) were formed during the Early Devonian. These basins are interpreted as wrench induced and as having subsided rapidly when being filled with thick sequences of Old Red Sandstone (Richards 1990). The formation of these basins is associated with the post orogenic crustal re-organizations which took place within the newly formed Laurasian continent, Figure 2.6 shows the relationship between these basins and the major faults in Mid Devonian times. Such post orogenic crustal movements are thought to have occurred on the Great Glen Fault. Estimates of the direction and amount of movement on the Great Glen Fault vary enormously (Glennie, 1990). Deposition of Old Red Sandstone did however occur on both sides of the Great Glen Fault in the Orcadian Basin. Crustal deformation continued throughout the Devonian as is evidenced by the Late Devonian deformation of Mid Devonian sandstones in northern part of the Orcadian Basin. After folding, these sandstones were intruded by dykes and sill like sheets suggesting deep crustal tension in the Late Devonian followed the compressive folding event. Ziegler (1982) points out that the deposition and deformation of the Old Red Sandstone in the Orcadian Basin was probably the result of an interplay between tensional and compressional stresses associated with horizontal displacements on faults such as the Great Glen Fault. Across the Walls Boundary Fault on Shetland, which is probably the northern extension of the Great Glen Fault, there is a marked lithological contrast in the Devonian sediments and there is also a marked offset in the depth of the Moho (McGeary, 1989, Klemperer and Hurich, 1990). This is good evidence that major crustal structures which obviously exerted a lot of influence on the sedimentary and tectonic evolution of the area as long ago as the Devonian and earlier are still mechanically significant. The rest of the Great Glen Fault on shore Scotland is well known for being seismically active (Ambraseys and Jackson, 1985). Although recent seismic events are small, the activity on this fault is still significantly higher than surrounding areas.
A major Old Red Sandstone Basin seems to have occupied much of the North Sea area by Mid Devonian times (Figures 2.5 and 2.6). Glennie (1990) points out that this may reflect crustal weakness in the North Sea area inherited from the northern Iapetus Suture between Laurentia and Fennoscandia. He goes on to suggest that it is perhaps significant that the Devonian Basin bears some similarity to the Cenozoic North Sea Basin. Figures 2.5 and 2.6 also show a narrow marine invasion coming from the proto-Tethys to the south into the Old Red Sandstone continent. This marine embayment follows almost exactly the same line as the Mesozoic Central Graben. This may be another example of the inheritance of an old line of weakness having a controlling effect on the structural evolution of the crust.

By the Late Devonian, progressive overstepping of basin margins, related to rising sea levels, caused the individual basins to become connected with each other (Figure 2.5C). The upper Old Red Sandstone spans the boundary between the Devonian and Carboniferous and the actual position of the chronostratigraphic change is often difficult to determine.
2.3.3 The Carboniferous (360Ma to 290Ma)

The Carboniferous of North-West Europe is very important to the hydrocarbon industry because it was during this period that the carbonaceous source rocks for the whole of the southern North Sea and Dutch-German-Polish gas belt were deposited. The main period of coal deposition was during the Late Carboniferous Westphalian stage, where up to 2500m of strata were deposited locally, with a cumulative thickness of some 75-100m of coal (Glennie, 1990). With the slow northerly drift of Laurasia, Early Carboniferous sedimentation represents a transition from the relatively arid conditions of the southern hemisphere tropics that prevailed at the end of the Devonian to the more humid equatorial conditions of coal measure deposition (Glennie, 1990).

The Carboniferous rocks of the British Isles and North West Europe are divided into two sub-systems, the Dinantian (360Ma to 326Ma) and the Silesian (326Ma to 290Ma) (Harland et al., 1989).
Throughout most of the Carboniferous, the North Sea area lay to the north of the Variscan orogenic belt. This orogeny was the result of Gondwana migrating northwards at a faster rate than Laurasia (Figure 2.3D) causing the proto-Tethys Ocean to be subducted beneath Laurasia. To the north of the North Sea area the Caledonian Highlands remained an area of positive relief and a source for sediments throughout the Carboniferous.

The North Sea and adjoining areas were largely peripheral to the Variscan Orogen until its last phases. However, they have a complex tectonic history which was probably largely controlled by events within the Orogen to the south (Besly, 1990). The Carboniferous Basin of Northern England and its eastward extension into the North Sea was separated from the northward migrating foreland flexural basin of the Variscan Mountains to the south by the Wales Brabant High. This area of positive relief may have been of Caledonian origin but its structural importance increased during the Carboniferous (Besly, 1990).

2.3.3.1 Lower Carboniferous.
To the north of the Wales-Brabant High, the Early Carboniferous saw the break-up of the stable Old Red Sandstone Continent due to widespread crustal extension. A zone of rapidly subsiding graben and half graben was established in Northern England. This zone is inferred to have extended eastward into the North Sea area. Sequences of condensed sediments are seen to have been deposited on the horsts (Figure 2.7). The location of the zones of lower subsidence rate appear to have been controlled by the location of basement heterogeneities, particularly the presence of Devonian and Precambrian granite plutons and by the tectonic grain inherited from earlier orogenic events (Besly, 1990). Much of the sedimentation during the Early Carboniferous was carbonate facies deposited as a result of a major northerly transgression of the equatorial proto-Tethys Ocean which covered most of the British Isles and North Sea area to the east of the British Isles (Anderton et al., 1985). The northern North Sea area is thought to have remained an area of positive relief until the Mid-Permian (Doré and Gage, 1987).
Figure 2.7 Structural Framework of Lower Carboniferous graben and half graben development. "highs" are mainly located over Devonian granite masses. D = Derbyshire Block; DS = offshore Durham Shelf; G = Gainsborough Trough; H = Humber Basin; IS = Inde Shelf; ND = North Dogger Shelf*; S = Silver Pit Basin*; SH = South Hewett Shelf*; W = Widmerpool Trough. Structural elements marked * are of Mesozoic age but may have been active in the Carboniferous. (After Besly, 1990)

The orientation of the prevailing stress field during the Lower Carboniferous remains a subject for debate, Leeder (1988) argued for a N-S oriented stress field, driven by back arc extension to the north of the Variscan subduction system. It has also been suggested that extension may have occurred in an east-west direction, in response to an early phase of opening in the future North Atlantic (Haszeldine, 1984).

The end of the graben development in Mid-Carboniferous maybe related to the impingement of the Variscan orogen from the south, creating north-south compression (Glennie, 1990).
2.3.3.2 Upper Carboniferous

During the Mid-Carboniferous there was a rejuvenation of clastic sources such as the Scottish highlands and other sources to the west, north and north-east. The marine influence of the proto-Tethys was driven back by coastal plain progradation (Anderton et al., 1985; Besly, 1990; Glennie 1990). Shales, siltstones, sandstones and coals of the Millstone Grit series were deposited as a series of overlapping, turbidite fronted, fluvial dominated deltas which prograded into the UK from the north and east (Kirby et al., 1987). Figure 2.8 shows the Mid-Late Carboniferous palaeogeography of the UK and North Sea area (Doré and Gage, 1987). Note that the northern North Sea area remains an upland area and the Mid-North Sea High area is a basinal area, i.e. the Mid North Sea High had not yet formed.

The Westphalian is the major coal bearing stage in the British Isles and NW Europe. It occupies most of the latter half of the Upper Carboniferous. Just before the onset of the Westphalian, sedimentation rates exceeded subsidence rates and deltaic deposits rapidly infilled the basinal topography (Kirby et al., 1987). The London-Brabant platform remained a positive feature during the Westphalian although southward pro-grading sediments may have crossed it as coals are seen to have been deposited on the south flank of the platform in Kent, where the Westphalian coal measures onlap Silurian rocks (Figure 2.9).
During the Late Carboniferous, the central and southern North Sea was occupied by distal parts of the Variscan foreland basin. In this basin, a southward expanding thick wedge of paralic coal measures accumulated (Ziegler, 1990). Siltstones with interbedded coals, laterally discontinuous sandstones and carbon-rich mudstones in marine bands were deposited.

During the upper stage of the Carboniferous, the Stephanian, the first evidence of a system of NW-SE and conjugate NE-SW trending wrench faults is seen (Oudmayer and Jager, 1993). These faults developed following a proposed change in relative movement between Laurasia and Gondwana, are thought to have aided the collapse of the Variscan fold belt (Glennie, 1990). In the southern North Sea, many of these wrench faults are
aligned parallel to Precambrian lineaments and may be related to a rejuvenation of these structures (Glennie, 1990).

Figure 2.9 NW European Carboniferous Basin. (After Glennie, 1990)

The Mid North Sea-Rinkøbing-Fyn trend of highs also developed at the end of the Carboniferous (Ziegler, 1990). This structural high may have been created by the Variscan compression or by the intrusion of granitic plutons along an E-W line (Doré and Gage, 1987). Either of these two mechanisms of creating a high feature may have exploited a pre-existing weakness. The trend of these systems of highs is continuous offshore from the Southern Uplands, the southern boundary of which is delineated by the Iapetus Suture (Glennie, 1990). The Iapetus Suture has been identified on deep seismic reflection profiles and seen to extend into the North Sea along the line of the Mid North Sea High at least as far as the Central Graben (Klemperer and Hurich, 1990).
2.3.4 The Permian (290Ma - 245Ma)

At around the Carboniferous-Permian time boundary, two different processes followed each other in the North Sea region in fairly rapid succession. The first was the development of the Variscan Mountains which involved north-south compression and probably ceased in the Late Westphalian. The next process began very early in the Permian and resulted in NW-SE oriented transtension (Glennie, 1990; Anderton et al., 1985). This Early Permian transtension was associated with volcanism and also aided the collapse of the Variscan Mountains. This same transtensional stress field also led to the subsidence of the Southern and Northern Permian Basins within the northern foreland (Glennie, 1990). In Permian times the area between the Hercynian fold belt to the south and the Caledonian fold belts of Scotland and Scandinavia to the north was dominated by the two east-west trending, Northern and Southern Permian Basins. These two basins are separated by the Mid North Sea-Ringkøbing-Fyn system of highs (Figure 2.3).

Another system of east-west and south-west - north-east trending horsts and grabens such as the English Channel, Southwestern Approaches, Bristol Channel and the Celtic Sea Basin lie within the zone of strong Variscan folding within the south and probably have their origins within the Early Permian (Glennie, 1990).

Running north-south and cutting the basins mentioned above, are a series of other basins. The Central Graben and its Northward extension the Viking Graben plus the Horn Graben and its northern extension the Oslo Graben. The Central and Horn Graben both cut right through the Mid North Sea-Ringkøbing-Fyn system of highs (Figure 2.2). Glennie (1990) suggests that mainly towards the end of the Variscan Orogeny, the Cantabrian Mountains (part of the Variscan Mountains to the south of the North Sea area) as they were moving north, acted as a wedge and attempted to split the former Laurentian and Fennoscandian continents (Figure 2.4C - 2.4F) along lines of north-south weakness such as the possible northern end of the Iapetus Suture. It is further suggested that such a wedge like action would initiate grabens that would propagate from north to south and that this could be the origin of the north-south grabens named above.
The end of the Variscan Orogeny was accompanied by substantial uplift which also affected the northern foreland. This uplift is evidenced by a substantial unconformity between the Upper Carboniferous and the Lower Permian (Anderton et al., 1985).

2.3.4.1 Lower Permian

The Rotliegend rocks consist of both volcanics and sediments and, as their German name suggests, are largely red. The sediments comprise continental red beds, basinal shales and evaporites and reach a maximum thickness of 1500m in the southern basin and 600m in the northern basin (Glennie, 1990a).

The Rotliegend sands which are developed in a broad belt along the southern margin of the southern North Sea Basin form the reservoir rocks for a number of major gas accumulations in this region.

The Rotliegends are divided into upper and lower. The Lower Rotliegend is limited in distribution in comparison to the Upper Rotliegend. The Lower Rotliegends contain volcanic rocks which are distributed adjacent to known or inferred faults and are probably related to the extensional movements responsible for the formation of the basins.

The Upper Rotliegend is made of four distinctive facies interpreted as products of deposition in fluvial (wadi), aeolian, sabkha, and lacustrine environments. All four facies are widely distributed throughout the asymmetric basin, whose floor sloped from south to north over much of the area (Glennie, 1990a).

Figure 2.10 shows the distribution of the palaeogeography of the Upper Rotliegends. These rocks are rarely drilled in the area of the Northern Permian Basin and so the facies distributions are poorly known.

Figure 2.11 shows the stratigraphy of the Rotliegend sediments. As mentioned above (section 2.3.4) they are seen to lie unconformably over the Upper Carboniferous. Also shown in Figure 2.11 is the thin highly organic shale known as the Kupferschiefer (Taylor, 1990). This shale is present in both North and South Permian Basins and marks the top of the Rotliegend before the onset of the Zechstein flooding. It reflects the low
energy of the epicontinental basin with little or no circulation of bottom waters (Anderton et al., 1985).

Figure 2.10 Upper Rotliegend facies and palaeogeography, and limit of Zechstein transgression. Facies distribution poorly known in the Northern Permian Basin. (After Glennie, 1990a)

Figure 2.11 Rock Stratigraphic diagram of the Permian Upper and Lower Rotliegend Groups. Names in brackets refer to formations in the Northern Permian Basin. Dotted areas represent fluvial or mixed fluvial and aeolian sands. (After Glennie, 1990a).

Throughout the Lower Permian it has been suggested that the subsidence of the Rotliegend Basins greatly exceeded the rate of sedimentation with the subsequent development of basin-wide topographic depressions below the level of the world's
oceans (Ziegler, 1982). This sets the scene for the rapid marine flooding seen in the Upper Permian Zechstein.

2.3.4.2 Upper Permian

It was during the Upper Permian that the Zechstein Halite was deposited. These thick sequences of a material with quite different mechanical properties to other sediments have had a profound effect on the structural development of the North Sea Basin and certainly still do have a significant geomechanical impact.

The Upper Permian began with a major, very rapid, marine transgression which extended over parts of the Netherlands and Germany and led to the deposition of cyclical carbonate and evaporite sequences of the Zechstein Formation (Taylor, 1990). Fossil evidence indicates that the water was most likely derived from the open ocean somewhere between Norway and Greenland and that the rate of supply was controlled not by tectonics but by the waxing and waning of the Gondwana ice caps (Glennie, 1990). The glacio-eustatic control is evidenced by the cyclical nature of the Zechstein series. Evidence for the rapid nature of the initial transgression is provided by the presence of basal conglomerates and Rotliegend sand dunes which have been reworked and exhibit slumping compatible with sudden drowning.

In the North Sea, Zechstein deposits comprise five major sedimentary cycles, Z1 - Z5. Each cycle begins with shelf carbonates and grades up into evaporites (Taylor, 1990). The carbonates are best developed around the margins of the basin, and the thickest evaporites (up to 1000m) occur towards the centre. Salt diapirism has since complicated the thickness patterns considerably (Anderton et al., 1985). The cycles in both the North and South Permian Basins are correlative and indicate that there was communication across the Mid North Sea-Rinkøbing-Fyn High (Glennie, 1990).

The “ideal” Zechstein cycle illustrates the influence of increasing salinity and decreasing water depth through evaporation of a marine incursion, commencing with a thin clastic member which passes upwards through limestone/dolomite, anhydrite, halite and occasionally, to very soluble magnesium and potassium salts (Taylor, 1990). Although
glacio-eustatic sea level changes were probably the dominant cyclic control, local
tectonics in some parts of the basins, such as the propagation of the northern North Sea
Graben system is also likely to have played a role. Indeed, the various cases of locally
missing or reversed cycles may be explained by local tectonics.

2.3.5 Triassic (245Ma - 208Ma)
During the Triassic the North Sea area saw the extensional development of the rift
system which was to dominate the structural evolution and the deposition of sediments
throughout the rest of the Mesozoic. The break-up of Pangaea which began with crustal
thinning and rifting along the axis of the incipient Atlantic had already established a new
structural framework in North-West Europe (Fisher and Mudge, 1990). The subsidence
that began in the Permian rapidly accelerated with block faulting leading to a complex
set of multidirectional horsts and grabens (Anderton et al., 1985). The localization of
these Mesozoic grabens was, to a large extent governed by the tensional reactivation of
pre-existing fracture systems (Ziegler, 1982; Bartholomew et al., 1993). The Viking and
Central Grabens had been initiated in the Permian (see section 2.3.4) but during the
Triassic they became the major structure in the North Sea, stretching for over 1000 km
and transecting the Northern Permian Basin and the Mid North Sea High (Fisher and
Mudge, 1990).

The close of the Permian saw the end of widespread marine sedimentation and a return
to predominantly non-marine depositional environments. The final regressive stage of
the Zechstein Sea saw the migration of continental environments into the centres of the
basins. As a consequence, the Triassic series consists mainly of continental red beds
which include alluvial fan, aeolian, sabkha, and lacustrine facies (Fisher and Mudge,
1990).

Throughout the Early Triassic, sedimentation is thought to have kept pace with
subsidence (Ziegler, 1982) and it has been proposed that the northern and southern
basins continued to subside along the fault lines established in the Permian (Ziegler,
1982).
The rift pattern in the Triassic was dominantly north-south (Figure 2.12). The Central Graben, north of the Mid North Sea High, runs north-west - south-east but there is no clear evidence that rifting took place in this part of the graben during the Triassic (Fisher and Mudge, 1990; Bartholomew et al., 1993). The north-south part of the Central Graben, the Viking Graben, the Horn Graben and others all show evidence of active rifting during the Triassic (Fisher and Mudge, 1990). The influence of rifting on sedimentation patterns seems more pronounced north of the Mid North Sea High, where sedimentation is much less laterally continuous than in the Southern North Sea Basin.

2.3.5.1 The Southern North Sea Basin

The Triassic Southern North Sea Basin occupied roughly the same position as that of the Southern Permian Basin (Figure 2.12). Areas of positive relief such as the London-Brabant Massif and the Mid North Sea-Ringkøbing-Fyn High were still in existence. The new tectonic features in this area are the Central and Horn Grabens which dissected the Mid North Sea-Ringkøbing-Fyn High and were rapidly infilled throughout the Triassic with up to 2000m of sediments accumulating in the southern part of the Central Graben and up to 4000m in the Horn Graben (Fisher and Mudge, 1990). Within the basin itself, the Sole Pit, Broad Fourteens and West Netherlands sub-Basins were major Triassic tectonically controlled depocentres (Figure 2.2) (Fisher and Mudge, 1990).

The Triassic sediments of the southern North Sea falls naturally into two major groups the Bacton Group and the Haisborough Group (Figure 2.13). The Bacton Group represents a phase of clastic sedimentation of sandstones, shales and mudstones. The Haisborough Group is a largely fine grained clastic and evaporite sequence with a marked cyclic character.
Figure 2.12 Triassic structural elements. (After Fisher and Mudge, 1990).
The Permian-Triassic transition is represented by a distinct facies change with the lower formation of the Bacton Group, the Bunter Shale, overlying the basinal carbonates, anhydrites and halites of the Zechstein (Fisher and Mudge, 1990). The sandstones of the Bacton Group such as the Bunter Sandstone Formation, result from the periodic rejuvenation of a major southern source area such as the London-Brabant Massif (Anderton et al., 1985).

Greater tectonic stability is evident during the deposition of the Haisborough Group (Fisher and Mudge, 1990) with the Mid North Sea-Ringkøbing-Fyn High still forming the northern limit of the basin (Anderton et al., 1985). Local tectonic activity combined with eustatic sea level changes have been proposed as the cause of periodic marine transgressions from the Tethyan marine realm which deposited Muschelkalk shelly limestone series, the Keuper Marls and Halites some of which were cyclical in nature. At the top of the Haisborough Group is the Winterton Formation which onlaps onto structural highs and reflects the marine transgression that marked the passage from the Triassic to the Jurassic (Fisher and Mudge, 1990).

2.3.5.2 The Central North Sea Basin

The Central North Sea Basin lies to the North of the Mid North Sea-Ringkøbing-Fyn High and is located in approximately the same site as the Northern Permian Basin (Figure 2.12). The Central Graben, which may have been initiated as an area of active rifting in the Early Permian (Glennie, 1990) remained an area of deposition. However, positive evidence of major Triassic rifting is restricted to the parts of the central and Horn Grabens which breach the Mid North Sea-Ringkøbing-Fyn High, the Moray Firth Basin and some faults bordering the Norwegian-Danish Basin. Areas of Triassic sedimentation which occur over the Permian evaporites are thought to have been structurally controlled to a large extent by halokinesis (Fisher and Mudge, 1990).

The central North Sea Triassic succession can be divided into the Smith Bank Formation and the Skagerrak Formation (Figure 2.13)
The Smith Bank Formation comprises basinal silty mudstones with rare sandstone stringers and marginal conglomerate beds. The mudstones within the Smith Bank Formation are comparable with the Bunter Shale Formation of the southern North Sea (Fisher and Mudge, 1990).

The Skagerrak Formation is, in part, younger than, and in part of equivalent age to the Smith Bank Formation. It comprises interbedded conglomerates, sandstones, siltstones, and shales which are thought to have accumulated at the margins of the basin and are proposed to have been derived from the Fennoscandian Shield (Fisher and Mudge, 1990; Ziegler, 1990).

2.3.5.3 The Northern North Sea Basin

The structure of the northern North Sea Basin is dominated by north-south faulting which has formed deep and well defined grabens. Faulting is thought to have commenced during the Permian and with rapid subsidence occurring throughout the Triassic (Ziegler, 1982; Fisher and Mudge, 1990). The variation in thickness of the Triassic sediments in the graben indicate differential subsidence and the facies
relationships are compatible with fault controlled subsidence (Fisher and Mudge, 1990).

The Northern North Sea Basin was apparently separated from the Central North Sea Basin by a structural element in the vicinity of the triple junction as evidenced by sediment supply from the south in the Viking Graben and from the north in the Central Graben during the Rhaetian (Fisher and Mudge, 1990).

The Triassic sequences in this area has been divided into the Cormorant Formation south of the latitude 60°N, and the Hegre Group to the north. Both groups are overlain by the Statfjord Formation (Fisher and Mudge, 1990).

The Cormorant Formation is mainly composed of argillaceous sandstones with some siltstones, shales, conglomerates, and coarser-grained sandstones deposited in a fluvio-lacustrine environment (Fisher and Mudge, 1990). Correlation of this sequence between separate fault blocks has proven to be very difficult due to the unfossiliferous nature of the rocks and the use of this term has been proposed to be restricted to attenuated sequences on structural highs (Fisher and Mudge, 1990).

The Hegre Group, comprising the Teist, Lomvi, and Lunde Formations, is used as a broad term to cover the Triassic sediments in this area. The rocks comprise interbedded sandstones, shales and claystones resulting from deposition in fluvial, aeolian and lacustrine environments (Fisher and Mudge, 1990). Detailed correlation between the Hegre Group and the Cormorant Formation is not feasible from the current published data (Fisher and Mudge, 1990).

The Statfjord Formation overlies both the Cormorant Formation and the Hegre Group. Its base is defined by a coarsening-upward succession of grey, green, and red shales interbedded with thin siltstones, sandstones and dolomites indicative of a slow marine transgression (Fisher and Mudge, 1990).

2.3.5.4 The End of the Triassic
At the end of the Triassic, peneplanation was almost complete. The basins contained thick sediments that onlapped the eroded highs. The Early Jurassic transgression invaded the vast continental floodplains and tidal flats of the North Sea Basins and re-established marine conditions in North West Europe.

2.3.6 The Jurassic (208Ma - 145Ma)

From the point of view of the hydrocarbon industry the Jurassic is the most important system in the sedimentary basins of the North Sea. The Jurassic is economically important for both reservoir rocks and perhaps more importantly, the source rock for most of the North Sea oil, the organic rich Kimmeridge Clay. As a result of its importance the Jurassic is very commonly drilled, logged and imaged seismically and so its structure and stratigraphy are generally well known.

(During the Jurassic the North Sea was flanked by large active rift systems within the Arctic North Atlantic and Tethyan Oceans). Jurassic strata in the North Sea area occur for the most part in fault basins related to the development, through regional crustal extension, of the graben system which was initiated in the Permian and further developed during the Triassic (see sections 2.3.4 and 2.3.5). The Jurassic was a period of active faulting with fault controlled differential subsidence and syn-tectonic sedimentation having a marked influence on stratigraphic thickness and facies distribution, especially in the Late Jurassic (Brown, 1990).

The phase of rifting which commenced in Mid-Late Jurassic (Oxfordian) and ceased at the end of the Jurassic (in the Ryazanian) represents the dominant tectonic pulse in the evolution of the North Sea Mesozoic rift system (Bartholomew et al., 1993). Of the faults active in the Jurassic, many have been assigned to older structural trends and evidence of reactivated Precambrian, Caledonian and Variscan lineaments has been reported from different parts of the North Sea (Brown, 1990). The discordance between the NE-SW Caledonian orogenic belt and the trend of the Viking and Central Grabens has been noted (Watson, 1985). However, as mentioned in section 2.3.3.2 there are NW-SE trending basement lineaments present in the southern and central North Sea which are related to the latter stages of the Variscan Orogeny and may be
reactivated even older structures (Glennie, 1990). Also, as noted in section 2.3.4, the late stages of the Variscan Orogeny are thought to have initiated north-south trending rifts when the Variscan Mountains drove wedge-like into the foreland and reactivated what could be the northern extension of the Iapetus Suture (Glennie, 1990).

The Jurassic development of the main North Sea rift system, that is the Viking, Central and Moray Firth/Witch Ground Grabens (Figure 2.14), was initially thought to have been due to plume generated domal uplift centred at the triple point of the three grabens where there is a Mid Jurassic volcanic pile (Brown, 1990). However as pointed out by Bartholomew et al. (1990), this requires concurrent unique extension directions for each of the grabens. This type of basin formation is akin to the so-called active rift formation, where the word "active" refers to the active upwelling of a mantle plume. The other mode of basin formation is the so-called passive rifting, where the word passive refers to the passive response of the mantle as it upwells beneath stretched lithosphere. This type of interpretation emphasises the presence of anomalously thin continental crust under much of the main grabens (Christie and Sclater, 1980) which relies on a hypothesis of regional lithospheric stretching (after McKenzie, 1978). Wood and Barton (1983) associated the extrusion of basalts during the mid-Jurassic with the most significant and rapid phase of extension. This model envisaged stretching and fault-controlled subsidence throughout the mid to Late Jurassic and into the Early Cretaceous, when, as extension in the North Sea area ceased, thermally induced subsidence became the dominant influence on basin formation (Brown, 1990).
Figure 2.14 Structural elements of the North Sea basin. The principal Jurassic play fairways are located in the Central Graben, Witch Ground Graben and in the Viking Graben, including its marginal terrace, the east Shetland Basin. (After Brown, 1990).
Bartholomew et al. (1990) suggest that a simple stress field with maximum horizontal stress oriented approximately north-south is responsible for regional extension oriented roughly east-west across the whole of the North Sea area with a dominant
structural style of extensional faulting and block rotation as is seen in much of the area (e.g. the Viking Graben). However they go on to point out that basin architecture in some areas, notably the central North Sea, is considerably more complex than would be expected if the region had merely been accommodating crustal extension by simple dip-slip deformation. These complexities are explained in terms of the pre-existing basement structure inherited from Palaeozoic deformation which exert a geometrical control on basin development and lead to oblique movement on faults with associated transpression and transtension.

Another major influence on Jurassic sedimentation as well as the extensive tectonism, was eustatic sea level change (Brown, 1990). The Jurassic period is thought to have been characterised by a gradual rise in sea level until the Kimmeridgian (Brown, 1990). Marine transgression transformed much of Jurassic Europe into an enormous, generally shallow, epicontinental sea (Anderton et al., 1985). The interplay between subsidence, sedimentation and eustatic sea-level change are difficult to disentangle.

The main basins and structural elements of the Jurassic are shown in Figure 2.14. The generalised lithostratigraphy of the Jurassic within the North Sea Basin is illustrated in Figure 2.15.

2.3.6.1 The Lower Jurassic

The Lower Jurassic, consisting of broadly transgressive marine argillaceous deposits has the most restricted distribution of the three Jurassic series in the North Sea Basins (Brown, 1990). Lower Jurassic strata are absent, or at best sparsely distributed over much of the northern North Sea, however, such strata attain maximum thicknesses of 500m in the north Viking Graben, 750m in the Danish Basin and 250-500m in the Central Graben (Ziegler, 1990). Away from the uplifted areas, the Lower Jurassic succession reaches around 900m in the Sole Pit Basin and includes the organic-rich Posidonia Shale (Brown, 1990).

Within the Inner Moray Firth, the Lower Jurassic is represented by a transgressive sequence, consisting of shales with minor sandstones which is overlain by a
coarsening-upward sandy sequence formed in a broadly deltaic setting capped by the Middle Jurassic Brora Coal Formation (Brown, 1990). As illustrated in Figure 2.15 these facies do not continue into the Outer Moray Firth/Witch Ground Graben.

Lower Jurassic marine shales above a basal sand are widely distributed throughout the Viking Graben (Brown, 1990). These shales overlie the upper Triassic Statfjord Formation and show evidence of a diachronous marine transgression (Anderton et al., 1985; Brown, 1990).

2.3.6.2 The Middle Jurassic

During the Middle Jurassic, crustal extension accelerated in the Arctic-North Atlantic and Tethys-Central Atlantic Rift. This was accompanied by Late Aalenian-Bajocian upwarping of a large rift dome centred over the Central Graben and by the development of a large volcanic complex at the triple junction between the Viking Graben, Central Graben and Moray Firth-Witch Ground Graben system (Ziegler, 1982; 1990). This volcanic complex displayed a bimodal mafic-felsic alkaline chemistry typical of intracontinental rifts (Fall et al., 1982). The volcanic rocks appear to thicken southwards and reach a maximum thickness of 1500m in presumed vent areas (Fall et al., 1982). The bulk of the volcanic rocks were extruded by the end of the Bajocian although radiometric ages support both extrusive and intrusive activity into the Early Cretaceous (Fall et al., 1982).

The upwarping and development of the volcanic dome which has been proposed to have had a structural relief of 2-3 km, created a barrier between the Boreal and Tethys oceanic provinces (Ziegler, 1982). More importantly the dome provided a major clastic source for a series of radially draining fluvio-deltaic systems which followed major structural alignments. One such system is the Brent Group deposited within the proto-Viking Graben. This sequence of rocks is probably the single most productive reservoir unit in the North Sea (Brown, 1990). It comprises 5 formations: the Broom, Rannoch, Etive, Ness and Tarbert Formations which are composed of fluvial sands, prograding shoreface-foreshore sandstone and marine mudstones, coastal plain deposits and a transgressive sandstone unit (Brown, 1990).
Within the South Viking Graben, the continually rising sea-level resulted in the deposition of thick shallow-marine and, eventually, deep-marine sediments (Brown, 1990).

In the southern North Sea, sedimentation was continuous throughout the Lower and Middle Jurassic, shallow-marine lagoonal sandstones and shales, and occasional calcareous sediments conformably overlie the Posidonia Shales (Anderton et al., 1985; Brown, 1990). Distribution of Jurassic strata is delimited in much the same way as in the Triassic with the Mid North Sea-Ringkøbing-Fyn High to the north, the Pennine High to the west and the London-Brabant massif to the west (Figure 2.14). The areas of deposition can be subdivided into the Anglo-Dutch Basin, consisting of a number of smaller, narrow NW-SE trending basins and ridges, and the southern extension of the Central Graben (Figure 2.14). Mobilisation of the Permian Salt also played an important role in controlling sediment distribution (Brown, 1990).

In the Witch Ground Graben, Middle Jurassic strata rest unconformably on Triassic continental sediments (Figure 2.15) and consist of basaltic lavas and tuffs of the Rattray Formation which are thought to be the products of subaerial eruptions (Fall et al., 1982). Within the Forties and Piper oilfields, a 740m-thick sequence of undersaturated alkali basalts is interbedded with Bajocian-Bathonian volcaniclastic sediments, tuffs, and coal seams of the Pentland Formation which have been interpreted as being fluvio-deltaic sediments coeval with the Brent Group in the north Viking Graben (Fall et al., 1982; Anderton et al., 1985).

In addition to the mid-Jurassic volcanics to the north of the Central Graben, the fluvio-deltaic Bryne Formation rests unconformably on either the Late Triassic Winterton Formation or directly on continental Triassic deposits (Figure 2.15).

2.3.6.3 The Upper Jurassic
Late Jurassic seas reached their maximum extent during the upper Oxfordian and resulted in a diminution in the availability of coarse grained clastic material. The high
eustatic sea level coupled with rapid subsidence led to the extensive development of restricted basinal marine environments and the accumulation of laminated bituminous Kimmeridge Shales and stagnant or near stagnant marine bottom conditions (Anderton et al., 1985). The Kimmeridge Clay Formation forms the principal source rocks within the northern North Sea area (Brown, 1990).

Tectonic activity in the Upper Jurassic was concentrated around the Viking, Central and Moray Firth-Witch Ground Graben systems, normal faulting in these areas is proposed to have been accompanied by local dextral wrench deformations (Ziegler, 1990). By mid-Kimmeridgian time marine connections had been established between the central and southern North Sea across the Mid-North Sea High (Ziegler, 1988, 1990).

Along the margins of the Central Graben, sand bodies accumulated contemporaneously with the shales adjacent to a number intra basinal 'highs'. These form important reservoir units such as the Fulmar and Ula Formations and are thought to result from a complex interplay of local tectonic activity and halokinesis (Brown, 1990).

In the Moray Firth area, Upper Jurassic strata consist of shales and minor beds of sandstone with pronounced spatial thickness variations which are overlain by shales equivalent to the Kimmeridge Clay Formation (Brown, 1990).

The Upper Jurassic sequence of the Outer Moray Firth and Witch Ground Graben comprises three widely distributed units: the deltaic to shallow-marine heterolithic deposits of the Sgiath Formation, the shallow-marine sandstones of the Piper Formation and the overlying Kimmeridge Clay Formation (Figure 2.14).

The Upper Jurassic sequence within the Viking Graben consists of marine shales and turbiditic sandstones of the Heather Formation. These sediments are overlain by submarine-fan sandstones and conglomerates of the Brae Formation (Figure 2.15).
which is in turn blanketed by the Kimmeridge Clay Formation and its Norwegian equivalent, the Draupne Formation (Brown, 1990).

Along the western graben-margin fault, in the South Viking Graben, coarse, proximal submarine fan sediments originating from the Fladen Ground Spur were deposited whilst distal sands were deposited in the East Shetland Basin (Brown, 1990). The coarse clastic sediments are overlain by a thin veneer of Kimmeridge Clay which passes laterally into intercalations of sandy siltstones and shales (Brown, 1990).

Within the literature, the identification of a 'base Cretaceous' unconformity is commonplace (Brown, 1990). It has been proposed that a major 'Late Cimmerian' rifting pulse during the earliest Cretaceous produced a regional unconformity however, on closer inspection, an unconformity between Jurassic and Cretaceous strata is documented from marginal and intra-basinal highs (Brown, 1990), but for the most part, the Jurassic-Cretaceous transition is a conformable one.

2.3.7 The Cretaceous (145Ma to 65Ma)
During the Cretaceous, the overall style of tectono-stratigraphic evolution in the North Sea changed for good. The Cretaceous saw the end of rift dominated basin formation and the beginning of a broad general subsidence (Christie and Sclater, 1980). In the Early Cretaceous there were extensive areas of land with tectonically controlled sedimentation, whereas the Late Cretaceous experienced fully marine conditions with regional subsidence centred over the axial graben system (Hancock, 1990).
Figure 2.16 Thickness and structural setting of the Lower Cretaceous. (After Hancock, 1990).

Much of the tectonic setting of the Early Cretaceous was a continuation of that established in the Jurassic with ancient massifs remaining stable, fault controlled
basins subsiding further and in some cases, such as the Viking Graben, broadening of
the basins due to the formation of new faults outside the graben margins (Hancock,
1990). In some basins, such as parts of the Central Graben, subsidence was faster
than sediment supply, causing an increase of water depth. More commonly however,
the rate of sediment supply kept pace with subsidence in the Early Cretaceous basins
causing a common pattern of thick Lower Cretaceous in centres of local basins,
passing laterally through rapid changes of thickness across a series of half-grabens, to
disappear altogether within a few kilometres and the whole being blanketed by more
uniform Upper Cretaceous sediments deposited after fault movements had ceased
(Hancock, 1990).

Figure 2.17 Shows the approximate correlation of Cretaceous formations in the North
Sea area.

2.3.7.1 Lower Cretaceous
During the Lower Cretaceous, the North Sea was cut off from the southern European
Ocean by the London-Brabant Massif (Figure 2.16) and its extension across the
Rhenish Massif. Marine connections existed to the east through Germany and to the
north via the Viking Graben into the early Atlantic (Hancock, 1990). The major
transgressions of the Aptian and Albian more than doubled the submerged area of the
North Sea (Hancock, 1990). Outside the main depositional basins, Aptian and Albian
sediments dominate the Cretaceous system (Hancock, 1990).

In the Moray Firth Basin, the thickest sequence of arenaceous Cretaceous sediments
occurs. This 1150m thick pile consists of submarine fan conglomerates and turbidites
at the base of fault scarps and is associated with gravity-collapse on the upthrown
sides of contemporaneous faults and deposition in a deep topographic low (Hancock,
1990).
Figure 2.17 Approximate correlations of Cretaceous formations in the region of the North Sea. (After Hancock, 1990).
On the uplifted horst blocks in the Outer Moray Firth and Witch Ground Graben, a thin veneer of pelagic marls and limestones were deposited (Boote and Gustav, 1987). In the basins, on the other hand, debris flows and turbidites are again apparent (Hancock, 1990). These deposits are capped and sealed by Aptian marls and marly limestones which complete the transition to a low energy sedimentary basin.

Within the Viking Graben, deposition was continuous from the Jurassic, and hundreds of metres of shales of the Cromer Knoll Group (Figure 2.17) were deposited in relatively deep water (Ziegler, 1981; Hancock, 1990).

In the Central Graben, the Lower Cretaceous succession of grey shales and occasional marls such as the Plenus Marl (which form important marker horizons), reach thicknesses of up to 800m (Hancock, 1990).

The Lower Cretaceous succession in the Sole Pit Basin consists of basal sandstones of the Spilsby Formation, which are overlain by pyritic clays of the Speeton Clay Formation, and a ferruginous marlstone of Aptian-Albian age, known as the Red Chalk Formation (Hancock, 1990). Within the southern North Sea, both marine and non-marine Lower Cretaceous sediments are also present in the Broad Fourteens, Central and Western Netherlands Basins.

2.3.7.2 Upper Cretaceous

The Upper Cretaceous was characterised by a major rise in global sea level (Hancock, 1990). Most of the London-Brabant Platform, and the Mid-North Sea and Ringkøbing-Fyn High became submerged during the Albian (Figure 2.18). Submergence of the highs was completed during the Campanian, when the sea spread over the Fladen Ground Spur and the Halibut Horst (Hancock, 1990). It is clear from Figure 2.18 that, in the trenches away from the relatively uplifted areas, large thicknesses of Upper Cretaceous sediments were deposited: 1800m in the north Viking Graben, 1200m in the Central Graben, and 1000m in the South Halibut Basin.
and Witch Ground Graben. It is thought that much of the chalk in the trenches was deposited by mass-flows from the flanks (Hancock, 1990).

Figure 2.18 Isopachs of the Chalk Group (Upper Cretaceous plus Lower Paleocene) and Shetland Group in the region of the North Sea. (After Hancock, 1990).
In the southern North Sea, south of 57°N, most of the Upper Cretaceous is represented by chalk. This continues across the Cretaceous-Lower Palaeocene boundary as the Danian or Ekofisk Formation (Hancock, 1990).

Pelagic carbonate deposition gives way northward to calcareous claystone deposition, either as a consequence of climatic change or of increased fine clastic input from the Atlantic Rift shoulders (Doré and Gage, 1987). Over the Shetland Platform and in the Viking Graben north of 59°30', the succession is dominantly clastic and comprises 1000 to 2500 metres of pelagic marls and clays; only a few metres of Maastrichtian chalk are apparent (Hancock, 1990).

Within the Witch Ground Graben, Late Cretaceous carbonate deposition occurred in a stable marine basin, the topography being infilled and buried by pelagic and hemipelagic shales, marls and chalks (Boote and Gustav, 1987).

2.3.7.3 Cretaceous Inversions

During the Late Cretaceous most of the North Sea area was experiencing tectonic quiescence and broad crustal downwarping. However, some basins in the southern North Sea, such as the Sole Pit, Broad Fourteens and parts of the southern Central Graben, were experiencing inversion (Figure 2.18) (Olsen, 1987; Doré and Gage, 1987; Ziegler 1982). During basin inversion there is a change from basin subsidence to basin uplift often involving folding of the basin fill and reactivation of extensional faults as compressional faults. The basin sediments are often eroded during inversion.

The beginning of the Late Cretaceous saw the regional downwarping of wrench induced basins such as the Sole Pit Basin. (Ziegler, 1982). In the Senonian, subsidence of these basins, and also southern parts of the Central Graben, was reversed and the Sediment fill was folded and eroded. The Late Senonian to Danian saw the sea level advance and deposit the upper chalk unconformably on the inverted basin sediments (Ziegler, 1982).
The cause of these inversions is generally assumed to be linked to some kind of change in plate boundary configuration. Ziegler (1982) notes that the inversions are more intense in the basins closer to the Alpine deformation front to the south-east. Although the basin inversions at this time all seem to have occurred south of the Highland Boundary Fault Zone (Figure 2.2), Doré and Gage (1987) point out that the initial inversion of the Sole Pit Basin occurred in the Albian and was thus synchronous with the opening of the Rockall Trough (west of the Shetland Isles) indicating that reorganisation of the Atlantic plate margin might have influenced the stress regime in these southern North Sea Basins. A change in the direction of tectonic compression from NW-SE to NNE-SSW is documented at the end of the Early Cretaceous within the Danish part of the Central Graben (Olsen, 1987). This change resulted in right lateral reactivation of older left lateral shear zones in the Central Graben and led to compression in the southern North Sea. (Olsen, 1987).

The Late Cretaceous inversion events of the southern North Sea are discussed by Oudmayer and Jager (1993) in relation to reactivation of basement faults. These authors also emphasise the observation that the occurrence of halokinesis is related to inversion in the southern North Sea.

A subsequent and generally more intense inversion event occurred within the same basins during the Late Palaeocene (Ziegler, 1982). Concomitant with these inversions, the Rhenish and Bohemian Massifs became dissected and uplifted along a series of wrench and steep reverse faults. Further to the west however, the Channel and Hampshire Basins were only mildly deformed in the Late Cretaceous and Palaeocene events, their main inversion occurring at a similar time to that of the Western Approaches, Celtic Sea and Bristol Channel Troughs in the Oligocene to Miocene (Ziegler, 1982).

2.3.8 Cenozoic (65 Ma to Present)
The North Sea Basin exhibited its present day north-south configuration by Cenozoic times, the axis following the trend of the rift systems described throughout this chapter. Sediments deposited during this time are largely unfaulted and only slightly
deformed, halokinesis of Zechstein salts has a marked influence on formation 
thickness (Ziegler, 1982; Lovell, 1990).

Within the Cenozoic rocks of the North Sea Basin, two main phases of pyroclastic 
sedimentation are recognised. The first, dated around 58-57Ma, the other around 55-
52Ma. The top of the latter, the Balder Tuff Formation, is a widespread seismic and 
stratigraphic marker (Lovell, 1990). These ashes result from contemporaneous 
igneous activity to the west of the British Isles that was associated with a particularly 

Figure 2.19 illustrates the great changes in the palaeogeography of the British Isles 
from almost total submergence in the Upper Cretaceous to the emergence of a 
landmass exhibiting the shape of the present day British Isles in the Early Tertiary 
(Lovell, 1986).

The map illustrating the top of the Chalk structural contours (Figure 2.1) is roughly 
equivalent to a map of Cenozoic isopachs. The thickest Cenozoic sequences can be 
seen from to be centred over the Viking and Central Graben (Figure 2.1). The 
southern North Sea, outside of the southern end of the Central Graben, does not have 
a great thickness of Cenozoic sediments.

2.3.8.1 Palaeocene

During the Lower Palaeocene (Danian), subsidence of the North Sea Basin continued. 
Chalk deposition persisted within the Central Graben, whilst in the Viking Graben, 
clastic sediments became the dominant lithotype (Ziegler, 1981; Lovell, 1986).

Four major oil or gas-bearing sandstone units can be recognised within the 
Palaeocene of the central North Sea. Two of them were derived from the Scottish 
Highlands, and the other two were derived from the Shetland Platform (Lovell, 1990; 
Stewart, 1987). These clastic rocks comprise sands of deltaic, shelf, and submarine 
fan origin (Stewart, 1987). The large submarine fans are separated in basinal areas by 
hemipelagic muds (Lovell, 1990).
It has been proposed that a large Palaeocene fluvial delta built out into the Moray Firth and East Shetland Platform (Stewart, 1987). The Moray Firth delta developed a series of extensive submarine fans which prograded into the Central Graben and were responsible for the deposition of the major sand bodies of the Montrose, Forties and Frigg. At the end of the Palaeocene, coarse clastic input was halted by a basin-wide rise in sea-level and sedimentation was reduced to hemipelagic fallout of clays and tuffs (Anderton et al., 1985).
2.3.8.2 Eocene
Prograding sedimentation from the margins to the centre of the basin was coupled with slow subsidence (Lovell, 1986). Clastic supply from the west diminished, subsidence rates and rising sea levels are thought to have been greater than sedimentation rates, and the basin deepened (Ziegler, 1982).

2.3.8.3 Oligocene, Miocene and Pliocene
Throughout the Oligocene, water depths decreased (Lovell, 1986). This was followed by an increase in the rate of sedimentation and a slight increase in subsidence rates due to sedimentary loading (Ziegler, 1982; Lovell, 1986).

2.3.8.4 Quaternary
Quaternary sediments within the Central North Sea are between 500 and 1000m thick and are associated with an acceleration of subsidence rates which may have been brought about by a renewed build up of NNW-SSE oriented compressional stresses (Ziegler, 1990).

Following the Pleistocene glacial events the present-day geography of the North Sea Basin and surrounding areas was revealed. Following the removal of the ice-sheet from Northern Britain the UK mainland has undergone isostatic uplift centred on the point of maximum loading in SW Scotland (Anderton et al., 1985).

2.4 Crustal Scale Structure of the North Sea Basin
Crustal Scale Geophysical Studies of the North Sea Basin include deep seismic reflection profiling (Klemperer and Hurich, 1990), gravity (Holliger and Klemperer, 1990, Fichler and Hospers, 1990) and seismic refraction profiling (Christie and Sclater, 1980). These techniques can be used individually or in unison to determine the thickness of the crust and there appears to be generally good agreement between the different methods. Features such as crustal scale faults (e.g. Great Glen Fault)
and suture zones (e.g. Iapetus Suture) can also be detected with seismic methods (Klemperer and Hurich, 1990).

The present crustal configuration derived from the geophysical methods listed above shows that the Mesozoic North Sea Rifts are related to thinned crust as is predicted by models of basin formation (McKenzie, 1978; Sclater and Christie, 1980; Wernicke, 1985). The zone of maximum crustal thinning coincides with the trace of the Viking and Central Graben (Ziegler, 1990). Refraction and deep seismic reflection lines indicate that the Moho rises from a depth of 33-34 km under the coast of Norway to 22-24 km and 20 km under the Viking and Central Grabens respectively, and descends to a depth of 30 and 32 km under Britain and the Shetland platform, respectively (Ziegler, 1990). Under the axial parts of the Central and Viking Grabens, in which sediments reach a thickness of 8-10 Km, the continental crust is 60% thinner than beneath the surrounding land masses (Ziegler, 1990). The centre of Cenozoic subsidence coincides with high gravity anomalies and also with the culmination of the mid-Jurassic North Sea Rift dome (see section 2.3.6.2).

Figure 2.20 shows crustal scale cross sections across the Viking Graben constructed from deep seismic reflection data. Sections such as this can be constrained using deep refraction and gravity data. It can be seen from Figure 2.20 that the Moho, although it is much smoother, broadly mirrors the top of the faulted crust. Crustal scale cross sections of this type have been used to assess the degree of symmetry across the North Sea Rifts in an attempt to distinguish between the pure shear (symmetrical stretching) model and the simple shear (asymmetric detachment) model (White, 1990).
Recent free-air gravity mapping of the North Sea using satellite obtained data clearly reveals some of the crustal structure, with gravity highs following the axes of the major Mesozoic rifts (Figure 2.21). The gravity anomalies shown in Figure 2.21 are free-air gravity anomalies the highest of which occur along the north-western British
Continental Margin including the North and West Shetland Platforms and the West Shetland Basin. There is a noticeable rim of gravity highs around the coast of Norway. It has been noted on previous gravity maps that there is a trend of highs following the line of the Mid North Sea High (Cowgill, 1994). This trend is discernible but not very clear in Figure 2.21.

There is a deep gravity low of limited areal extent in the Inner Moray Firth which is bordered to the north-west by the Great Glen Fault. There are also deep gravity lows in the Rockall Trough. The southern North Sea is largely occupied by a gravity low which seems to be broadly coincident with the Southern Permian Gas Basin (section 2.3.4) and is dissected by the southern end of the Central Graben.

Gravity modelling is performed using gravity measurements and assumed values for the densities of different parts of the crust and mantle. In the North Sea, gravity modelling has been used to constrain the crustal structure and map the Moho (Fichler and Hospers, 1990; Holliger and Klemperer, 1990). Across the North Sea Grabens, the Moho is imaged using deep seismic reflection beneath the platform areas on either side but only sporadically imaged beneath the graben itself (Fichler and Hospers, 1990). The position of the Moho obtained from gravity modelling, the gravity Moho - defined by a density contrast, coincides with the seismic Moho under the platform areas either side of the graben (Fichler and Hospers, 1990; Holliger and Klemperer, 1990). This coincidence implies that the Moho is the petrological boundary between the mantle and crust.

Structural models based on calculated and observed Bouguer gravity anomalies and BIRPS (British Institutions' Reflection Profiling Syndicate) deep seismic reflection profiles have been constructed for several lines of section across the North Sea (Holliger and Klemperer, 1990). Figure 2.22 shows two such models at the crustal and basin scale for sections across the central and southern North Sea.
Figure 2.21 North Sea free air gravity map. (Constructed by R. Holt, pers. com. 1996). These models show the Moho broadly mirroring the faulted upper surface of the crust with the maximum thickness of sediments occurring above the thinnest parts of the crust. It can be seen that the crust of the southern North Sea is not as severely thinned as in the central North Sea, which is representative also of the northern North Sea (Fichler and Hospers, 1990). The southern North Sea did not experience the same degree of Mesozoic rifting as the central and northern parts which were affected by
the Central and Viking Grabens. In the Central and Viking Grabens, dense mantle is present beneath the thinned crust, causing positive gravity anomalies over the rifted areas. The southern extension of the Central Graben in the southern North Sea is also underlain by dense mantle causing a higher anomaly than the surrounding area but it is weaker than the equivalent anomalies in the central and northern North Sea (Figure 2.21 and 2.22).

The absence of large negative Bouguer gravity anomalies on a basin wide scale in the North Sea implies that the area is largely isostatically compensated by the mantle anti-root vertically beneath the basins (Holliger and Klemperer, 1990). On a local scale some structures may not be compensated. Loads at the base of columns from the surface to the Moho in the models of Holliger and Klemperer (1990) increase by up to 350 MPa from the basin margins to the centre of the grabens and require an elastic thickness of the lithosphere of 2-3 Km. Barton and Wood (1984) found that a lithospheric elastic thickness of about 5 km best fitted the observed gravity field of the North Sea.

The models of Holliger and Klemperer (1990) imply that the Palaeozoic loads (granites or basins) are uncompensated in contrast to the large degree of compensation found for most of the Mesozoic structures. This means that the elastic strength of the lithosphere is enough to support the loads of the Palaeozoic basins and agrees with the observation of a gravity low over the Southern Permian Basin in the southern North Sea.
Figure 2.22: (A); Map of North Sea showing location of profiles in (B) and (C). (B); Gravity model for NSDP85-5 shown on (A). (C); Gravity model for SNST83-7 shown on (A). (After Holliger and Klemperer, 1990).
2.4.1 Palaeozoic Crustal Structure of the North Sea Basin

As discussed in section 2.3, it has frequently been speculated that the structural evolution of the Mesozoic triple rift system has been controlled by Palaeozoic structures which form fundamental zones of weakness in the crust (Glennie, 1990). It has been suggested, for example, that the Viking Graben overlies a Palaeozoic transform plate boundary (Watson, 1985). This type of suggestion can rarely be tested because this type of structure is rarely reached during drilling.

Some of the major Palaeozoic features or postulated features have been investigated using deep seismic reflection profiling (Klemperer and Hurich, 1990). The Iapetus Suture formed around the middle of the Palaeozoic at the closure of the Iapetus ocean and separates the continental blocks of Laurentia and Baltica (Figure 2.4). As such it is a very important and fundamental boundary, yet it is not exposed in Britain, being everywhere buried by Late Palaeozoic sedimentary cover. The zone of the suture can be seen as a north dipping reflector on deep seismic reflection profiles and traced for over 900 km from west of Ireland, through Britain and into the North Sea (Klemperer and Hurich, 1990). Its trend in the North Sea is ENE, which despite being almost orthogonal to the Mesozoic Rift in this part of the North Sea, is parallel with trends established in Carboniferous and Permian Basin formation (see sections 2.3.3 and 2.3.4).

The Iapetus Suture coincides closely with the Mid North Sea High (Figure 2.23). The Mid North Sea High is not a fault bounded structure and is defined only on its topographic expression through time. It seems plausible that the crust of the suture zone may have been thickened during continental collision and become a zone of positive relief (Klemperer and Hurich, 1990). The Mid North Sea High extends east towards the Ringkøbing-Fyn High on the eastern side of the Central Graben, however, due to lack of suitable seismic data it is not known whether the Iapetus Suture can be traced this far (Klemperer and Hurich, 1990).
The Great Glen Fault (and its probable northern extension, the Walls Boundary Fault) is an important predominantly strike slip fault (Figure 2.3) the reactivation of which has been demonstrated during basin development from Devonian to the Mesozoic (Klemperer and Hurich, 1990). The deep seismic profile across the Great Glen/Walls Boundary Fault shows that this fault is a crustal penetrating fault even today and that it appears to offset the Moho (Klemperer and Hurich, 1990).

![Map illustrating the relationship between the Iapetus Suture mapped from north-dipping reflectors on deep seismic reflection profiles (the tip of each arrow head is located at the position where the reflectors meet the Moho) and the Mid-North Sea High (drawn as area where Zechstein halites are absent). The synform symbol (opposed arrow heads) marks a zone 40 km wide where the crust is still thickened. (After Klemperer and Hurich, 1990).](image)

Other important Palaeozoic features such as the Highland Boundary Fault (Figure 2.2) and the Tomquist Zone (Figure 2.22) are well defined on land and were obviously important Palaeozoic features although their offshore extensions can not be seen on deep seismic reflection profiles which brings into question the degree of influence they have had since the Palaeozoic (Klemperer and Hurich, 1990).

2.5 Seismotectonics of the North Sea Area

Earthquake focal mechanisms (section 4.2.1), from most of the North Sea region, are rarely obtained. There are two reasons for this, firstly it seems that the level of seismicity
is generally low within large parts of the North Sea (Muir Wood, 1985), and secondly, the small amount of seismicity that occurs in most of the North Sea, is rarely recorded accurately enough to determine the focal mechanism, due to the lack of suitable seismic instrumentation within a sufficiently local area (pers com, Ritchie, 1996). Seismically active areas in the North Sea and surrounding regions, have been presented by Muir Wood (1985) (Figure 2.24).

In Figure 2.24, seismically active areas are those which have experienced either "felt seismicity", much of which may be historical, or instrumentally detected seismicity, which is of course relatively recent. Some areas, such as western Scotland and western/offshore Norway are singled out as areas in which the instrumentally recorded seismicity is similar to the historically recorded seismicity. In these areas, the levels of seismicity are relatively high, with earthquakes occurring frequently.

The stress regime (section 3.1.1) can only be determined if a focal mechanism is obtained for the earthquake. The only focal mechanisms available for the North Sea itself, are from offshore Norway. In fact, this part of the northern North Sea, which is approximately marked in Figure 2.24 as the area in which instrumental and historic
seismicity is similar, is one of the most seismically active regions of Northwest Europe (Lindholm and Marrow, 1990). A significant proportion of this seismicity, as well as the seismicity in western Scotland, has been related to the effects of post glacial rebound (Muir Wood, 1985) which it is thought disguises the effects of tectonic stress. However, some of the larger recent Earthquakes of approximate surface wave magnitude (M_s) 5, from offshore Norway, have been studied in detail, and indicate focal mechanisms consistent with intraplate stress orientations derived from neighboring areas (Hansen et al., 1989). These earthquakes, as well as numerous other smaller earthquakes in this region (Lindholm et al., 1995) predominantly show significant reverse fault components, indicating that the vertical stress is the minimum stress (Figure 2.25). The focal mechanisms also indicate an approximately NW/SE to E/W maximum horizontal stress direction, which is consistent with the hypothesis that the maximum horizontal stress direction arises from the ridge push force from the mid Atlantic Ridge.
Figure 2.25 shows the focal mechanisms compiled by Lindholm et al. (1995) from the offshore Norway region of the northern North Sea. The focal depths for these earthquakes have been determined as being between 5 and 15 km on the Horda platform and between 20 to 30 km beneath the Viking Graben (see Figure 2.2 for location of these structural elements). The triangle plot included in Figure 2.25 shows the relative proportions of inferred stress regime from the focal mechanisms. It is clear that at these depths in this area of the northern North Sea, the stress regime is predominantly reverse to strike slip. The rose diagram in Figure 2.25 shows the direction of maximum horizontal stress directions inferred from the focal mechanisms.
It seems clear that the orientation of maximum horizontal stress in this part of offshore Norway is due to the tectonic stress associated with the mid Atlantic ridge. However, it has been noted by Lindholm et al. (1995) that there is some evidence of regions with shallower focal depths being dominated by normal faulting, and that this variation in the relative stress magnitudes with depth may be to some extent due to the effects of glacial rebound.

Although these focal mechanisms from offshore Norway are the only ones available from the North Sea itself, focal mechanism from onshore areas surrounding the North Sea in Britain and Northwest Europe have been determined.

The area to the south east of the southern North Sea (northern France, Belgium and The Netherlands) is shown in Figure 2.24 as being seismically active. The area which is shown as having historical seismicity similar to recent seismicity, experiences frequent seismic events, and corresponds broadly to the area shown in Figure 2.26. The Roer Valley Graben and associated structures shown in Figure 2.26 form part of the lower Rhine embayment which is part of the Rhine Graben system. The focal mechanism shown in Figure 2.26 (the 17 km deep, almost purely normal focal mechanism of the Roermond M, 5.3 earthquake, and the M, 4.6 strike slip Liege earthquake) correspond to the largest seismic events that have occurred in the area in the last 20 years. The faults mapped in Figure 2.26 are all quaternary faults. The quaternary fault system is dominated by normal faults which strike approximately NW/SE. Earlier focal mechanisms from in and around the Rhine Graben and the lower Rhine embayment have been compiled by Ahorner (1975). These show a variety of focal mechanisms at depths of between 5 and 20 km which are clearly dominated by normal movement on NW/SE striking faults. The stress regime at these depths in the crust, in this part of Europe, is clearly a predominantly normal faulting regime. It is not clear whether these fault trends extend into the southern North Sea, and so inferences about the stress regime in the southern North Sea must be tentative.
The majority of Britain is shown in Figure 2.24 as being seismically active. This is based on both historical seismicity and instrumentally detected seismicity. Although seismicity has been instrumentally detected for many parts of Britain, the only focal mechanisms available are all close to the west coast. Figure 2.27 shows the British earthquakes for which focal mechanisms have been determined. It can be seen that the focal mechanisms are predominantly strike slip, although some (e.g. the Dunoon earthquake) have a component of normal movement. In most cases the fault plane can not be linked to any surface geological features. The largest of these earthquakes, the Lleyn Earthquake of 1984 (also estimated to be the largest in Britain this century) had a body wave....
magnitude of 5.2 (Marrow and Walker, 1987) and a focal depth of around 20 km (Marrow and Walker, 1987; Trodd et al., 1985).

Figure 2.27 Locations of focal mechanisms (shaded areas indicate compressional first arrivals) in the UK. These indicate a predominantly strike-slip mechanism with a component of normal faulting. (After Marrow and Walker, 1988)

Seismicity in and around the North Sea: A Summary.
In the parts of the northern North Sea close to the Norwegian coast at depths below 5 km, the stress regime is predominantly reverse faulting, although both strike slip and normal faulting focal mechanisms also occur. In the northern North Sea, further from the Norwegian coast, and in the central and southern North Sea the stress regime is unknown. Focal mechanisms have been determined in parts of western Britain where they are predominantly strike slip, and within the lower Rhine embayment where they are predominantly normal. However, extrapolation of these stress regimes into the North Sea would be very tentative. Moreover, extrapolation of stress regimes from the depth of focal mechanisms, to depths sampled by oilwell drilling in the North Sea may also not be valid.

The focal mechanisms in the areas surrounding the North Sea are included in the World Stress Map. The European section of the World Stress Map has been enlarged and included here (Figure 2.28). It can be seen from this part of the World Stress Map that the North Sea as whole is surrounded by a variety of stress regimes, and also that, at
present, the stress regime in the North Sea, other than in the area offshore Norway, is unknown.
Figure 2.28. The European part of The World Stress Map showing the direction of azimuth of the maximum horizontal stress and the stress regime where known. Downloaded from http://www-gpi.physik.uni-karlsruhe.de/pub/wsm/index.html
Chapter Three - Lithospheric Stress

3.1 Introduction
This chapter briefly introduces the concept of stress and then goes on to discuss the origins, distribution and measurement of stress in the lithosphere. Since any discussion of lithospheric stress will necessarily be inextricably linked to the theory of plate tectonics, a brief outline of the main principles of the latter has also been included.

3.1.1 Definitions of Stress State and Other Terms
For a thorough review of fundamentals see Jaeger and Cook (1979). A few basic definitions of terms which are used in this study are briefly described here. Stress is force per unit area and is measured in Newtons per square metre (Pascals) or bars (atmospheric pressures) or pounds per square inch (psi). Stresses are either normal stresses, which act perpendicular to a surface, or shear stresses, which act parallel to a surface.

The three dimensional state of stress at a point can be represented by the stress components on the sides of an infinitesimal cube which approximates a point in a body. The state of stress is then completely defined by a nine part tensor (three normal stresses and six shear stresses). If the axes of this cube are defined as being parallel to the axes $x_1$, $x_2$, $x_3$, then the complete stress tensor, $\sigma_{ij}$, can be written:

$$
\sigma_{ij} = \begin{pmatrix}
\sigma_{11} & \sigma_{12} & \sigma_{13} \\
\sigma_{21} & \sigma_{22} & \sigma_{23} \\
\sigma_{31} & \sigma_{32} & \sigma_{33}
\end{pmatrix}
$$

Where the index $i$ in $\sigma_{ij}$ is the pole to surface on which the stress acts, and the index $j$ is the direction in which the stress acts. For example, the stress $\sigma_{11}$ is the stress acting on the surface containing the $x_2$ and $x_3$ axes and is acting in the $x_1$ direction which is perpendicular to that surface, it is therefore a normal stress, as are $\sigma_{22}$ and $\sigma_{33}$. The stress component $\sigma_{12}$ again acts on the surface containing the axes $x_2$ and $x_3$ but it acts in the direction of $x_2$ which is parallel to the plane and it is therefore a shear stress as are $\sigma_{13}$, $\sigma_{21}$, $\sigma_{23}$, $\sigma_{31}$ and $\sigma_{33}$. When the body is in equilibrium the opposing shear stress
components must balance such that: $\sigma_{12} = \sigma_{21}$, $\sigma_{13} = \sigma_{31}$, $\sigma_{23} = \sigma_{32}$. There are then only six independent stress components.

- **Principal Stresses:** At any point there are always three mutually perpendicular planes on which no shear stresses act. These are the planes of principal stress and the normal stresses acting on these planes are the principal stresses (the three normal stresses in the stress tensor: $\sigma_{11}, \sigma_{22}$, and $\sigma_{33}$). The three principal stresses are usually denoted simply $\sigma_1 > \sigma_2 > \sigma_3$. In geological literature, the convention is for compressive stresses to be positive and tensile stresses to be negative. This is because tensile stresses are uncommon in the lithosphere except at very shallow depths.

- **Hydrostatic and Deviatoric Stress:** The stress tensor can be divided into two parts, the hydrostatic part and the deviatoric part. The hydrostatic part consists only of the mean normal stress (the mean of the three principal stresses) acting equally in all directions:

$$
P_{ij} = \begin{pmatrix}
\sigma_m & 0 & 0 \\
0 & \sigma_m & 0 \\
0 & 0 & \sigma_m
\end{pmatrix}
$$

(3.2)

Where $P_{ij}$ is the hydrostatic stress and $\sigma_m$ is the mean stress given by:

$$
\sigma_m = \frac{\sigma_{11} + \sigma_{22} + \sigma_{33}}{3}
$$

(3.3)

The deviatoric stress is the rest of the stress tensor, it therefore consists of shear stresses and deviations of the normal stresses from the mean. The deviatoric stress $\sigma_{ij}^d$ is given by:

$$
\sigma_{ij}^d = \begin{pmatrix}
\sigma_{11} - \sigma_m & \sigma_{12} & \sigma_{13} \\
\sigma_{21} & \sigma_{22} - \sigma_m & \sigma_{23} \\
\sigma_{31} & \sigma_{32} & \sigma_{33} - \sigma_m
\end{pmatrix}
$$

(3.4)
In a fluid at rest, because the fluid effectively has no long term shear strength and can support no shear stress over a sufficiently long period of time, all three principal stresses must be equal, hence there is no deviatoric stress (hence the term hydrostatic).

- **Differential Stress**: Distinct from deviatoric stress, differential stress is simply the difference in magnitude between two principal stresses in a plane. Usually the differential stress ($\sigma_d$) in a geomechanical context simply means the difference between the maximum and minimum principal stresses:

$$\sigma_d = \sigma_1 - \sigma_3$$

(3.5)

- **Stress Regimes**: The stress regime refers to the relative magnitudes of the vertical and horizontal stresses which dictate the style of faulting which will occur, should the differential stress be high enough. A simple classification was first introduced by Anderson (1905, 1951) where it is assumed that the vertical stress is a principal stress. As the surface of the Earth is a free surface and can support no shear stress, the vertical stress will be a principal stress at least at the surface, and will continue to be so at depth in the absence of topography and non-horizontal tectonic stresses. The Andersonian classification of stress regimes in the lithosphere is illustrated in Figure 3.1. The three stress regimes are:

  - Normal Faulting Regime - $\sigma_v > \sigma_H > \sigma_h$
  - Strike Slip Faulting Regime - $\sigma_H > \sigma_v > \sigma_h$
  - Thrust Faulting Regime - $\sigma_H > \sigma_h > \sigma_v$

Where $\sigma_H$ is the maximum horizontal stress, $\sigma_h$ is the minimum horizontal stress and $\sigma_v$ is the vertical stress.
Thrust-Fault Regime

Strike-slip Fault Regime

Normal-Fault Regime

\[ S_v = \sigma_1 \]
\[ S_h = \sigma_2 \]

\[ S_v = \sigma_3 \]
\[ S_h = \sigma_2 \]
\[ S_h = \sigma_1 \]

\[ S_v = \sigma_1 \]
\[ S_h = \sigma_3 \]

\[ z \]
\[ y \]
\[ x \]

Figure 3.1 The three states of stress associated with thrust, strike-slip and normal faulting as classified by Anderson (1951) (After Engelder, 1993).

- **Total and Effective Stress:** In a rock which consists of grains in contact with each other and connected pore spaces between the grains which are filled with a fluid, the total stress or confining pressure applied at the boundaries of the material is transmitted through the material partly by the fluid and partly by the framework of grains. The effective stress \((\sigma')\) is that part of the total stress which the solid framework effectively transmits, it is defined as:

\[ \sigma' = \sigma - \alpha \rho \]  \hspace{1cm} (3.6)

Where \(\sigma\) is the total stress, \(\rho\) is the pore fluid pressure, and \(\alpha\) is a poroelastic co-efficient. The co-efficient \(\alpha\) is dependent on the bulk modulus of the rock framework (grains plus pores) compared to that of the solid grains themselves. The co-efficient \(\alpha\) can be found from the equation derived by Nur and Byerlee (1971):

\[ \alpha = 1 - \frac{k}{k_s} \] \hspace{1cm} (3.7)

Where \(k\) is the bulk modulus of the rock framework and \(k_s\) is the bulk modulus of the solid grains of rock. When the bulk modulus of the framework is very low compared to the bulk modulus of the grains (e.g. in an unconsolidated sediment) \(\alpha\) will be close to 1.
If however, there is very little pore space and the rock is very compact (e.g. some igneous rocks) $\alpha$ may be significantly less than 1.

### 3.2 Plate Tectonics

The Earth has a cool and relatively rigid outermost shell called the lithosphere which comprises the crust and uppermost mantle. It is thinnest at mid ocean ridges and thickest in cratonic continental regions where its base is poorly understood (Fowler, 1990). An average lithospheric thickness of around 100 km has been suggested by Turcotte and Schubert (1982). Below the lithosphere is the asthenosphere where high temperatures and pressures cause the viscosity of the material to be low enough to allow flow on a geological time scale (Fowler, 1990).

The lithospheric shell is divided into the tectonic plates illustrated in Figure 3.2. The theory of plate tectonics describes the motions and interactions of lithospheric plates as they move on a sphere over the underlying asthenosphere. Plate tectonic theory recognises the lithospheric plate as the fundamental kinematic unit, each plate having its own velocity vector. The deformation which results from the motion of the plates is thought to take place mainly at the plate edges, i.e. at the plate boundaries. Figure 3.2 shows the major plates which make up the Earth's surface layer.

#### 3.2.1 Oceanic and Continental Plates.

Oceanic plate material is generated at mid-ocean ridges where sea floor spreading occurs. Oceanic crust is composed largely of basaltic and gabbroic material which makes oceanic lithosphere more dense than continental lithosphere and as it moves away from the mid-ocean ridge and cools it becomes more dense than the underlying asthenosphere. The Pacific plate (Figure 3.2) is almost entirely oceanic. Most of the other major plates are part oceanic, part continental.

Continental plates are generally of lower density than oceanic plates (section 3.3.4.1). The upper continental crust is composed mainly of granitic material and is enriched in lithophile elements which tend to form low density minerals. As a result of the density contrast,
Figure 3.2 The major tectonic plates, mid-ocean ridges, trenches and transform faults. (After Fowler, 1990).
the continents act as rafts of lower density material which remain on the surface while
the denser oceanic material can be subducted back into the Earth’s interior.

3.2.2 Plate Boundaries

Plate boundaries are of three general types: divergent, convergent and conservative
(Fowler, 1990).

At divergent plate boundaries (also called accreting or constructive boundaries) plates
move away from each other. At such boundaries, new material, derived from the mantle,
is added to the lithosphere. The divergent plate boundary is represented by the mid-ocean
ridge system, along the axes of which new plate material is produced. The Mid-Atlantic ridge (Figure 3.2) is an example of a divergent plate boundary. Continental rifts
may also become divergent plate boundaries.

Convergent boundaries are also called consuming or destructive boundaries and occur
where plates are approaching each other. Most such boundaries are represented by
subduction zones where the cold dense oceanic plate, being gravitationally unstable with
respect to the underlying asthenosphere, founders and begins to sink (beneath either a
continental or another oceanic plate) into the interior of the earth (Turcotte and
Schubert, 1982).

A second type of convergent boundary occurs where continental parts of plates collide
following the complete subduction (or obduction) of the intervening oceanic lithosphere.
An example of such a boundary is the Alpine-Himalayan mountain belt which has
formed as the result of a collision between the continental parts of the African, Indian
and Eurasian plates (Figure 3.2).

Conservative boundaries occur where adjoining plates move past each other with an
essentially horizontal strike-slip sense of displacement along transform faults such as the
San Andreas fault in California and the Alpine fault in New Zealand. In this case
lithosphere is neither emplaced or displaced (Turcotte and Schubert, 1982).
These three types of plate boundary describe only purely orthogonal or parallel relative plate motions and as such are idealised end members of an array of possible boundaries. Plate motion vectors may not be purely orthogonal or parallel to the plate boundaries which consequently often display transpressional or transtensional features. Examples of transpression and transtension are seen along parts of the San Andreas fault (Zoback, 1991).

3.3 Forces acting on the Lithosphere
In this section I describe the main types of forces which act on the lithosphere and which are thought to be important sources of stress. The first part of this section (section 3.3.1) attempts to define the types of forces (body forces and surface forces) which act on the lithosphere. The second part (3.3.2) attempts to distinguish between tectonic forces and non-tectonic forces. Stress states in the absence of plate tectonics (sometimes called reference states) form the starting point for the discussion of lithospheric stress. Moving on to also consider tectonic forces, the relationship between tectonic forces and lithospheric stress is discussed in the context of plate tectonics and the World Stress Map. Opposing views on the rheology of the lithosphere as a whole are outlined in the last part of this section.

3.3.1 Body Forces and Surface Forces
The forces acting on an element of a solid are of two types: body forces and surface forces. Body forces act throughout the volume of the solid, such as the effect of gravity acting on a mass so that the body force is the weight of this mass. Surface forces act on the surface area bounding an element of volume. They arise from interatomic forces exerted by material on one side of the surface against material on the opposite side (Turcotte and Schubert, 1982). Both body forces and surface forces act on the lithosphere and both can be either tectonic or non-tectonic forces, examples of these forces are given in section 3.3.5.

3.3.2 Tectonic and non-Tectonic Forces
Tectonic forces are those which exist directly as a consequence of a dynamic plate tectonic system. Non-tectonic forces are those which exist in the lithosphere today, but would exist anyway, even in the absence of plate tectonics. Stress in the lithosphere then, is not due entirely to tectonic forces. Non-tectonic body and surface forces are present under static conditions such as a block of continental crust simply floating on mantle of a higher density (Figure 3.3). In this simple case the body forces are equal to the overburden weight (assuming an isostatic stress state, see section 3.3.3.1) and are balanced by the surface forces which exist at the sides of the block exerted by the crust and the mantle.

Tectonic forces can also be divided into both body forces and surface forces. Tectonic forces are ultimately a consequence of density differences derived from the mechanisms of cooling by which the Earth releases internal heat. Tectonic forces are discussed in section 3.3.5, but first the state of stress in the absence of tectonic forces is examined in sections 3.3.3 and 3.3.4.

### 3.3.3 Simple Reference States of Stress

While developing an understanding of lithospheric stress, it is convenient to start with a reference state which may occur in the absence of tectonic stress (Engelder, 1993). Two such reference states, the so called lithostatic reference state (Engelder, 1993) and the uniaxial reference state are very simple and can be defined by making assumptions about the rock properties and boundary conditions which are discussed below.

#### 3.3.3.1 Lithostatic Reference State

The simplest reference state is that of isostatic stress, where the stress is the same in every direction. This reference stress, which has been termed the lithostatic reference state when applied to the lithosphere, is analogous to the hydrostatic stress state found in a fluid at rest. A fluid has no shear strength and so cannot support any differences between the three principal stresses which therefore become equal ($\sigma_1 = \sigma_2 = \sigma_3$).

An isostatic stress state in the lithosphere ($\sigma_v = \sigma_h = \sigma_k$) will develop if the rock has no long term shear strength such that it creeps plastically. Although rocks such as weak
shales and halite have very low shear strength, they seem to be able to support small shear stresses over long periods of time so that a perfectly isostatic stress state in the lithosphere is rare (Engelder, 1993). The lithostatic stress reference state however, remains a useful concept.

3.3.3.2 Uniaxial-Strain Reference State

The uniaxial-strain reference state is based on the postulated boundary condition that strain is constrained at zero across all fixed vertical planes (Price, 1966; Engelder, 1993). Such a boundary condition leads to a stress state which approximates newly deposited “weak” sediments in a sedimentary basin with “rigid” boundaries and is discussed in more detail later in this chapter (section 3.4.1). The uniaxial strain reference state arises, as its name suggests, through strain in only one direction, the vertical direction, due to the weight of the overlying rock. The rock is prevented from expanding laterally by the “pressure of the surrounding rock” (Price, 1966).

The assumption of uniaxial strain, applied to an elastic body of rock, leads via Hooke’s law to the relationship:

\[ \sigma_h = \sigma_h = \sigma_v \left( \frac{v}{1 - v} \right) \]  

where \( v \) is Poisson’s ratio and the stresses, where a pore fluid is present, are the effective stresses.

Typical values of Poisson’s ratio for rocks are between 0.1 and 0.35. The uniaxial strain reference state therefore predicts the two horizontal principal stresses to be equal to each other and to be between one ninth and around a half of the vertical stress.

The fact that observations of folding and faulting in the crust are common for rocks in most geological environments clearly shows that most rocks have non-zero horizontal strain. The uniaxial strain state, like the lithostatic state, is rare, but again it provides a useful reference state of stress.

3.3.4 Reference States of Stress in Non-tectonic Isostatically Compensated Cases.
Section 3.3.3 dealt with very simple reference states which do not consider the lithosphere as a whole and therefore do not include stresses induced by the fact that the lithosphere is "floating" on the asthenosphere. It is useful to consider some simple models of lithospheric stress in the absence of tectonic stress which make use of the assumptions of isostasy and static horizontal force equilibrium described by Artyushkov (1973). This type of model applies to any situation where one material is "floating" in another and so has been applied to both crust/mantle systems and lithosphere/asthenosphere systems. Two simple models are described below (3.3.4.1) where blocks of uniform thickness crust simply float on the underlying mantle. These models and the equations which are used to calculate the stresses both come from Turcotte and Schubert, (1982). A similar situation using the same assumptions is then considered which takes into account the effect of erosion of part of crust (3.3.4.2). This is the model of McGarr, (1988) which has been termed the "constant horizontal stress reference state" by Engelder, (1993).

3.3.4.1 Simple Block of Crust in Mantle Reference State.
This model uses the assumptions of isostacy and horizontal force balance that are described by Artyushkov, (1973) and by Turcotte and Schubert, (1982). The model also assumes that the horizontal stress due to the vertical stress is equal to the vertical stress (an isostatic stress state in the absence of other stresses). Two simple cases (Figure 3.3 and 3.4) are presented for which stress states have been calculated.

![Figure 3.3 A block of crust "floating" in mantle, where \( \rho_c \) = density of crust, \( \rho_m \) = density of mantle, \( h \) = thickness of crust, \( b \) = depth of compensation in mantle. (After Turcotte and Schubert, 1982).]
Figure 3.3 shows a uniform thickness block of continental crust floating on higher density mantle. Taking values \( h = 35 \text{ km} \), where \( h \) is the thickness of continental crust, \( \rho_m = 3300 \text{ kg.m}^{-3} \) and \( \rho_c = 2750 \text{ kg.m}^{-3} \), where \( \rho_m \) and \( \rho_c \) are the densities of the mantle and continental crust respectively, it is shown (Turcotte and Schubert, 1982) that in order to maintain the integrity of the continental block, the vertically averaged deviatoric horizontal stress must be tensile and have a value of \(-80.2 \text{ MPa}\). Deviatoric tensile horizontal stresses give rise to tensional and extensional faulting. Deviatoric tensile stresses of this level would not be sustained, at least in the upper parts of the crust. Tensional and extensional faulting would occur in order to dissipate the excess potential energy. In other words the block of continental crust would collapse extensionally under its own weight. The model illustrated in Figure 3.3 is therefore unrealistic.

The situation illustrated in Figure 3.3 is unrealistic of course because there is no oceanic crust or water. Figure 3.4 shows an alternative configuration of isostatically compensated crust, mantle and water. Taking values for thickness of continental crust \( (h) = 35 \text{ km} \), thickness of oceanic crust \( (h_{oc}) = 7 \text{ km} \), depth of water \( (h_w) = 5 \text{ km} \), density of oceanic crust \( (\rho_{oc}) = 2900 \text{ kg.m}^{-3} \), density of water \( (\rho_w) = 1000 \text{ kg.m}^{-3} \) and \( \rho_m = 3300 \text{ kg.m}^{-3} \) and \( \rho_c = 2800 \text{ kg.m}^{-3} \) and using the equations of Turcotte and Schubert (1982) and the same assumptions as were used in the case of Figure 3.3, the vertically averaged horizontal deviatoric stress is less than 1 MPa.

![Diagram](https://via.placeholder.com/150)

Figure 3.4 Isostatically compensated continental and oceanic crust structure where \( h_w = \) water depth, \( h_{oc} = \) thickness of oceanic crust, \( h_{cc} = \) thickness of continental crust, \( \rho_{cc} = \) density of continental crust, \( \rho_w = \) density of water, \( \rho_{oc} = \) density of oceanic crust, \( \rho_m = \) density of mantle. (After Turcotte and Schubert, 1982).
The situation illustrated in Figure 3.4 is a reasonable approximation to the tectonic configuration of much of the earth and seems to imply that generally, in the absence of topographic differences, it will be tectonic forces which are responsible for large deviatoric stresses. Such tectonic forces may actually be due lateral density/topography changes such as in the case of ridge push (see section 3.3.5.1) Although Figure 3.3 is unrealistic, it does illustrate how lateral density change and associated topographic differences can generate deviatoric stresses. Specifically, it also illustrates how if, for some reason, the oceanic crust shown in Figure 3.4 was not directly juxtaposed against the continental crust (or if it was moving in the same direction with a higher velocity than the continental crust) extensional deviatoric stresses would be generated in the continental crust.

3.3.4.2 Constant Horizontal Stress Reference State.

The constant horizontal stress reference state (Engelder, 1993) is somewhat ambiguously named. This model examines the effects of change in lithospheric thickness and temperature on the state of stress in the lithosphere. Again it is based on the assumptions of isostasy and static horizontal force equilibrium employed by Artyushkov (1973) and Turcotte and Schubert (1982). The constant horizontal stress reference state (Engelder, 1993) is reproduced from the model of McGarr (1988) in which the term lithosphere is used to denote the layer between the surface and the depth of compensation and so can be applied to either the crust/mantle or the lithosphere/asthenosphere systems.

The starting point is a piece of lithosphere of uniform thickness, \( Z_i \), in which an isostatic stress state exists. One part of the lithosphere is eroded (Figure 3.5) by a thickness, \( Z \), and isostatic compensation occurs elevating the base of the lithosphere by an amount \( W \):

\[
W = \frac{\rho_L Z}{\rho_m}
\]  

(3.9)

Where \( \rho_L \) and \( \rho_m \) are the densities of the lithosphere and underlying (asthenospheric) mantle respectively.
Figure 3.5 A portion of lithosphere, part of which has undergone erosion and has been isostatically compensated. The stress state within this portion of lithosphere conforms to the “constant horizontal stress” reference state (McGarr, 1988). Where $\rho_l$ = density of the lithosphere, $\rho_m$ = density of the mantle, $\sigma_h$ = horizontal stress in lithosphere of original thickness, $\sigma_h^*$ = horizontal stress in eroded and thinned portion of lithosphere, $w$ = elevation in base of eroded lithosphere. (After Engelder, 1993).

In the uneroded portion of the lithosphere the average horizontal stress is denoted as $\sigma_h$. The key assumption in the model of McGarr (1988) is that the stress in this thicker portion of the lithosphere is unaffected by the change in thickness of the eroded lithosphere, that is to say a “constant horizontal stress” boundary condition is imposed. This is true if there is perfect isostatic compensation, that is if the lithosphere acts as though shear stresses on vertical planes can not be sustained. If the portion of lithosphere is long compared to its thickness (ratio of length to thickness greater than 6) the above assumption is reasonable for most relevant materials (Artyushkov, 1973).

Under the assumption that the horizontal forces must balance after erosion, the average horizontal stress in the thinner portion of the lithosphere $\sigma_h^*$ is now higher and it is shown (McGarr, 1988) that:

$$\sigma_h^* = \sigma_h \left( \frac{Z_1}{Z_1 - Z} \right) - \rho_l g Z \left( \frac{\rho_l}{\rho_m} \right) \left( \frac{Z_1 - Z}{2(Z_1 - Z)} \right)$$

(3.10)

Where $g$ is the acceleration due to gravity.

Using this equation, McGarr (1988) calculated changes in horizontal stress due to erosion (both surface erosion and thermal erosion of the base of the lithosphere) and
isostatic response. It was shown that changes in horizontal stress relative to vertical stress are much less using the above approach than using the uniaxial strain approach and it is argued (McGarr, 1988; Engelder, 1993) that the constant horizontal stress reference state more closely fits stress data from a broad range of geological environments.

So far however, tectonic stresses have not been considered. There is a whole range of data (much of which is summarised by the world stress map) which points to the existence of tectonic stress in the lithosphere in a broad range of geological environments. It seems therefore that the simple models described so far, will be unlikely to account for the majority of observations of stress. For this reason, the rest of section 3.3 mostly looks at the role of tectonic forces in the origin of lithospheric stress.

3.3.5 The Origin of Tectonic Stress

Sources of horizontal stress in the lithosphere associated with tectonic processes are discussed in this section. These stresses act in addition to the stresses discussed so far (sections 3.3.3 and 3.3.4). The forces which act on the plates to cause and resist their motion are generally called plate boundary forces. The lithosphere scale stresses within the plates reflect the “pushes” and “pulls” of these plate boundary forces.

The plate boundary forces suggested by Forsyth and Uyeda (1975) are briefly described below and are illustrated in Figure 3.6. Bott and Kusznir, (1984) have suggested a further set of sources of stress which they refer to as non-renewable sources of stress. These are associated with tectonic processes but are not due directly to the forces that drive the plates. They propose that sources of tectonic stress can be divided into renewable sources and non-renewable sources. This division is adopted here, and the terms renewable and non-renewable are defined below.
3.3.5.1 Renewable Sources of Stress

The sources of stress discussed in this section are renewable in the sense that there is a dynamic equilibrium between the application of the stress and its dissipation by tectonic strain. Throughout this equilibrium the lithosphere acts as a reservoir of strain energy.

*Ridge-Push (RP)*

These are the body forces developed as a consequence of gravitational loading of the evolving oceanic plates, arising from the systematic increase in lithospheric thickness as the plates diverge from the mid ocean ridges and cool. The ridge push force can be thought of as a distributed force which accumulates from a minimum at the ridge crest (where there is zero lithospheric thickness) to a maximum in lithosphere of approximate age 80 Ma (Parsons and Ritcher, 1980; Richardson, 1992). It is thought that mid-ocean ridges are in approximate isostatic equilibrium as indicated by the lack of an associated free air gravity anomaly (Forsyth and Uyeda, 1975). Material of anomalously low density is thought to underlie the ocean ridges and isostatically support their elevation (Bott, 1993). The excess potential energy at the mid-ocean ridge structure is dissipated through lateral spreading of the oceanic lithosphere. The ridge-push force helps to push plates apart and causes lateral compression within the adjacent ocean plates with a vertically averaged stress through the lithosphere of 20-30 MPa (Bott and Kusznir, 1984).

Figure 3.6 Some possible plate boundary forces. (After Engelder, 1993; Adapted from Forsyth and Uyeda, 1975).
Slab-Pull (SP) and Slab Resistance (SR)

The slab-pull force acts on the subducting plate at a convergent plate boundary and results from the higher density of the cooler material of the sinking slab relative to the density of the asthenosphere into which it sinks (Bott and Kusznir, 1984). The body force acting on the descending part of the plate is transmitted to the rest of the plate, which is pulled towards the trench. The forces due to the negative density contrast are resisted by a viscous drag that develops between the slab and the surrounding asthenosphere. The viscous resistive force is thought to be proportional to the velocity of the plate (Forsyth and Uyeda, 1975). Slab-pull could be an important plate driving force depending on the degree of resistance to subduction.

Subduction Suction (SU)

This is a “less obvious” force, first recognised by Elsasser (1971), which affects the overriding plate at a subduction zone “pulling” it towards the trench (Bott and Kusznir, 1984). This force arises from lack of support (Bott and Kusznir, 1984) and is therefore probably not actually a suction force but a gravitational body force acting to reduce excess potential energy in the over-riding plate, i.e. the over-riding plate is collapsing under its own weight (see section 3.3.4.1). The stress induced in this situation has been estimated by Bott and Kusznir (1984) to be a tension of around 20 MPa but could presumably be higher if the unsupported lithosphere were thicker.

Drag Forces: Mantle Drag (MD) and Continental Drag (CD)

The mantle drag forces are defined as those which act on the base of a moving plate if the velocity of the plate is different from that of the underlying mantle (Forsyth and Uyeda, 1975). If the convecting mantle is moving faster and in a similar direction to the plate, the drag force acts as a driving force. If the mantle is moving more slowly than the plate or in the opposite direction, the drag force acts as a resistive force. Basal drag affects both oceanic and continental plates. However, it is likely that old cratonic blocks with very thick lithosphere, or a “lithosphere keel” will be more affected by the drag forces. Observations of absolute velocities of the plates has shown that many of the continental plates are the slowest moving, indicating that the lithospheric keel does allow more viscous coupling between the mantle and the plate and that the drag force is
generally resistive (Engelder, 1993). Uncertainty in values such as the viscosity of the asthenosphere makes calculations of mantle drag forces difficult. Bott and Kusznir (1984) give a maximum value of driving or resistive stress induced in the lithosphere by a large underlying convecting mantle cell of 40 MPa.

Resistive Forces: Transform Fault Resistance (TR), Collisional Resistance (CR) and others.
A number of other resistant forces can be produced in the vicinity of plate boundaries, including (i) resistance to the sinking tongue of lithosphere (mentioned above) and to its down-bending in subduction zones, (ii) resistance to overriding plate motion in the vicinity of convergent plate margins, (iii) a small amount of ridge resistance at ocean ridge crests, (iv) frictional resistance along transform faults (TR), and (v) collisional resistance in mountain belts such as the Alpine-Himalayan front (Bott and Kusznir, 1984). The stresses associated with these resistive forces are not easy to estimate although attempts have been made in some cases, for example estimates of shear stress at the Himalayan front have been made by England and Molnar (1991) (see section 3.3.9).

3.3.5.2 Non-renewable Sources of Stress
The sources of non-renewable stress act on one part of the lithosphere over a time scale that is relatively short compared to the time scale over which the plate as a whole evolves. This idea is best illustrated by the examples given below. These non-renewable sources can be quite significant in magnitude but because they are generally relieved in relatively short periods of geological time (Bott and Kusznir, 1984) they are often neglected.

Bending Stresses
Bending stresses arise from bending of the lithosphere at subduction zones and from uncompensated loading of the lithosphere by sediments or ice sheets. The magnitude of the bending stress, which depends on the degree and geometry of loading or down-bending and the elastic thickness of the lithosphere, can be very high (up to 500 MPa stress difference beneath a major delta, and up to 1000 MPa at subduction zones) but is likely to be rapidly relieved (Bott and Kusznir, 1984).
Membrane Stresses

Membrane stresses are caused by changes in the radius of curvature of a plate as it changes latitude. Again, it is thought that these stresses can potentially be quite large (a maximum of around 100 MPa) but are non-renewable and therefore likely to be short lived (Bott and Kusznir, 1984).

Thermal Stresses

Thermal stresses are caused by temperature changes within the lithosphere and are likely to occur in the oceanic crust as it initially cools on formation and later heats up during subduction. Tensile stresses induced by cooling are thought to be dissipated by brittle fracture and creep before reaching high values (Kusznir, 1991) although stress magnitudes of up to 500 MPa are estimated by Bott and Kusznir (1984).

Table 3.1 is adapted from Bott and Kusznir (1984) and summarises the main sources of renewable and non-renewable stress. It also lists whether the stress source induces compressive or tensional stresses, i.e. whether it “pushes” or “pulls”.

<table>
<thead>
<tr>
<th>Source of Stress</th>
<th>Renewable or Non-renewable</th>
<th>Compression or Tension</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ridge Push</td>
<td>Renewable</td>
<td>Compression</td>
</tr>
<tr>
<td>Slab-pull</td>
<td>Renewable</td>
<td>Tension (normally)</td>
</tr>
<tr>
<td>Subduction Suction</td>
<td>Renewable</td>
<td>Tension (normally)</td>
</tr>
<tr>
<td>Mantle Drag</td>
<td>Renewable</td>
<td>Both</td>
</tr>
<tr>
<td>Lithosphere Bending (due to loads)</td>
<td>Non-renewable</td>
<td>Both</td>
</tr>
<tr>
<td>Lithosphere Bending due to Subduction</td>
<td>Non-renewable but continually generated</td>
<td>Both</td>
</tr>
<tr>
<td>Thermal Effects</td>
<td>Non-renewable</td>
<td>Both</td>
</tr>
<tr>
<td>Membrane Effects</td>
<td>Non-renewable</td>
<td>Both</td>
</tr>
</tbody>
</table>

Table 3.1 Sources of tectonic stress. Adapted from Bott and Kusznir (1984).
3.3.5.3 The State of Stress in the Lithosphere (Tectonic Setting and Stress Regime).

If the system of driving and resisting forces acting on a plate is known, then the stress distributions within the plate can be speculated upon. Some simple cases have been summarised (Bott and Kusznir, 1984) and are shown in Figure 3.7.

![Figure 3.7 Examples of simple stress systems within lithospheric plates caused by plate boundary forces. (After Bott and Kusznir, 1984)](image)

1. A plate which has ocean ridges on opposite sides should be in compression throughout as a result of ridge push forces acting on opposite sides (Figure 3.7a). An example of such a system would be the present day African plate.

2. A plate which has an ocean ridge on one side and a subduction zone on the opposite side would be expected to show a stress system which grades from compression at the ridge to tension adjacent to the subduction zone assuming the driving forces are mainly balanced by mantle drag (Figure 3.7b and 3.7c). An example of this configuration is the Pacific plate with ridge push on the western side and slab pull on the eastern side.

3. If subduction occurred on opposite sides of an essentially continental plate, then the trench suction force could cause the whole of the plate to be subjected to lateral
tension (Figure 3.7d). No such configuration exists at present, but it is thought that the supercontinent Pangaea may have broken up in the Early Mesozoic partly as a result of stresses induced by this type of plate boundary force.

The simple stress systems described above ignore many of the possible complications such as irregularities in the shapes of the boundaries and lateral density or topography contrasts in the plates. There are many parts of the world which clearly do not conform to the simple models illustrated here, e.g. the East African Rifts, the western coast of South America, the Tibetan Plateau, the German, Belgian and Dutch Rhine and Roer Graben system.

3.3.6 Local Influences on the Stress Field

The effects of geological features such as faults, folds, intruded bodies and salt diapirs in the crust, impart an inhomogeneous nature which causes local variations in the stress field. Such local variations can easily overprint and disguise the regional pattern of stress, so that measurements made on a small (i.e. local) scale may give a misleading picture. Some examples of how both stress magnitudes and orientations can vary on a local scale are presented below.

3.3.6.1 Local Effects on Vertical Stress.

Vertical stress is usually assumed in stress studies to be a principal stress and to be equal to the overburden load. It is clear that the effect of topography will change the vertical stress from being a principal stress; for example at the foot of a mountain (a small distance below surface to avoid the free surface effect) the stress due to the weight of the mountain would be the biggest stress and therefore a principal stress, however it would not be coming from directly above, but from the side. In this case, the vertical stress would be equal to the weight of the overburden plus some component from the weight of the mountain.

Similar to the effect of topography is the effect of compressional tectonic forces which are acting out of the horizontal plane. Such forces can arise if the tectonic compression is not being applied uniformly to both ends of the piece of crust in question, for example,
if it were being transmitted partly from below. Also if non-vertical faults are present, the principal stresses can re-orientate themselves such that they are parallel and perpendicular to the fault plane (see section 3.3.6.2), in which case the vertical stress will no longer be a principal stress.

Even in the absence of topography, differing mechanical properties of different layers of rock, coupled with some form of structure deviating from the simple “layer cake” model can give rise to “stress shielding”. A relatively weak rock can be shielded from the overburden by a structure consisting of relatively strong rock. Figure 3.8 shows a simple case where rock unit C could be shielded by unit B. It is easy to imagine how such structures could occur as a result of either deformation (faulting and folding) or as sedimentary features (large channel fill sands etc.).

![Figure 3.8 Vertical stress shielding. Rock unit C could be shielded from vertical stress if rock units B and D have considerably higher stiffness than unit C. (After Katahara, 1996).](image)

Where vertical stress magnitudes vary, so will the magnitudes of the horizontal stresses generated by them (see section 3.4).

3.3.6.2 The Stress Field Around Faults, Cracks and Other Inclusions.
Cracks can act as free surfaces such that they support no shear stress. As a result, the stress field around such a crack must be oriented so that the principal stresses are
aligned parallel and perpendicular to the free surface. Faults can also act as free surfaces or soft inclusions to varying degrees depending on the strength of the fault material. A review and explanation of stress fields around joints is given in Engelder (1993). The influence of faults on the orientation of stresses has been widely noted (Bell *et al.*, 1992; Cowgill 1995).

3.3.6.3 The Stress Field Around Domal Structures.
The effect of domal structures on the stress orientations has been provided by a study of the halokinetically induced Ekofisk dome within the Central Graben of the North Sea (Teufel and Farrell, 1990). The orientations of the local stress field mapped in the vicinity of the dome suggests that the maximum horizontal stress is oriented perpendicular to the structure contours around the dome, and that a local stress field has developed which overprints the regional stress field seen throughout much of Western Europe.

Measured principal stress orientations, derived from borehole breakouts, around a salt dome in northern Germany have been compared with calculated principal stress directions from a finite element model by Brereton and Müller (1991). The model and the measured orientations are in good agreement (Figure 3.9) showing the principal stresses to realign around the edge of the dome.
Rotations of stress orientation have been noted in western Canada in sediments around a Precambrian basement high. It is suggested that the rotation is due to a contrast in elastic properties between the basement rocks and the sediments (Bell and Lloyd, 1989).

Another, somewhat more tangible example of local stress rotations is seen around volcanoes, and is mapped out by the radial pattern of dykes and fissures seen on the flanks of volcanoes. An example is given by Engelder (1993) who also notes that in some cases the dykes can be traced away from the vicinity of the volcano, and seen to rotate into the direction of the regional orientation of $\sigma_H$.

3.3.7 The World Stress Map.

Assuming that a plate behaves as a rigid body, the absolute velocity of the plate is determined by the resultant of all the forces acting on the plate. The stress field within the plate however, will be a function of proximity to the area of application of a particular force. The relative importance of various plate driving forces is therefore best addressed by examining the patterns of stress on a plate wide or global scale.
The World Stress Map (WSM) project is a global co-operative effort to compile and interpret data on the orientation and relative magnitudes of the contemporary in situ tectonic stress field. More than 7300 in situ stress orientations have been compiled as part of the WSM and of these over 4400 are considered reliable tectonic stress indicators recording horizontal stress orientations to within $\pm 25^\circ$ (Figure 3.10) (Zoback, 1992)

The stress data come from a variety of stress determination techniques (discussed in section 4.2) some of which are able to give orientation of maximum and minimum horizontal stresses and relative stress magnitude (e.g. earthquake focal mechanisms, hydraulic fracturing tests) and some are only able to give the horizontal stress orientations (e.g. borehole breakouts).

Some of the important results to come out of the WSM project have been discussed by Zoback (1992) and are summarised below:

- Remarkably good correlation is observed between stress orientations deduced from in situ stress measurements and geological observations made in the upper 1-2 km, well bore breakouts extending to 4-5 km depth and earthquake focal mechanisms to depths of around 20 km. This indicates a relatively uniform stress field with depth throughout the brittle part of the crust.

- Most intraplate stress regimes are either compressional or strike slip, with intraplate extensional faulting mostly being restricted to areas of elevated topography.

- Regionally uniform stress orientations and relative magnitudes are observed. This observation permits the definition of broad scale regional stress patterns often exceeding 20-200 times the approximately 20-25 km thickness of the upper brittle lithosphere.
Figure 3.10 The World Stress Map showing the orientations of maximum horizontal stress derived from various methods (see Key). This map was downloaded from the home page of the World Stress Map Project: http://www-gpi.physik.uni-karlsruhe.de/pub/bsm/index.html
Some of the conclusions which can be drawn from the above observations are discussed in section 3.3.8

3.3.8 Plate Motions, Driving Forces and Stress Fields.

Present day plate motions have been determined in both a relative framework, the motion of one plate relative to another, and an absolute framework, the motion of a plate relative to a fixed point (Minster and Jordan, 1978). To specify an “absolute” reference frame, an origin in angular velocity space must be chosen. A particular frame of interest in discussions of plate dynamics is one that is fixed with respect to the position of the deep mantle which is assumed to be rigid or at least have inertial motions much slower than the motions of the plates; such a frame of reference is provided by hot spots (Minster and Jordan, 1978).

There are several simple relationships between plate configuration and absolute plate velocity which have a bearing on the relative importance of plate driving forces. It is assumed that those plates subjected to larger body forces will move relatively faster and those experiencing larger resistive forces will move relatively slower and indeed, plate velocity varies directly with the percentage of plate boundary connected with subducting slab and varies inversely with the area of continental lithosphere (Engelder, 1993).

Torque poles are calculated as the integral of the product of the plate driving force and radius (distance between relevant portion of plate boundary and pole of rotation of the plate) over the length of the plate boundary (Richardson, 1992). The direction of the torque pole, when projected onto a map, can be thought of simply as the direction in which the plate driving force acts. Torque poles have been calculated for a variety of possible forces acting on the plates including ridge push, slab pull and collisional resistance (Richardson, 1992; Richardson and Reding, 1991). They have been compared with the directions of absolute plate motion and the patterns of orientation of the maximum horizontal stress ($\sigma_{H}$) as determined by the WSM project.
There is a strong correlation between direction of ridge push force and absolute plate motion for the North American, South American, Cocos and Eurasian plates (Figure 3.11); directions of slab pull forces correlate well with the absolute motion of the Pacific, Nazca and Cocos plates; collisional resistance torque poles correlate well with the absolute motion of the Eurasian plate (Richardson, 1992). These correlations are thought to support the idea that the absolute reference frame for plate motions is determined by the plates themselves rather than by the pattern of convection in the sublithospheric mantle (Richardson, 1992).

Figure 3.11 Comparison of ridge torque (solid arrows) and absolute velocities (dashed arrows) of Minster and Jordan, 1978. (After Richardson, 1992).

Torque poles for various forces and absolute plate motion directions have also been compared to long wavelength features of the global intraplate stress field as determined by the WSM project (Richardson, 1992; Richardson and Reding, 1991). The notion that plate boundary forces, in most cases, dominate the stress field arises from the fact that, in several cases, $\sigma_h$ is parallel to absolute plate velocity (Engelder, 1993). Ridge torque directions agree well with the orientations of maximum horizontal stresses for stable North America, western Europe and South America (Richardson, 1992). Collisional resistance forces (Richardson, 1992) or a combination of ridge push and collisional
resistance (Brereton and Müller, 1991) can also be used to explain the alignment of stresses in western Europe. Neither the North American, Eurasian or South American plate is attached to any significant length of subducted slab. The stress field in the Indo-Australian plate cannot be explained simply in terms of one plate boundary force.

Where the absolute direction of motion of the plate is parallel to the direction of $\sigma_H$, basal drag can always be evoked as the plate driving force. In the cases mentioned above, where the ridge push or collisional resistance torque pole is also parallel to the $\sigma_H$ direction, it is not possible to determine which force is responsible for the stress field from stress orientation data alone (Zoback, 1992; Zoback and Magee, 1992; Richardson, 1992; Richardson and Reding, 1991). There are however several lines of evidence which support the ridge push force over the basal drag force (Richardson, 1992):

(i) The ridge push forces would always predict compressional intraplate stresses away from the ridges, which is what is observed in the vast majority of cases in the WSM project, whereas basal drag might produce compressional or extensional stresses depending on the absolute motion of the plate and other boundary conditions.

(ii) The absolute motion of the Eurasian plate is very small. It is not likely that such a plate would be dominated by basal drag forces when other, faster moving plates are not.

(iii) Models using a uniform basal drag as the plate driving force predict large increases in the magnitude of the force across plates with long wavelength uniform stress fields. For the North American plate, a stress amplification ratio of the order of 50-100 from the east to the west of the plate is predicted when the western boundary is fixed. Absolute stress magnitude data is rare, however, there is no evidence in the relative stress magnitude data from tectonic regimes of a stress amplification by a factor of 50.

3.3.9 The Rheology of the Lithosphere.

In addition to the above discussion about plate driving forces and their relative importance, there has also been debate concerning how stress is transmitted through the
lithosphere. In particular, in continental lithosphere, which part of the lithosphere is responsible for transmitting most of the stress.

In contrast to the oceanic portions of the Earth's lithosphere, which move as essentially rigid bodies, the continents exhibit distributed deformation over horizontal length scales of hundreds to thousands of kilometers (England and McKenzie, 1982; England and Molnar, 1991). As a result of such observations, it has been suggested that the rheology of the lithosphere is best described as a viscous sheet (England and McKenzie, 1982). This viscous sheet model uses a vertically averaged rheology for the lithosphere incorporating brittle faulting in the upper crust and a combination of creep and brittle failure, depending on temperature and strain rate, in the lower lithosphere. The observations of strain in zones of active deformation are best fitted by results from the above model in which the rheology of the lithosphere is controlled by creep (England and Molnar, 1991).

The relationship of tectonic style to surface elevation in the Andes and Tibet (both show compressional features below about 4 km and extensional features above 4 km) has been used to calculate a force per unit length of $3-6 \times 10^{12}$ Newtons per Metre ($N.m^{-1}$) needed to deform the lithosphere of these regions (England and Molnar, 1991). Estimates of the level of shear stress on active thrust faults which bound the Himalayan Mountains have been derived from temperature measurements around the faults, and imply that the seismically active portions of these faults can contribute only $0.5-1 \times 10^{12} \text{ N.m}^{-1}$ (England and Molnar, 1991). It is argued that the upper, brittle part of the lithosphere does not control the deformation of the lithosphere as a whole, but passively follows a continuously deforming, stronger substrate which probably consists mainly of the uppermost mantle (England and Molnar, 1991). This somewhat counter-intuitive notion remains controversial. The only direct way of testing it's validity is to actually access the deeper parts of the lithosphere. What follows below, outlines the evidence for the more conventional theory that stress is transmitted primarily in the brittle upper crust, and describes how attempts have been made to verify this by drilling a very deep borehole.
Laboratory friction studies, combined with simple faulting theory (as well as extrapolation of in situ stress measurements from the upper 3 km of the crust) imply that if pore pressure is approximately hydrostatic at mid crustal depth, crustal strength is appreciable (hundreds of MPa) and would markedly constrain the nature of lithospheric deformation (Zoback et al., 1993). This hypothesis implies that in-situ differential stresses (and thus shear stresses) are limited by the frictional strength of well oriented and pre-existing faults, and attempts have been made to measure the stress magnitude in order to test this theory down to mid-crustal depths in the KTB (Kontinentales Tiefbohrprogramm Bundesrepublic Deutchland) deep borehole site in southern Germany.

To consider the strength of the upper crust at the KTB site in the context of how the lithosphere transmits plate driving forces, a model must be constructed which incorporates the ductile creep strength of the lower crust and upper mantle. This creep strength depends on the geothermal gradient (estimated from surface and down hole heat flow measurements), estimated strain rates, and rheological values of the rocks which constitute the lower crust and upper mantle (determined from laboratory experiments).

Such a model is used in the KTB region (Zoback et al., 1993) for two possible lower crustal lithologies, Adirondack and Piikwitonei granulites, which have very low and very high ductile strengths respectively at the strain rate and temperature estimated to be applicable to the lower crust of this region. The average strain rate is taken as $10^{-16} \text{ s}^{-1}$ which is assumed to be the upper limit for plate interiors. The lower crust temperatures are based on the observed geotherm and relatively high heat flow (74 mW.m$^{-2}$) at the KTB site (Zoback et al., 1993). The upper mantle has been assumed to have the rheology of dry dunite (essentially olivine) which, because of the high temperatures in the upper mantle, has very low strength. The upper crust however is shown to be able to support between $2 \times 10^{12} \text{ N.m}^{-1}$ in the case of the Adirondack granulite (which yields a brittle ductile transition at around 12 km depth for the given geothermal gradient) and $5 \times 10^{12} \text{ N.m}^{-1}$ in the case of Piikwitonei granulite (which yields a brittle ductile transition at around 19 km depth for the given geothermal gradient). Figure 3.12 shows the
differential stress and cumulative force profiles respectively for these two rock types under conditions at the KTB site. The cumulative force profile shows what magnitude of force per unit length the lithosphere is capable of transmitting and can therefore be thought of as a cumulative strength profile.

![Diagram of stress and force profiles](image)

Figure 3.12 (a) Theoretical maximum differential stress ($\sigma_1-\sigma_3$) profiles for conditions similar to those in the region of the KTB drill site. Data points shown as o indicate the difference between $\sigma_H$ and $\sigma_b$ as determined from hydraulic fracturing data to 3 km depth in the pilot hole. The data points shown as horizontal lines indicate a generalisation of the values from the breakout and tensile fracture analysis. A strike-slip faulting stress regime with hydrostatic pore pressure and a coefficient of friction ($\mu$) of 0.7 was used to calculate the strength of the upper crust. (b) cumulative crustal force corresponding to the strength profiles shown in (a). (After Zoback et al., 1993).

As the available tectonic force per unit length from ridge push ($2-5 \times 10^{12} \text{ N.m}^{-1}$) is comparable in magnitude to the cumulative lithospheric force which the lithosphere is estimated to be able to transmit ($2-5 \times 10^{12} \text{ N.m}^{-1}$), the lithosphere in this part of western Europe is assumed to be undergoing steady state failure equilibrium (distributed failure on numerous small faults throughout the region giving rise to steady state deformation of the lithosphere over the whole region) reflected by wide spread seismicity, with the force transmission through the lithosphere being largely in the upper 10-20 km of the crust (Zoback et al., 1993).

Estimates of the magnitudes of in-situ stresses made in the KTB borehole to depths of 6 km indicate a high strength upper crust, in which the state of stress is in equilibrium with its frictional strength (Zoback et al., 1993). More recently, fluid induced seismicity tests have been conducted to depths of 9 km in the KTB borehole. The induced seismic
events were either strike-slip, reverse, or a component of both and showed the orientation of maximum compression to be oriented close to NNW, parallel to the direction of maximum horizontal stress observed in the borehole (Zoback and Harjes, 1996). It is estimated that many of the seismic events were induced by pore pressure increases of less than 1%, indicating that these fault planes are critically stressed which supports the hypothesis that Byerlee’s law (i.e. in-situ differential stresses are limited by the frictional strength of well oriented, pre-existing faults) is valid to these depths (Zoback and Harjes, 1996) and that a considerable amount of tectonic force can be transmitted through the brittle upper crust.

The relationship between lithospheric stress and depth is primarily controlled by rheology, which depends on the geothermal gradient, crustal composition and thickness. Kusznir (1991) shows how a variety of different stress depth relationships can occur, including both strong upper crust and strong upper mantle, by varying these parameters and using different strain rates. The sensitivity of stress predictions in the crust based on laboratory flow laws is highlighted (England and Molnar, 1991) by the statement “Even in the absence of uncertainty in material properties determined in the laboratory, uncertainty in temperature and lithology at any given depth in the lithosphere is sufficiently large to make a prediction of deviatoric stress from laboratory flow laws uncertain by an order of magnitude at least”.

It appears then, that although stress-depth relationships can be estimated using the above methods proposed by Zoback et al. (1993) in a few areas (where due to the presence of a deep borehole, at least some of the parameters can be determined to at least some of the depth of relevance) in the vast majority of the Earth’s lithosphere there is likely to be a considerable degree of uncertainty.

3.4 Models of Stress Distribution with Depth in Sedimentary Basins
Models of stress within sedimentary basins, as opposed to within the lithosphere as a whole, have evolved because sedimentary basins occur in specific structural and tectonic settings within the lithosphere. Most of the work on stress in sedimentary basins has been oil industry led. As a consequence of the industrial interest, the models which have
given good empirical results have tended to dominate as have models which use parameters that are easily determined, from geophysical logs for example.

3.4.1 Elastic Models

The most simple elastic model, the uniaxial strain model, has been briefly described in section 3.3.3.2, and is the basis for most predictions of horizontal stress magnitude in the oil industry. In its simplest form, the conditions of zero lateral strain and the absence of tectonic stress, are thought to approximate a tectonically relaxed sedimentary basin (Engelder, 1993).

The first models of this type related the horizontal stress to the vertical stress, pore pressure and a ratio of horizontal to vertical effective stresses which was an empirically derived ratio (e.g. Mathews and Kelly, 1967). Later models used Poisson’s ratio to estimate the effective stress ratio, where Poisson’s ratio was either determined from sonic logs, or back calculated from previous fracture gradients (Eaton, 1969; Anderson et al., 1973; Daines 1982). A summary of most of these models is given by Breckels and van Eekelen (1981). Modifications to the simple uniaxial strain model have been made to account for transverse isotropy and horizontal tectonic stress (Daines, 1982; Thiercelin and Plumb, 1991) and have been outlined below. These models have also been discussed by Katahara (1996).

3.4.1.1 Isotropic and Transverse Isotropic Uniaxial Strain Model.

The isotropic uniaxial strain model is the most simple of the elastic models. It assumes that the rock mass is a semi-infinite poroelastic material subjected only to gravitational loading such that the vertical stress is a principal stress, as are the two horizontal stresses. It is shown (Thiercelin and Plumb, 1991) that the assumption of the uniaxial strain condition (there is no strain in the horizontal directions) reduces Hooke’s law for a poroelastic material to:

\[
\sigma_h - \alpha p = \left[ \frac{\nu}{1-\nu} \right] (\sigma_v - \alpha p)
\]  

(3.11)
Where \( \sigma_h \) is the total horizontal stress (equal in all directions in this case), \( \sigma_v \) is the total vertical stress (equal to the overburden in this case), \( p \) is the pore fluid pressure, \( v \) is the Poisson's ratio of the rock, \( \alpha \) is the poroelastic constant (often assumed to be equal or very close to 1).

This model predicts different stress magnitudes in different lithologies because of the dependence on Poisson's ratio.

The transverse isotropic model applies to horizontally layered or bedded rocks which are isotropic within the bedding plane, but not in the direction normal to this plane. Under uniaxial strain conditions, the horizontal stress is shown (Thiercelin and Plumb, 1991) to be:

\[
\sigma_h' - \alpha p = \frac{E}{E'} \left( \frac{\nu'}{1 - \nu'} \right) \left( \sigma_v - \alpha (1 - \xi) p \right)
\]  

(3.12)

Where \( E \) is the Young's modulus in the plane of isotropy, \( E' \) is the Young's modulus normal to the plane of isotropy, \( \nu \) is the Poisson's ratio in the plane of isotropy for stresses acting in the plane of isotropy, \( \nu' \) is the Poisson's ratio for stress acting normal to the plane of isotropy, \( \xi \) is a poroelastic constant. This model has been evaluated with real stress measurements and laboratory determined rock properties for approximately nine East Texas intervals (Thiercelin and Plumb, 1991) but because of the number of rock properties which must be determined in the laboratory from rock core corresponding to the rock tested in-situ, the use of this model is rare. With the assumption that \( \xi = 0 \) (the assumption used by Thiercelin and Plumb, 1991) the transverse isotropic model predicts that, if the rock is stiffer (in-situ) in the plane of isotropy than it is in the direction perpendicular to that plane, the horizontal stress will be higher than in the simple isotropic case.

In their evaluation, Thiercelin and Plumb (1991) show that the transverse isotropic model above yields a significantly better result than the simple uniaxial strain model in only one case out of nine, where the rock in question is a shale. They have assumed that \( \alpha = 1 \) and \( \xi = 0 \) but then show that realistic values of \( \alpha \) worsen the results of both models considerably, with predicted values of horizontal stress being much too low.
3.4.1.2 Elastic Models Incorporating Horizontal Strain.

Both of the models discussed in section 3.4.1.1 can be modified if the assumption of zero horizontal strain is relaxed. If a uniform (with respect to depth) amount of anisotropic tectonic strain is added Thiercelin and Plumb (1991) give equations for the isotropic elastic model:

\[ \sigma_h - \alpha \rho = \left[ \frac{\nu}{1-\nu} \right] (\sigma_v - \alpha \rho) + \frac{E}{1-\nu^2} e_h + \frac{E \nu}{1-\nu^2} e_H \]  

(3.13)

\[ \sigma_H - \alpha \rho = \left[ \frac{\nu}{1-\nu} \right] (\sigma_v - \alpha \rho) + \frac{E}{1-\nu^2} e_H + \frac{E \nu}{1-\nu^2} e_h \]  

(3.14)

Where \( \sigma_h \) is now the minimum total horizontal stress, \( \sigma_H \) is the maximum total horizontal stress, and \( e_h \) and \( e_H \) are the minimum and maximum principle horizontal strains, respectively.

These equations show a dependence of the stress on the Young's modulus such that the greater the Young's modulus the higher the stress. This model can therefore account for situations where the level of stress in sandstone beds (low \( \nu \) and high \( E \)) is higher than in shales (high \( \nu \) and low \( E \)) such as occurs in some of the Devonian rocks of the Appalachian Plateau (Evans et al., 1989; Plumb et al., 1991). However, it is not clear whether rocks in a sedimentary basin environment, such as that of the North Sea, would be likely to have undergone the degree of horizontal strain experienced by the Devonian rocks of the Appalachian Plateau.

For the case of transverse isotropy, the addition of vertically uniform, anisotropic horizontal strains causes the horizontal stresses to become:

\[ \sigma_h - \alpha \rho = \frac{E}{E'} \left[ \frac{\nu'}{1-\nu'} \right] (\sigma_v - \alpha (1-\xi) p) + \frac{E}{1-\nu^2} e_h + \frac{E \nu}{1-\nu^2} e_H \]  

(3.15)

\[ \sigma_H - \alpha \rho = \frac{E}{E'} \left[ \frac{\nu'}{1-\nu'} \right] (\sigma_v - \alpha (1-\xi) p) + \frac{E}{1-\nu^2} e_H + \frac{E \nu}{1-\nu^2} e_h \]  

(3.16)
Daines Method

The method which has become known as Daines method (Daines, 1982) is fairly widely used in the oil industry to predict the magnitude of $\sigma_h$ (Exlog, 1982). The model predicts that $\sigma_h$ is simply the value predicted by the isotropic uniaxial strain model (with the assumption that $\alpha=1$) with the addition of an empirically determined component of horizontal tectonic stress. The tectonic stress component is assumed to increase linearly with depth. $\sigma_h$ then becomes:

$$\sigma_h = \left[ \frac{\nu}{1-\nu} \right] (\sigma_v - p) + p + \sigma_t$$

(3.17)

Where $\sigma_t$ is the tectonic stress component.

During well drilling, the tectonic component (per unit of depth) is determined at the first leak-off test/fracture test (see section 4.6) made in compact formation. This is done by equating the leak-off pressure to the minimum horizontal stress magnitude, estimating the pore pressure from a variety of drilling parameters (known collectively as the drilling exponent - section 5.4.2) and using standard (previously empirically derived) Poisson's ratio values for the lithology in question.

Predictions of $\sigma_h$ at deeper levels in the well are then made for various lithologies using the standard values of Poisson's ratio, and the tectonic component extrapolated to the desired depth. One problem with this model is that, because different beds have different values of Young's modulus, it predicts different amounts of horizontal strain in different beds.

None of the elastic strain models takes account of processes such as cracking or creep (the effects of which are often seen in sedimentary basin rocks) by which stresses are relaxed irreversibly.

3.4.1.3 Poisson's Ratio

The parameter that is key to all forms of the uniaxial strain model is Poisson's ratio. Poisson's ratio, as applied to perfectly elastic materials, is an elastic constant, and is defined as the ratio of radial to longitudinal strains when a longitudinal stress is applied to the ends of a long stiff cylinder with unconfined sides. However, real rocks in the
earth are much more complex than the perfectly elastic rod in the definition of Poisson’s ratio for two main reasons:

1. Real rocks contain anisotropies such as aligned layering (bedding, cleavage etc.) as well as aligned systems of cracks, both of which impart anisotropic elastic properties, so that Poisson’s ratio will vary according to the direction in which it is measured. Poisson’s ratio can also vary according to the level of stress (including confining pressure) under which it is measured. This is discussed further (with respect to the results from the North Sea) in section 7.3.2.

2. Most real rocks, especially sediments, are porous, with a fluid which is usually aqueous, occupying the pore spaces. The rock is then a poroelastic material. When an external stress (total stress) is applied to such a rock it is carried both by the solid framework of mineral grains and the pore fluid, but each carry different amounts. If such a poroelastic rock is tested to measure it’s Poisson’s ratio, the value obtained will depend on whether the fluid is allowed to escape as the external load increases (drained condition) or if the system is sealed in which case the fluid cannot escape (undrained condition). In the experiment carried out under drained conditions, the Poisson’s ratio is dependent only on the elastic properties of the grain framework and the value of Poisson’s ratio obtained can be called the drained Poisson’s ratio $v_{drained}$. In the experiment carried out under undrained conditions, the value of Poisson’s ratio obtained is dependent both on the fluid and the elastic properties of the grain framework and can be called $v_{undrained}$. The property of the fluid that will affect $v_{undrained}$ is the compressibility. For example, if an undrained test is carried out on a rock with water in the pore spaces, the water, being largely incompressible, will mean that the rock as a whole will have to expand laterally to balance the stress. A gas in the pore spaces on the other hand, being much more compressible than water, will allow pore spaces within the rock to become smaller, in this way the stress can be balanced to a large extent by internal elastic strain, and lateral expansion of the rock as a whole is less. The ratio of lateral expansion to vertical compression (i.e. the Poisson’s ratio) of the gas filled rock is smaller than the water filled rock, therefore $v_{undrained}$ gas filled rock is less than $v_{undrained}$ water filled rock. It can also be assumed that if the gas has a high compressibility, then the drainage conditions do not matter, so that $v_{undrained}$ gas filled rock is approximately equal to $v_{drained}$ gas filled rock.
In an elastic isotropic material, Poisson’s ratio can be expressed in terms of the rigidity (G) and the incompressibility (K):

\[ \nu = \frac{3K - 2G}{6K + 2G} \]  

(3.18)

The P-wave (V_p) and S-wave (V_s) velocities of seismic and sonic waves passing through an elastic material depend on the rigidity, incompressibility and density (\( \rho \)) of the material. Seismic waves travel in isotropic solids with velocities given, for example by Thomsen (1996), as:

\[ V_p = \sqrt{\left(\frac{K + 4G/3}{\rho}\right)} \]  

(3.19)

\[ V_s = \sqrt{\frac{G}{\rho}} \]  

(3.20)

It can be shown from equations 3.18, 3.19 and 3.20 that:

\[ \nu = \left[\left(\frac{V_p}{V_s}\right)^2 - 2\right] \sqrt{\left(\frac{2V_p}{V_s}\right)^2 - 2} \]  

(3.21)

which expresses Poisson’s ratio in terms of the wave velocities, assuming the rock is elastic and isotropic. However, as real rocks have pores and cracks containing fluid, equation 3.21 must be qualified by consideration of the drainage conditions. Because seismic and sonic waves have frequencies which compress the rocks for only short periods of time, it is assumed (Thomsen, 1996) that the pore fluid does not have time to flow out of the rock during the passage of the seismic wave, and therefore the Poisson’s ratio measured by the wave speeds is representative of undrained conditions. The Poisson’s ratio measured by seismic waves is generally called the dynamic Poisson’s ratio.

When stresses such as tectonic and gravity induced stresses are applied to rocks in the Earth’s crust, the time scale over which they are applied is obviously much greater than that of a passing seismic wave. The pore fluid in this case does have time to flow out of the rock, if other conditions allow, and induce pore fluid flow. The Poisson’s ratio
applicable to these circumstances is the drained Poisson's ratio. The term static Poisson’s ratio is often applied to situations where the time scale is long, however a static condition could be drained or undrained.

For a gas filled rock, the undrained (dynamic) and drained Poisson’s ratio will be virtually the same (see above) so Poisson’s ratio derived from wave velocities in gas filled rocks can be used to predict the stresses in the uniaxial strain model.

For water filled rock, it is not immediately obvious whether the drained or undrained Poisson’s ratio be used in the uniaxial strain model. The uniaxial strain model relates the effective stresses to Poisson’s ratio. In a drained experiment the pore pressure has no effect on the Poisson’s ratio, so it is this Poisson’s ratio which relates the effective stresses. During the burial of sediments, under “normal” conditions, the formation pore fluid pressure increases as the hydrostatic head increases, however it is also free to drain away as the rock compacts. Under “normal” conditions then, it seems that the drained Poisson’s ratio is the suitable Poisson’s ratio to use in the uniaxial strain model. Rocks which contain pore pressures considerably higher than pressure due to hydrostatic head are not uncommon. In these cases the pore fluid is clearly not free to drain away and so it seems that the undrained Poisson’s ration would be appropriate for use in the uniaxial strain model.

3.4.2 Failure Models
Geological observations (e.g. Chapter 2) in sedimentary basins and laboratory measurements (e.g. Ayling, 1992) show that the assumption that the rocks behave elastically under crustal conditions is rarely true. The observation that most rocks contain faults and fractures on a variety of scales, clearly shows that they have experienced stress states which lie outside their elastic limits and leads to the suggestion that rock stress magnitudes are governed by the shear strength of the rocks. In a normal faulting regime, where $\sigma_1$ is the vertical stress, the magnitude of $\sigma_h$ will be limited by how much differential stress the rock can withstand i.e. its shear strength. If the rock is assumed to be in a state of incipient failure and the Mohr-Coulomb failure criterion is used, $\sigma_h$ is given (Perkins, 1976; Katahara, 1996) by:
Where $\phi$ is the internal friction angle and $C_u$ is the cohesive strength of the rock. Internal friction angles and cohesive strengths can be determined experimentally or can be obtained from empirical relationships with other more easily measured rock properties such as porosity and clay content (e.g. Plumb, 1994).

The Mohr-Coulomb criterion does not account for curved failure envelopes (i.e. the failure envelope described by equation 3.22 plots as a straight line on a Mohr-Coulomb diagram) and also predicts that failure, and therefore minimum stress magnitude, is independent of the intermediate stress. Katahara (1996) points out some limitations of this model by stating that in fact the intermediate stress can have a role in failure and also that real failure envelopes are curved.

If the rock has cracks of a favourable orientation (it is assumed that the pre-existing cracks have no cohesive strength) then the failure criterion in a normal fault stress regime becomes:

$$\sigma_h = \frac{1 - \sin \phi}{1 + \sin \phi} (\sigma_v - p) + p$$

(3.23)

The states of stress described by these equations can be represented on a Mohr diagram (Figure 3.13) where the normal stresses are effective stresses ($\sigma'$) and the shear stresses ($\tau$) are those acting on the plane whose normal makes an angle $\beta$ with $\sigma_1$. Because there is no cohesion across the failure surface, the effective stress law is the simple Terzaghi law (Thiercelin and Plumb, 1991):

$$\sigma' = \sigma - p$$

(3.24)

Failure models predict the magnitude of the minimum principal stress ($\sigma_3$) from a knowledge of the magnitude of the maximum principal stress ($\sigma_1$) or vice versa. Generally in sedimentary basins, the only stress magnitude which can be easily determined without direct stress measurement is the vertical stress. Therefore, when the friction angles and pore pressure are known, in normal faulting regimes, $\sigma_h$ can be estimated from a knowledge of the vertical stress. In reverse faulting regimes, $\sigma_h$ can be estimated from a knowledge of the vertical stress.
determined from a knowledge of the vertical stress. In strike slip regimes, where $\sigma_1$ and $\sigma_2$ are both horizontal stresses, the failure model is of limited use without further assumptions, or some stress measurements.

The main restriction of this model is that, in order to apply it confidently, some independent observations about the present stress regime are needed. Such observations could be from stress measurements or from active faulting. Stress measurements would generally be restricted to a small area and a small depth range, so application of the failure model would have to assume that the stress regime is the same in the area in which the measurements were taken and the area in which the model is to be applied. Observations of active faulting tend to be made over a wider area, although focal mechanisms are generally from depths below sedimentary basins. Inferring a particular stress regime at one depth from observations at another may be risky. One reason why this is so is because horizontal and vertical effective stresses change by different amounts.

Figure 3.13 Mohr-Coulomb failure envelopes for the case of a rock with no cohesion (solid straight line) and the case where the rock has a cohesion, $C_u$, (dashed straight line). The states of stress (for a normal faulting stress regime) corresponding to failure of the above rocks are also shown.
during pore pressure change (e.g. Teufel et al., 1991) thus pore pressure changes can easily change the stress regime.

The Mohr-Coulomb failure criterion has been evaluated using data from hydraulic fracturing stress measurements and laboratory determined rock properties by Thiercelin and Plumb (1991), using the same rocks and stress measurements as the evaluation of the elastic models (3.4.1). The data for the evaluation come from east Texas where recent fault movements have occurred and the formations show extensional and shear fractures indicating a normal faulting regime. Thiercelin and Plumb (1991) concluded that the failure model gives good predictions of $\sigma_h$, and also predicts the stress contrast between sandstone and shale (with shales showing higher values of stress). The best predictions are made by using internal friction angles obtained by measuring peak strength values under non zero confining pressure as opposed to zero confining pressure.

In the study made by Thiercelin and Plumb, 1991, both the uniaxial strain model (particularly incorporating transverse isotropy) and the Mohr-Coulomb failure model both give reasonably good predictions of the stress state as measured by hydraulic fracturing. This has been found in a number of other studies which have been summarised by Katahara (1996) who suggests that the reason for this agreement between two fundamentally different models is the empirical equivalence of $\sin(\phi)$ and $1-2\nu$. Taking typical values for these two parameters: $\nu = 0.25$, and $\phi = 30^\circ$, both $\sin(\phi)$ and $1-2\nu$ give a value of 0.5 and so would both predict the same horizontal stress magnitude for a given vertical stress.

Lithologic stress contrasts, with higher values of $\sigma_h$ in shales than sands, are also predicted by both models. This means that shales are weaker and more deformable than sands in that they can support less shear stress (lower $\phi$) and expand laterally (if unconfined) more than sands in response to a given vertical stress and thus have a higher $\nu$. 

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A particular problem with the failure model, and the reason perhaps that it is not widely used in the oil industry, is a lack of knowledge of the friction angles and how to predict them from logs.

3.5 Conclusions

This chapter has introduced some of the ideas and theories regarding the nature of stress in the lithosphere, its origins and distribution. In particular, models of stress distribution with depth, which have been proposed for other sedimentary basins around the world, have been presented in detail in this chapter.

It is clear that ideas on the nature of stress in the crust are wide ranging and there are many unresolved issues. A good source of further discussion on these topics is the 1991 Royal Society publication: Tectonic Stress in the Lithosphere, the papers of which were first published in Philosophical Transactions of the Royal Society of London, series A, volume 337, pages 1-194.

It is highlighted at this point, in view of the above discussions, that there is clearly a need to actually make stress measurements in the crust in an attempt to resolve some of the issues discussed. The next chapter describes crustal stress measurement techniques. In particular, the hydro-frac method is examined, and the leak-off test (from which the bulk of the data in this study is derived) is assessed as a potential stress determination technique.
Chapter Four - Methods of Crustal Stress Measurement

4.1 Introduction
This chapter describes some common crustal stress measurement techniques. Methods which are not used in this study are described only briefly. The majority of this chapter describes in detail the type of borehole measurements which have been used in this study. In particular the hydraulic fracturing (hydro-frac) technique is discussed. Hydro-frac data has been used in this study to compare leak-off pressures to hydro-frac derived stress estimates. Also, the theoretical framework which has been established to describe the processes which occur during hydraulic fracturing has been used to investigate leak-off pressures in terms of crustal stress. Particular emphasis is placed on the processes involved in hydraulic fracturing breakdown. Oilwell drilling procedures, the leak-off test and previous interpretations of the leak-off pressure in terms of crustal stress are discussed. Theoretical limits are placed on the range of crustal stress magnitudes which the leak-off pressure can represent. Leak-off test pressure/volume plots are examined to determine crack behavior during the leak-off test. Density log integration and other methods used to calculate the vertical stress magnitude in boreholes are also described.

4.2 General Crustal Stress Measurement Techniques
In most cases the term "stress measurement" should be replaced by "stress estimation" or "stress inferred by measurement". Stress itself is very difficult to measure directly as will be seen below, in most cases it is the strain which is measured and the stress is then inferred from this through some assumption about the stress/strain relationship.

There are a wide variety of stress measurement techniques in use, both by industry, particularly the mining and underground excavation industries, the geothermal and oil industry, and by research workers. There is also a wide range of scales at which stresses are measured, from the rock core scale of a few centimeters, through the borehole methods which sample scales of the order of several meters, up to the earthquake focal mechanism methods which sample stresses up to perhaps the thickness of the crust. There is some debate as to how the stresses measured at different scales are related (Harper and Szymanski, 1991).
The stress measurement techniques described here have been divided into indirect methods, borehole methods and direct methods.

4.2.1 Indirect Methods

*Geological Indicators* (these yield stress orientations and usually relative magnitudes):

Patterns of folding, faulting and jointing are the surface manifestation of the varying magnitudes and orientations of tectonic stress throughout geological time. A well documented example of a geological indicator of the present day tectonic stress regime is the alignment of volcanoes and feeder vents which propagate parallel to the maximum principal far field stress (Nakamura, 1977; Zoback *et al.*, 1989). Neotectonic joint patterns have also been used to interpret the present day stress field (Hancock, 1991).

Care must be taken when interpreting surface geological features in terms of the present day stress field, as they may reflect the influence of crustal weaknesses inherited from older geological events. Only late neotectonic features can be confidently used to interpret the present in-situ stress field.

*Earthquake Focal Mechanisms* (these yield stress orientations and relative magnitudes):

Contemporary stress orientations and relative stress magnitudes deeper within the crust can be determined from the study of earthquake focal mechanisms. 54% of the data points compiled by the World Stress Map project (see section 3.3.7) were derived from earthquake focal mechanism studies. When seismic waves from an earthquake are recorded by a suitably distributed network of seismic stations, the distribution of first motions can yield information on the orientation (azimuth and dip) of the nodal planes, one of which is the fault plane, and the nature of the movement, (normal, strike slip, or reverse) which puts limits on the orientation of the stress direction, and reveals the relative magnitude of the stresses.

A major problem with this method is that it is not known whether or not crustal earthquakes occur on pre-existing planes of weakness, or newly formed fractures. The orientation of the principal stress axes relative to the fault plane can therefore not be confidently assumed and thus reliability can not be placed on single measurements.
However, when a large number of such measurements are analysed, the average is likely to provide information which accurately measures the contemporary stress field.

Due to the relatively low level of seismicity in the UK and the North Sea, it is rare for seismic events to be recorded in detail in these areas. Seismicity in and around the North Sea is discussed in section 2.5.

4.2.2 Borehole Methods

*Borehole Breakouts* (these yield stress orientations)

Borehole breakouts, first reported by Leman (1964) from horizontal boreholes in South African gold mines, and subsequently named by Babcock (1978), are sections of borehole which have become elongated in one direction so that the cross sectional shape of the borehole is elliptical.

Two independent studies (Bell and Gough, 1979; Hottman *et al.*, 1979) offered an explanation for breakouts which is now generally accepted. Both studies point out that, if the subsurface horizontal stresses are unequal, a vertical, circular borehole would concentrate compressive stresses at the part of the borehole wall (see section 4.3) which intersects the azimuth of the minimum stress in the plane perpendicular to the borehole. The stress concentration could become large enough to cause shear failure and spalling at the borehole wall thus causing elongation of the borehole in the direction of the minimum stress (Figure 4.1). With the assumption that one principal stress is vertical, the orientation of the maximum and minimum horizontal principal stresses can be determined by measuring the azimuth of maximum elongation of the borehole.
Maximum Horizontal Stress ($\sigma_H$)

Maximum hoop stress $3\sigma_H - \sigma_H$

Breakout

Zone of shear failure

Minimum Horizontal Stress ($\sigma_T$)

Figure 4.1 Stress concentration around a vertical borehole with unequal horizontal stresses. The stress concentration is shown to cause shear failure and incipient spalling of the borehole wall giving rise to a borehole breakout aligned with the direction of $\sigma_H$.

The degree of elongation and its azimuth can be measured by a variety of down-hole logging tools. The most common is the four arm dipmeter which has four orthogonally opposed spring loaded arms. Although this tool is run for the purpose of determining the dip of beds, in the process it measures the eccentricity of the hole and the azimuth of the eccentricity. Acoustic and electrical imagery logging tools such as the BoreHole TeleViewer and the Formation MicroScanner are now increasingly used in the detection of breakouts.

Numerous borehole breakout studies have been conducted in various parts of the World (Bell and Gough, 1979, Bell, 1989; Bell at al, 1993, Brereton and Müller, 1991) and stress orientations derived from borehole breakouts have make up the majority of contributions to the World Stress Map in the top 4 km of the crust. A comprehensive study of borehole breakouts has been conducted in the North Sea (Cowgill, 1995; Cowgill et al., 1993).
The main advantages of the borehole breakout method are that the measurements are often made on a basin-wide scale and over a depth range of several kilometers. Further, the data necessary to detect breakouts is routinely recorded in oilwell logging. The drawback of breakout measurements, is that taken alone, they can not yield relative or absolute stress magnitudes. Observations of breakouts with combined with estimates of $\sigma_h$ from leak-offs or hydro-fracs in the context of some rock failure model applied to the borehole wall can be used to put limits on the magnitude of $\sigma_H$. This is discussed further in section 7.4.

*Hydraulic Fracturing* (these yield stress magnitudes and orientations):
The technique of hydraulic fracturing (hydro-fracing) is described in detail in the next section.

*Leak-off Tests* (these yield stress magnitudes):
Leak-off tests are described in detail in the next section

### 4.2.3 Direct Methods

*Flat Jack* (this yields stress magnitudes and stress orientations):
This is a simple and reliable stress magnitude measurement technique that can be used only at the surface or at shallow mine faces. A series of rectangular slots are cut, at orthogonal azimuths between measuring pins securely fixed into the rock, using a diamond saw. As the slots are cut, the previously secured measuring pins record the stress released in the form of recovered strain. Hydraulic flat jacks are then cemented into the slots and the pressure required to recover the initial strain is recorded.

This relatively cheap, robust technique has been used throughout Western Europe for the last 15 years (*e.g.* Froidevaux *et al.*, 1980). The main disadvantages of the technique are that six emplacements are normally needed for the assessment of the complete stress tensor, and only surface or shallow mine measurements are possible. Additionally, due to the near surface nature of the technique, it seems likely that measurements may be adversely influenced by local variations in the stress field. However, this method, together with the hydraulic fracturing technique, is perhaps
the most direct measure of stress magnitude. A stress, or at least a pressure, is actually measured as opposed to a strain from which the stress is inferred.

*Overcoring* (this yields stress magnitudes and orientations):

Overcoring is one of a number of strain relaxation methods which rely on the principal of strain recovery upon the release of the in-situ stress field (Engelder, 1993). Figure 4.2 summarises the steps taken during an overcoring measurement. Essentially, a borehole is drilled into a rock mass to a specific required depth, and a second, narrower hole is then drilled beyond the base of the first hole. The narrower, second hole is drilled to a diameter matching that of the measuring instrument (the deformation meter, comprising an array of strain gauges) that is to be used. The deformation meter is then cemented into the second hole and its surrounding rock volume is finally overcored, with the strain in the rock annulus being measured as the rock expands due to the overcoring stress relief.

![Figure 4.2](image)

*Figure 4.2* The overcoring technique for in situ stress measurement at shallow depths or within mines. (a) A 15cm diameter borehole is drilled into the rock mass under investigation; (b) a second smaller borehole is drilled to a depth of 30cm beyond the base of the original hole; (c) the deformation meter is installed in the smaller hole; (d) the deformation meter is then overcored and the strain relief measured. (After Cowgill, 1994; Adapted from Herget, 1993).

A knowledge of the elastic properties of the rock then leads to determination of the in-situ stress magnitudes. This method has an obvious advantage over the flat jack.
method in that the test can be carried out at depths of a few tens of metres in the rock mass, thus eliminating surface perturbations to the stress field. However, technical problems are increased as the deformation meters require careful handling and calibration. A minimum of four to six consistent tests are required to be certain of a reliable in-situ stress measurement. A good review of the technique is provided by Engelder (1993).

Anelastic Strain Recovery (this yields stress orientations and magnitudes):
The strain relaxation of a rock core after removal from its in-situ stress state comprises an instantaneous, elastic component and an anelastic (time-dependent) component. The anelastic strain recovery (ASR) technique measures the time dependent component of strain recovery which then theoretically enables an estimate of the entire in-situ stress tensor to be made. The method relies on the fact that the rock will expand most in the direction of maximum in situ stress, probably due mainly to the opening of microcracks, and requires oriented core which is placed in a strain cell able to measure the strain recovery of the core as it relaxes over time (Matsuki, 1991; Butterworth, 1993). Although ASR can theoretically recover the full stress tensor, in practice it is useful for the determination of in-situ horizontal principal stress orientations but is less useful for the determination of stress magnitudes, since estimates of stress magnitude depend upon assumptions about the stress strain relationship of the rock which in the inelastic domain are very poorly constrained.

Other methods which have been used in attempts to characterise the microcrack fabric of rocks in the laboratory and in the field in order to provide estimates of the in-situ stress, such as Acoustic Shear Wave Anisotropy (Yale et al., 1991) Differential Strain Curve Analysis (Batchelor and Pine, 1986; Teufel and Farrell, 1990), Kaiser Effect Gauging (Stuart, 1992; Holcomb, 1993), Core Disking (Kim et al., 1986; Dyke, 1989; Li, 1987), Extensive Dilatency Anisotropy (Crampin, 1987), Holographic Stress Measurements (Schmitt et al., 1984) and Borehole Slotting (Bock, 1993), have some individual advantages for particular measurement purposes, but none are, as yet, in widespread or routine use.
4.3 Stress Concentration around a Borehole

When a borehole is drilled into a rock mass, the presence of the borehole disturbs the pre-existing stress field in the rock. Stresses are concentrated at and around the borehole wall. The stress concentrations die out back to their unperturbed levels over a distance comparable to the diameter of the borehole.

The stress concentration around a borehole can be studied by considering the stress concentration caused by a circular hole in an infinite elastic plate subject to an external uniaxial stress (Figure 4.3). The solutions for the stress concentrations in such a plate are given by Kirsch (1898) and by Timoshenko (1982) in terms of a uniaxial remotely applied stress, $\sigma_s$. When considering a vertical borehole drilled into rock in which one principal stress is vertical, the stress $\sigma_s$ will be one of the principal horizontal stresses.

The solutions for the stress concentrations are best expressed in terms of polar co-ordinates with the centre of the hole as the origin and the plane stress components: radial stress $\sigma_r$, circumferential stress $\sigma_\theta$, and tangential shear stress $\tau_{r\theta}$ (Figure 4.3).

For a plate subject to a uniaxial stress $\sigma_s$, the stress components at a point on the borehole wall, or within the rock body, with polar co-ordinates $(r,\theta)$, where $r$ is the distance from the centre of the wellbore (the origin) and $\theta$ is the angle measured from the direction of $\sigma_s$, are:

$$\sigma_r = \frac{\sigma_s}{2} \left[ 1 - \frac{\alpha^4}{r^4} \right] + \frac{\sigma_s}{2} \left[ 1 + \frac{3\alpha^4}{r^4} - \frac{4\alpha^4}{r^4} \right] \cos 2\theta$$  \hspace{1cm} (4.1)

$$\sigma_\theta = \frac{\sigma_s}{2} \left[ 1 + \frac{\alpha^4}{r^4} \right] - \frac{\sigma_s}{2} \left[ 1 + \frac{3\alpha^4}{r^4} \right] \cos 2\theta$$  \hspace{1cm} (4.2)

$$\tau_{r\theta} = -\frac{\sigma_s}{2} \left[ 1 - \frac{3\alpha^4}{r^4} + \frac{2\alpha^4}{r^4} \right] \sin 2\theta$$  \hspace{1cm} (4.3)
Where \( a \) is the radius of the hole.

Figure 4.3 Circumferential stress concentrations \((\sigma_\theta)\) at \( \theta = 0^\circ, 90^\circ, 180^\circ \) and \( 270^\circ \) due to uniaxial stress \((\sigma_a)\) around a circular opening of radius \( a \), in an elastic plate.

It can easily be seen from equations 4.1 and 4.3 that at the borehole wall, where \( r = a \), the radial stress and the tangential shear stress are both zero. As the borehole wall is a free surface there can be no shear stress, and as at this stage there is no pressure within the hole, the radial stress must also be zero. Figure 4.3 represents the stresses present at the borehole wall under the above conditions. It can be seen that the expression for the circumferential stress reduces to the simple terms shown in Figure 4.3 for angles of \( \theta \) equal to 0, 90, 180 and 270 degrees.

For the case of a vertical borehole in the Earth where one principal stress is vertical, there are two horizontal principal stresses orthogonal to each other. In order to determine the stress field at and around the borehole wall as a result of the two stresses, we can simply superpose the results of two uniaxial stresses such as the one described above. The results of such a superposition, where the two horizontal principal stresses are unequal, are shown in Figure 4.4. The expressions for the circumferential stresses at
the borehole wall are, at the intersect of the borehole wall with the direction of the maximum horizontal stress ($\sigma_H$):

$$\sigma_q = 3\sigma_H - \sigma_h$$

(4.4)

and at the intersect of the borehole wall and the direction of the minimum horizontal stress ($\sigma_h$):

$$\sigma_q = 3\sigma_h - \sigma_H$$

(4.5)
Figure 4.4 The resultant circumferential stress field around a vertical borehole due to the concentration of two unequal far-field horizontal stresses, $\sigma_H$ and $\sigma_h$. 
4.4 Hydraulic Fracturing: Background and Test Procedures.

Hydraulic fracturing is a procedure which initiates and propagates tensile fractures from boreholes drilled into the brittle part of the Earth's crust. Hydraulic fractures were first intentionally propagated in the late 1940's by the oil industry. Oil bearing reservoirs were fractured in order to increase the permeability of the rock and thus assist the flow of oil from the formation into the wellbore.

The technique involves isolating a section of open hole usually either by using inflatable packers which are lowered downhole or perforating the well casing at the desired depth. The wellbore pressure is increased in the sealed off section by pumping fluid from the surface until the isolated formation breaks down by tensile failure (Figure 4.5).

![Figure 4.5 A section of borehole in the Earth's crust isolated using inflatable packers. The isolated section is then pressurised by pumping from the surface to induce a hydraulic fracture in the surrounding rock.](image-url)
The general consensus of opinion in the early days was that the formation was breaking
down by the propagation of bedding parallel horizontal fractures. Although it had been
pointed out that the total pressure at breakdown was less than the total overburden, it
was not until the subject was addressed theoretically (Hubbert and Willis, 1957) that it
was shown that most fractures, in stress regimes other than the reverse faulting regime
(see section 3.1.1), were in fact likely to be parallel to the axis of the borehole, i.e.
generally vertical.

Although hydraulic fracturing was not performed by the oil industry for the purpose of
measuring the in situ stress magnitudes, it was soon realised that much useful
information pertaining to the magnitude of the in situ stresses (particularly the minimum
in situ stress) can be read from the pressure records recorded during hydraulic fracturing
(e.g. Zoback and Haimson 1982).

During the past 20 years, the hydraulic fracturing or hydro-frac technique has become
widely regarded as being the most reliable method of determining the magnitude of the
minimum stress (generally the minimum horizontal stress magnitude, $\sigma_h$) in the Earth's
crust. Hydro-fracs performed for the purpose of measuring stress magnitudes tend to be
small volume tests, a typical volume of fluid pumped is a few gallons to a few tens of
gallons. Hydro-fracs performed specifically for the purpose of stress determination are
sometimes known as micro-fracs in the oil industry or simply hydro-fracs in the scientific
community. A medium volume test, often called a mini-frac or calibration frac, is
performed in the oil industry prior to the main fracture. The volume of fluid pumped into
the fracture in a mini-frac is of the order of several hundred barrels, or a few tens of
thousands of gallons. An oil industry calibration-frac is performed in order to determine
relevant parameters for the main frac design. Amongst these parameters, some of the
more important ones are a fluid leak-off coefficient and $\sigma_h$. The size of an oil industry
hydro-frac (the main fracture propagated to improve reservoir returns) will vary
according to the type of reservoir being fractured. The volume of fluid pumped into the
fractures which stimulate the southern North Sea gas reservoirs is of the order of several
thousand barrels or a few hundred thousand gallons.
4.5 Interpretation of Hydro-frac Pressure/Time Records

The method of hydro-frac stress determination is based on the interpretation of the pressure/time plot which is recorded throughout fracture initiation, propagation and closure. The test interval may be pressurised several times to reopen the fracture and measure the re-opening pressure and check the closure pressure. The downhole pressure, generally taken to be the pressure in the fracture, can be determined either by a downhole pressure gauge, or by measuring the pressure at the surface and then adding the pressure due to the hydrostatic weight of the fracturing fluid in the borehole above the fracture. Differences between the surface pressure and the downhole pressure due to fluid compressibility and frictional losses during pumping, mean that the downhole method is usually more accurate.

Figure 4.6 shows an idealised pressure/time record from a small volume hydro-frac stress measurement test with different points on the curve interpreted in terms of the crack behavior. The pumping rate up to breakdown is constant and the initial increase of pressure with time is linear. The breakdown pressure \( (p_b) \) in an idealised test such as that shown in Figure 4.6 is also the maximum pressure, it is a distinct peak and is interpreted as the point at which a tensile fracture both initiates and instantaneously propagates at a very fast rate. As will be seen in section 4.5.2, the pressure at which tensile rupture of the borehole wall initiates, is not always the pressure at which the fracture propagates rapidly. The breakdown pressure in more general terms then, is the pressure at which the fracture grows faster than fluid can be supplied from the surface and is therefore taken as the point where the pressure drops off. The physical background to the breakdown pressure is discussed in detail in section 4.5.2

The fracture will propagate if the pumping continues above a certain rate and a fracture propagation pressure \( (p_p) \) is sometimes determined, particularly in reservoir stimulation hydro-fracs where the intention is often to extend fractures for long distances. The fracture propagation pressure, once the fracture has grown to a length comparable to a few diameters of the borehole, will be slightly higher than the stress magnitude acting perpendicular to the fracture. The difference between the propagation pressure and the
minimum in situ stress is sometimes referred to as the net fracturing pressure. The net fracturing pressure is an important parameter for understanding the geometry (particularly the height) and growth behavior of large hydraulic fractures, for example, where a fracture is confined to one formation by a stress contrast, the net pressure is seen to rise (Economides and Nolte, 1989).

After pumping is stopped the system is shut in, i.e. it is sealed so that any drop in fluid pressure, after initial dynamic pressure differences have equilibrated, is due to fluid leaking into the rock of the wellbore wall or the fracture. When the pumping has stopped, the pressure in the fracture rapidly drops as the fracture soon stops growing and the pressure becomes equal to the pressure required only to hold the fracture open. This pressure is known as the instantaneous shut in pressure ($p_{iisp}$).

The closure pressure ($p_c$) is the pressure in the fracture just as it closes, it is therefore the pressure required to just hold open the fracture. Along the length of the fracture, apart

Figure 4.6 Idealised pressure/time record during a hydraulic fracture stress measurement test. $p_b =$ breakdown pressure, $p_p =$ fracture propagation pressure (injection pressure), $p_{iisp} =$ instantaneous shut in pressure (close to fracture closure pressure for small volume tests), $p_r =$ fracture re-opening pressure. (Adapted from Ervine and Bell, 1987).
from the near wellbore part where the stress concentration is present, the stress acting perpendicular to the fracture will generally be the minimum principal stress as it is against this stress that the fracture will have opened (Hubbert and Willis, 1957).

For fractures of small volume, such as those performed for the purpose of measuring the stress magnitude, the $p_{uip}$ and $p_c$ are virtually identical, whereas for large volume fractures, due to their geometry, there can be a significant difference between $p_{uip}$ and $p_c$ (Economides and Nolte, 1989).

For the purpose of stress determination, in a small volume fracture, the point on the pressure/time curve which corresponds to the $p_{uip}$ and the $p_c$ is taken as being equal to the minimum stress magnitude. In larger volume fractures such as the main frac or mini frac performed in reservoir stimulation, the $p_{uip}$ and the $p_c$ pressure are two different points on the pressure/time plot, and it is the closure pressure which is generally taken to be equal to the minimum stress magnitude. In the reservoir stimulation main frac, where a proppant has been pumped into the fracture, the pressure at which the fracture closes on the proppant is not necessarily equal to $\sigma_h$ (Economides and Nolte, 1989). Analysis of fracture closure combined with experimental results from small volume laboratory and field hydro-fracs, has lead some workers to conclude that the pressure at which the crack tip starts to close is the pressure closest to the minimum stress (Hayashi and Sakurai, 1989; Hayashi and Haimson, 1991).

After the fracture has closed and the pressure has fallen back to the hydrostatic pressure in the wellbore, subsequent pressurisation cycles are generally performed. During this re-pressurisation the fracture re-opens, and a re-opening pressure ($p_r$) can be detected from the pressure/time plot. The re-opening pressure is defined as the pressure at which the pressure/time relationship deviates from linearity during re-pressurisation (Lee and Haimson, 1989; Ito and Hayashi, 1992). A distinction should be made here between re-opening pressure from a fast re-frac, and the re-opening pressure from a slow re-frac. A fast re-frac is sometimes performed to measure the stress concentration (i.e. to calculate $\sigma_H$ once $\sigma_h$ has been determined) at the borehole wall. During a fast re-frac it is assumed that the fracture induced in the breakdown cycle has closed perfectly and is thus not
permeable to the fracturing fluid. In this case, the re-opening pressure is the pressure at which the fracture at the borehole wall just starts to open, the re-opening pressure is therefore equal to the breakdown pressure minus the tensile strength. During a slow refrac, it is not assumed that the fracture has closed perfectly. The slow re-frac is often used to measure just $\sigma_h$. In this study, the term re-opening pressure (unless otherwise indicated) is used to imply a pressure equivalent to that obtained in the slow re-frac. The interpretation of re-opening pressure in terms of rock stress magnitudes is discussed in section 4.5.1.

4.5.1 Determining $\sigma_h$ from Analysis of Hydro-frac Pressure/Time Records.

In a vertical borehole, the most common case is for a vertical fracture to be produced by hydraulic fracturing. This is the case when the minimum principal stress is horizontal (normal or strike slip stress regime). It is therefore usually $\sigma_h$ which is determined from the pressure/time record.

The value of $\sigma_h$ can be determined either by identifying the closure pressure (equivalent to the isip in small volume tests) or by identifying the re-opening pressure in the case where the fracture is hydraulically open during wellbore pressurisation (equivalent to slow re-frac - see above).

**Closure Pressure**

After shut-in, the fracturing fluid slowly leaks into the rock through the surface area of the fracture and through the well bore wall. As the fracture closes the surface area of the fracture becomes smaller and so the rate of fluid leakage into the formation decreases. At the point where the fracture closes there is a relatively sharp change in the area of formation exposed to the fluid. The transition from fluid leaking into both fracture surfaces and wellbore wall to fluid leaking into just wellbore wall produces a discrete change in the rate of pressure decay. If this change in pressure decay rate can be detected from the pressure/time plot, the closure pressure and thus $\sigma_h$ can be determined.
In the ideal case (Figure 4.6), the closure pressure (or $p_{c_{u}}$) can be detected from the pressure/time plot as the inflection point on the pressure decay slope (Ervine and Bell, 1987). However, the change in slope of the pressure decay/time plot is often very subtle and numerous methods have been proposed to extract the correct information from the data. Such methods include:

- Plotting pressure against the square root of time since shut-in, commonly termed the shut in decline test (see below and section 5.5.2) (Economides and Nolte, 1989; Fjaer, 1992).

- Plotting rate of pressure decline against time (Nirex Hydro-frac factual reports, 1991-1994 section 5.5.1).

- Extracting the point of maximum curvature of pressure against time to be equal to the pressure at the onset of crack tip closure and thus $\sigma_b$ (Hayashi and Sakurai, 1989).

- Plotting the inverse of the pressure decrease rate against pressure to identify different stages of crack closure (Hayashi and Haimson, 1991).

- Plotting log (pressure) vs. log (time) to reveal the change gradient at fracture closure (Zoback and Haimson, 1982).

The “shut in decline test” is used by Dowell Schlumberger after a calibration fracture or mini frac (i.e. when there is no proppant in the fracture) to determine the fracture closure pressure. The pressure after shut in is plotted against square root of time since shut in. The curve produced by this plot approaches a straight line as long as the fracture remains open (Economides and Nolte, 1989; Fjaer et al. 1992). When the fracture closes, the pressure decline curve will depart from this linear trend. As illustrated in Figure 4.7, this change in slope can be in either direction, depending on the formation properties.
It is recognized that the subtlety in change of slope can introduce errors in picking the point of departure from linearity and because of these ambiguities, the step rate test (see below) is commonly performed as well as the shut-in decline test. Using both techniques in conjunction is the most reliable method of $\sigma_b$ determination.

**Re-opening Pressure**

During the re-pressurisation cycle of a hydro-frac, after a fracture has been propagated and then allowed to close in the first cycle, the reopening pressure is generally taken as the pressure at which the pressure/time plot deviates from linearity. The re-opening pressure has been interpreted as being equal to the breakdown pressure minus the tensile strength (Bredehoeft et al., 1976; Ervine and Bell, 1987). This interpretation assumes the fracture remains hydraulically closed during re-pressurisation. If the rate of pressurisation is high and the fracturing fluid viscous, this might be the case. The re-opening pressure obtained during a fast re-frac is sometimes interpreted in this way. However, it has been noticed that in many cases that the re-opening pressure is actually very close to the value of the closure pressure (Lee and Haimson, 1989). It has been demonstrated theoretically and in laboratory experiments, that if the pressurisation rate
is not very fast, and the fluid is not very viscous, then the fracture is likely to be hydraulically open during re-pressurisation (Ito and Hayashi, 1992). It seems likely that when the fracture closes at the end of the first cycle, it will not close perfectly. Only a very small amount of shear movement between the two fracture surfaces will cause a mismatch when they close so that asperities on each surface will prop the fracture open slightly. In this case, the fracture is mechanically closed, but hydraulically open, so that the wellbore fluid can penetrate the fracture during pressurisation. In this case, the fracture will start to open when the pressure inside the fracture is equal to the minimum stress acting across it. For a fracture of length comparable to the diameter of the borehole or more, this stress will be close to $\sigma_b$.

A test to determine $\sigma_b$ across oilwell hydro-fracs, which relies on the fact that the re-opening pressure is close to the closure pressure (or $\sigma_b$) is the step rate test (equivalent to the so called slow refrac). A step rate test involves re-pressurising the fractured interval after the first pressurisation cycle. The interval is re-pressurised at incrementally increasing pumping rates (fluid injection rates), each increment of pumping lasting for the same length of time (Figure 4.8a). A plot is made of the pumping rate vs. wellbore pressure at the end of each increment (Figure 4.8b). The pressure at which this plot deviates from linearity is the fracture re-opening pressure, and is taken to be close (although slightly above - due mainly to frictional effects) to $\sigma_b$. In practice, the pressure
at which the pumping rate vs. wellbore pressure plot becomes a plateau is often read from the plot (Figure 4.8b). This pressure is assumed to be the fracture extension pressure which is slightly higher than the fracture re-opening pressure and is taken as an upper bound for $\sigma_b$. The difference between re-opening pressure and extension pressure is best illustrated by the real example from the step rate test record from the southern North Sea shown in Figure 5.13. During a step rate test, the fracture extension pressure determined is thought to be between 50 to 200 psi above the closure pressure, the difference being due to fluid friction in the fracture and the fracture toughness of the rock (Economides and Nolte, 1989). The re-opening pressure is somewhere between the closure pressure and the fracture extension pressure. For the rest of this study (unless otherwise mentioned) the re-opening pressure is assumed to be equivalent to the pressure obtained from a slow re-frac or step rate re-opening test, and thus be a good measure of $\sigma_b$.

4.5.2 Breakdown Pressure Under Various Wellbore Boundary Conditions.

The leak-off pressure obtained during a leak-off test (see section 4.6.2) is to some extent physically equivalent to the breakdown pressure in a hydro-frac test. For this reason the physical background to the breakdown pressure is examined in some detail in this section.

The analysis of the breakdown pressure involves a consideration of the stress field acting in the rock at and around the wellbore wall. This stress field can be analysed in terms of the superposition of stresses from various sources which affect the rock. The stress fields around the wellbore analysed in this section are:

(i) The stress field due to the far field stresses
(ii) The stress field due to the fluid pressure within the wellbore
(iii) The stress field due to the presence of a pore fluid in the rock around the wellbore
(iv) The stress field due to the radial flow of fluid from the wellbore into the surrounding rock.
(v) The stress field due to fluid pressure in a pre-existing crack which is permeated by the wellbore fluid during wellbore pressurisation.
These stresses can be progressively added as the boundary conditions which are applied to the physical character of the wellbore wall and the surrounding rock are progressively relaxed. In all cases the rock is assumed to be elastic, homogeneous and isotropic.

4.5.2.1 Tensile Fracturing of an Impermeable Wellbore Wall and Non-porous Rock.

The stresses at and around the borehole wall due to the far field stresses are given in section 4.3. The application of a fluid pressure within the borehole produces additional stresses. For the case where the wellbore wall is impermeable, these stresses can be derived from the Lamé solution for the stresses in a thick walled elastic cylinder which is given by Timoshenko (1982). If the outer radius of the cylinder is allowed to become very large and the external pressure is set equal to zero, the solution becomes applicable to the wellbore problem and the radial, circumferential and axis parallel stresses become:

\[ \sigma_r = \frac{p_w a^2}{r^2} \]  \hspace{1cm} (4.6)

\[ \sigma_\theta = -\frac{p_w a^2}{r^2} \]  \hspace{1cm} (4.7)

\[ \sigma_z = 0 \]  \hspace{1cm} (4.8)

in which \( \sigma_z \) is the stress parallel to the wellbore axis and \( p_w \) is the fluid pressure applied within the wellbore.

To obtain the complete stress field, the stress caused by the pre-existing regional stress (given by equations 4.1 - 4.3, and shown in Figure 4.4) is superposed on the stress caused by the wellbore pressure. The radial stress at all points on the wellbore wall is equal to \( p_w \). At \( \theta = 0 \) and 180 degrees on the wellbore wall the circumferential stress is a minimum. The criteria for tensile failure is satisfied when the circumferential stress at the wellbore wall is less than the tensile strength (\( T \)) of the rock (where tensile stress is negative and compressive stress is positive).

The circumferential stress at the wellbore wall at \( \theta = 0 \) and 180 degrees is:

\[ 3\sigma_\theta - \sigma_z - p_w \]
So tensile failure occurs when:

\[-T > 3\sigma_b - \sigma_H + p_w\]  \hspace{1cm} (4.9)

It is assumed that once tensile failure has occurred, the formation will breakdown, that is to say that as soon as the fracture is initiated, it will propagate rapidly. So the breakdown equation for this most simple condition is:

\[p_b = 3\sigma_b - \sigma_H + T\]  \hspace{1cm} (4.10)

Where \(p_b = p_w\) at breakdown.

**4.5.2.2 Tensile Fracturing of an Impermeable Wellbore Wall in Porous Rock.**

The assumption that the rock is non porous is not very realistic. Even for crystalline rocks of very low porosity the effective stress law is thought to still hold in most cases (Schmitt and Zoback, 1993). It is therefore the effective circumferential stress at the borehole wall which must be equal to the negative value of the tensile strength. In this analysis the borehole wall is still considered to be impermeable (the rock itself will have some permeability, but the viscosity of the drilling fluid combined with a relatively high pressurization rate may be equivalent to an impermeable borehole wall).

Effective stress (where compressive stresses are positive) is assumed to be the Terzaghi effective stress, i.e. the total stress minus the local pore pressure \((\sigma' = \sigma - p)\) where \(\sigma\) is any normal stress, \(\sigma'\) is the effective stress and \(p\) is the pore fluid pressure in the rock.

The effect of the wellbore on the far field stresses is to concentrate the effective stresses, such that the effective stress at the wellbore wall (at \(\theta = 0\) and 180 degrees as these remain the points where this stress is a minimum) due to the far field effective stresses is \(3\sigma' - \sigma_H\). Thus the total stress (due to the far field stresses) at these points on the wellbore wall \((\sigma_{TB})\), is now: \(3\sigma' - \sigma_H + p\).

Now that it is effective stresses which are being considered at the borehole wall, the pressure in the wellbore must initially be set equal to the pore pressure in the rock. It is therefore now only increases above this pressure in the wellbore which cause additional
stresses in the wellbore wall, i.e. the total stress (due to fluid pressure in the wellbore) at the wellbore wall \((\sigma_{Twp})\) is now: \(-(p_w-p)\).

The total stress acting at the wellbore wall can be obtained by superposing the total stress due to the far field stresses \((\sigma_{Tf})\) and the total stress due to the wellbore fluid pressure \((\sigma_{Twp})\).

Total stress at the wellbore wall at \(\theta = 0\) and 180 degrees is thus:

\[ (3\sigma'_h - \sigma'_H + p) - (p_w - p) \]

The failure criteria is that breakdown occurs when the effective stress at the wellbore wall overcomes the tensile strength i.e. the effective stress becomes less than the negative value of the tensile strength. The effective stress at the borehole wall is now the total stress given above minus the pore pressure:

Effective stress at the borehole wall at \(\theta = 0\) and 180 degrees:

\[ (3\sigma'_h - \sigma'_H + p) - (p_w - p) - p \]

And so the breakdown occurs when:

\[ (3\sigma'_h - \sigma'_H + p) - (p_w - p) - p = -T \]  \(\text{(4.11)}\)

in terms of total far field stresses this becomes:

\[ (3\sigma_h - p) - (\sigma_H - p) + p - (p_w - p) - p = -T \]  \(\text{(4.12)}\)

Therefore the breakdown equation (where again \(p_b\) is \(p_w\) at breakdown) is now:

\[ p_b = 3\sigma_h - \sigma_H - p + T \]  \(\text{(4.13)}\)

This is the so called classical breakdown equation which is generally attributed to Hubbert and Willis (1957). During wellbore pressurisation, the breakdown pressure is generally higher \(\sigma_h\). It can be seen from equation 4.13 that this will be the case when the far field stress anisotropy is not too great, that is when:

\[ \sigma_H < 2\sigma_h - p + T \]  \(\text{(4.14)}\)
4.5.2.3 Tensile Fracturing of a Permeable Wellbore Wall in Porous Rock.

The next boundary condition which can be relaxed is that of permeability of the wellbore wall. If the wall is permeable, fluid from the wellbore will permeate into the rock during wellbore pressurisation and this will cause additional stresses. Some new boundary conditions must be introduced. These are that the fluid flow into the rock obeys Darcy's law and that the permeability is isotropic so that the flow will be radial. Also that the wellbore fluid has properties similar to the formation pore fluid.

The principle of superposition of stresses can be used again here. The additional stress caused by this radial flow of fluid into the rock can be added on to the stresses described for the non-penetrating case. Fluid flow through porous media gives rise to stresses and displacements in the material analogous to those caused by heat conduction through solids and known results from the theory of thermoelasticity may be easily modified to solve problems in poroelastic materials (Haimson and Fairhurst, 1967). In the case of a thick walled cylinder where the outer radius is much bigger than the inner one, the stresses caused by the fluid flow are given by Haimson and Fairhurst (1967):

\[
\sigma_r = \frac{\alpha(1-2\nu)}{r^2(1-\nu)} \int_0^r \Delta p(r) r dr \\
\sigma_\theta = -\frac{\alpha(1-2\nu)}{(1-\nu)} \left[ \frac{1}{r^2} \int_0^r \Delta p(r) r dr - \Delta p(r) \right] \\
\tau_{r\theta} = 0
\]  

Where \(\alpha\) is Biot’s poroelastic parameter and \(\nu\) is Poisson’s ratio and \(\Delta p\) is the increase in pore fluid pressure above the original value \(p\).

In order to obtain values for the stress induced by this flow within the rock around the wellbore, the distribution of pressure with distance form the wellbore must be known.
However, to obtain the value of this stress at the wellbore wall, only the pressure at the wellbore wall needs to be known. As the wellbore wall is now permeable, the pressure at the wellbore wall will now be equal to the wellbore pressure. At the wellbore wall, the radial stress due to the fluid flow is zero and the total circumferential stress due to the fluid flow at the wellbore wall ($\sigma_{\text{fn}}$) is a compressive stress:

$$\sigma_{\text{fn}} = \frac{\alpha(1-2\nu)}{(1-\nu)}(p_w - p)$$  \(4.18\)

The total stress acting at the wellbore wall can be obtained by superposing the total stress due to the far field stresses ($\sigma_{\text{f}}$), the total stress due to the wellbore fluid pressure ($\sigma_{\text{w}}$), and the total stress due to radial fluid flow ($\sigma_{\text{r}}$).

Total stress at the wellbore wall at $\theta = 0$ and 180 degrees is thus:

$$\left(3(\sigma_h - p) - (\sigma_H - p) + p_w - p\right) + \left(\frac{\alpha(1-2\nu)}{(1-\nu)}(p_w - p)\right)$$

The effective stress at the wellbore wall is the total stress at the wellbore wall minus the pore pressure. As the wellbore wall is permeable, the pore pressure is now equal to the wellbore pressure and so the effective stress at the wellbore wall at $\theta = 0$ and 180 degrees is now:

$$\left(3(\sigma_h - p) - (\sigma_H - p) + p_w - p\right) + \left(\frac{\alpha(1-2\nu)}{(1-\nu)}(p_w - p)\right) - p_w$$

When this effective stress becomes less than the negative value of the tensile strength tensile fracturing will occur. The breakdown equation now, is thus:

$$p_b = \frac{3\sigma_h - \sigma_H + T - \alpha \frac{1-2\nu}{1-\nu} p}{2 - \alpha \frac{1-2\nu}{1-\nu}}$$  \(4.19\)

This is the same as the breakdown equation obtained by Haimson and Fairhurst (1967).

4.5.2.4 Tensile Fracturing of the Wellbore Wall Containing Pre-Existing, Permeable Cracks

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In sections 4.5.2.1 - 4.5.2.3 it has been assumed that the wellbore wall is intact until breakdown. In most real rocks there are pre-existing micro-cracks into which the fracturing fluid will penetrate during wellbore pressurisation, prior to crack propagation (Rummel and Hansen, 1989). Often the rock that forms the wellbore wall can be particularly cracked as a result of the mechanical processes associated with drilling.

Cracks aligned perpendicular with the minimum stress are most likely to be permeable. If fluid penetrates the cracks prior to breakdown, the distribution of stresses acting on the rock is fundamentally different to those considered so far. Also the presence of permeable cracks means that the breakdown criteria used so far, that of breakdown occurring when the stress at the wellbore wall overcomes the tensile strength, is no longer valid. New stresses, which arise from the pressure of the fluid within the crack, and a new breakdown criterion must be considered.

The case of pre-existing, permeable cracks in the wellbore wall can be analysed in terms of fracture mechanics (Abou-Sayed et al., 1978; Rummel, 1987; Detournay and Carbonell, 1994). The fracture mechanics approach takes account of the fluid pressure on the walls of the pre-existing crack prior to breakdown. In the theory of fracture mechanics, critical crack growth occurs when the stress intensity at the crack tip overcomes the fracture toughness of the material. It is assumed here that hydro-frac breakdown corresponds to critical crack growth.

An outline of the approach used by Rummel (1987) to obtain a fracture mechanics breakdown equation is given here. The problem is considered in two dimensions with a vertical borehole and unequal horizontal stresses. It is assumed the rock is isotropic and that there is no fluid loss to the formation. The stress intensity in the region of the crack tip is formulated by the superposition of stress intensity factors from each of the following loading sources:

- $K_1(\sigma_{H})$ - stress intensity due to the far field maximum horizontal stress
- $K_1(\sigma_{h})$ - stress intensity due to the far field minimum horizontal stress
- $K_1(p_w)$ - stress intensity due to the pressure acting on the borehole wall
- $K_1(p_i)$ - stress intensity due to the pressure acting on the walls of the crack
Where $K_i$ indicates stress intensity factors for mode 1 (tensile) crack propagation.

Stress intensities in the region of the crack tip due to these loading sources are given by Rummel (1987), for $\sigma_H$:

$$K_i(\sigma_H) = -\sigma_H \sqrt{a} f(b)$$  \hspace{1cm} (4.20)

Where $a$ is the radius of the borehole, $f(b)$ is a dimensionless stress intensity factor, and $b$ is the normalised crack length $b = 1 + (l/a)$ where $l$ is the length of the crack. Two solutions for $f(b)$ are shown graphically in Figure 4.9a.

For $\sigma_h$:

$$K_i(\sigma_h) = \sigma_h \sqrt{a} g(b)$$  \hspace{1cm} (4.21)

where $g(b)$ is a dimensionless stress intensity factor which is represented in Figure 4.9b.

The stress intensities due to the pressure in the wellbore, $p_w$, and the pressure in the crack $p_i$ are given:

$$K_i(p_w) = p_w \sqrt{a} h_w(b)$$  \hspace{1cm} (4.22)

$$K_i(p_i) = p_i \sqrt{a} h_i(b)$$  \hspace{1cm} (4.23)
Where $h_0(b)$ and $h_1(b)$ are dimensionless stress intensity factors. The function $h_1(b)$ is dependent on the pressure distribution within the crack, $p_i(x)$ where $x$ is the Cartesian co-ordinate axis parallel to the direction of the crack and with origin at the centre of the wellbore.

A combined stress intensity function, $h(b) = h_0(b) + h_1(b)$ is shown graphically in Figure 4.9c for some possible pressure distributions in the crack.
Figure 4.9c Dimensionless stress intensity function $h$, as a function of normalised crack length $b$ ($b=1+l/a$) for various pressure distributions within the crack: 1, pressure at all points in the crack equal to wellbore pressure; 2, pressure at all points in the crack equal to 0.75 of the wellbore pressure; 3, linear pressure drop in the crack from wellbore pressure at the wellbore wall to zero at the crack tip; 4, quadratic pressure drop from wellbore pressure at the wellbore wall to zero at the crack tip. (After Rummel, 1987).

Superposition of the stress intensities due to the above loading sources yields the relationship for the critical borehole pressure at unstable crack propagation (breakdown pressure):

$$p_b = \frac{1}{h(b)} \left( \frac{K_{ic}}{h} + \sigma_n f(b) + \sigma_n g(b) \right)$$  

(4.24)

Where $K_{ic}$ is the critical stress intensity or the fracture toughness.

If this equation is compared to the classic breakdown equation (Hubbert and Willis, 1957), the term:

$$\frac{K_{ic}}{h(b)\sqrt{a}}$$

can be interpreted as a pseudo tensile strength term under zero external stresses (Rummel, 1987). It is clear that the pseudo tensile strength decreases with increasing wellbore radius, and so the breakdown pressure is predicted to decrease with increasing wellbore radius, a phenomenon which is indeed often observed (Rummel, 1987; Guo et al., 1993; Ito et al., 1990).
Due to the dependence of the fracture mechanics breakdown pressure on the fluid pressure distribution in the crack, the fracture mechanics approach is also able to explain some other variations in breakdown pressure which are not predicted by the simple breakdown equations of Hubbert and Willis (1957) or Haimson and Fairhurst (1967). For a crack of given hydraulic conductivity, the distribution of fluid pressure within the crack, as the wellbore pressure begins to increase, will be affected by the rate of increase of the wellbore pressure and the viscosity of the fluid. For fast rates of wellbore pressurisation and high viscosity fluids, there is likely to be a decrease in fluid pressure within the crack between the wellbore end and the tip. For slow pressurisation rates and low viscosity fluids, the pressure distribution with distance from the wellbore is likely to be linear.

In the case of a pressure drop within the crack between the wellbore and the tip, the wellbore pressure is not all being applied to the walls of the crack. Where there is no pressure drop in the crack, all the wellbore pressure is being transmitted to the walls of the crack. As fluid pressure on the walls of the crack promotes breakdown, it is clear that where there is a pressure drop in the crack, the breakdown pressure will be higher than when there is no pressure in the crack, all other factors being the same.

An increase in breakdown pressure with wellbore pressurisation rate (and less often fluid viscosity) has commonly been recorded and discussed (Guo et al., 1993; Ito and Hayashi, 1991; Detournay and Cheng, 1992).

Another effect of fluid pressure gradient in the crack between the wellbore and the tip is suggested by Detournay and Carbonell (1994). They argue that the criteria for unstable crack propagation (and therefore breakdown reflected in the pressure peak on hydrofrac pressure/time plot) is not only that (i) stress intensity at the crack tip be equal to the fracture toughness of the rock, but also that (ii) an increase in the length of the crack leads to an increase in stress intensity at the crack tip under the prescribed loading conditions. When condition (i) is fulfilled but not condition (ii), the crack may propagate stably so that a further increase in wellbore pressure is required to induce breakdown.
They go on to show that under fast loading conditions, where the fluid pressure in the wellbore is not transferred to the wellbore walls, stable fracture propagation can occur initially, followed by unstable propagation at higher pressures. The onset of stable fracture propagation corresponds to a fracture initiation pressure, which may be reflected in the pressure/time plot as a deviation from linearity as the fracture begins to open, and the onset of unstable fracture propagation corresponds to the breakdown pressure which is the peak pressure.

It should be noted that the case of a pre-existing crack which is oriented perpendicular to the minimum stress and which a permeable length comparable to the diameter of the borehole, is equivalent to the case of fracture re-opening, discussed in section 4.5.1 in which the re-opening pressure, under the condition of slow wellbore pressurisation and a fluid which does not have a very high viscosity, is equal to the minimum stress magnitude.

### 4.6 Oilwell Drilling and Leak-off Tests.

During the drilling of oilwells, there needs to be a column of fluid, called drilling mud, in the wellbore for several reasons. Firstly to control the well, that is to supply a pressure to the exposed rock in the walls of the well in order to prevent the rocks pore fluid (which is generally at or above hydrostatic pressure) from flowing out of the rock into the well, causing a “kick” or if not stopped in time, a “blowout” which can be very dangerous. Secondly, the drilling mud is circulated, by pumping down the drillstring and then back up the annulus (Figure 4.10). The rheology of the mud is such that it brings the cuttings from the drillbit to the surface. Some other effects of the mud are to lubricate the drill bit, and sometimes to chemically stabilise certain lithologies like clays where swelling can cause problems. However, the mud weight (density, or equivalent circulating density if frictional flow and effect of cuttings is taken into account) must not be so high as to cause a pressure on the rock that will propagate tensile hydraulic fractures (see section 4.5.2). When tensile fracturing is accidentally induced during drilling, formation fluid can flow uncontrollably into the fracture and lost circulation occurs. The pressure exerted by the drilling mud must be kept at a value between the formation pore pressure and the hydraulic fracturing breakdown pressure, in highly
overpressured areas this can be very difficult (French and Mclean, 1992) as there may be very little difference between the two.
Figure 4.10 Schematic representation of oilwell drilling during a leak-off test, showing the relationship between the open hole, casing, cement, drill pipe, annulus, and mud flow.
During the drilling of deep boreholes, steel casing is generally set at stages of roughly every few thousand feet (Figure 4.10). The casing has cement pumped down it so that the cement comes back up the outside, between the rock and the casing, holding it firmly in place. The cased section is now protected so that it can not be responsible for lost circulation or blowouts. An example of where casing might need to be set is where a known, or suspected overpressured region was to be drilled. In order to balance the overpressure, the mud weight would need to be raised. However, if it was thought that this increase in mud weight would be high enough to fracture the normally pressured formations above, casing would be set down to the onset of overpressure, so that the normally pressured formations would be protected.

After the setting of the casing, the next section of the hole is ready to be drilled. The diameter of the hole decreases by a few inches after each casing string. A section of a few feet to a few tens of feet is then drilled through the cement and casing shoe (a drillable guide piece on the end of the casing) and into the next body of rock to be drilled. At this stage a leak-off test is performed. Mud is pumped down the hole to increase the pressure on the cement and newly drilled formation at the bottom. This tests the integrity of the cement job and the integrity or “strength” of the formation in the open section of hole. The pressure to which the bottom of the hole was subjected during the test must not be exceeded during drilling of the next section of the hole. Details of leak-off test procedures and interpretations are given below.

4.6.1 Leak-off Test Terminology.

Leak-off test procedures and terminology vary between operators and drilling engineers. To the best of my knowledge there is no published standard glossary of terms pertaining to this type of test. Before leak-off tests are described in more detail, some terms must be defined.

*Standard leak-off test:* Leak-off test that is taken to leak-off but no further (see below).

*Limit test (Leak-off to):* Leak-off test that is stopped before leak-off occurs
**Formation integrity test:** Same as a standard leak-off test (although sometimes used to denote limit test).

**Fracture test:** A leak-off test where the pumping is continued beyond leak-off until formation breakdown occurs (sometimes there is no leak-off pressure before breakdown). The pressure decline after pumping has stopped may sometimes be monitored.

**Extended leak-off test:** The leak-off test is taken to the point of fracturing and the well is then shut in and pressure decline monitored. Subsequent pressurisation cycles may then be performed. This process is much more like a small hydro-frac test.

**Double leak-off test:** A standard leak-off test, limit test or fracturing test that is performed twice consecutively.

**Leak-off pressure:** The pressure at which the pressure/volume plot recorded during a leak-off test deviates from linearity (see section 4.6.2).

**Fracturing pressure:** The formation breakdown pressure, the maximum pressure where the test is taken to breakdown. Fracturing pressure and leak-off pressure are often used interchangeably, however in most cases the leak-off pressure (as defined above) is distinct and lower than the fracturing pressure.

**Fracture gradient:** The fracture pressure divided by vertical depth below surface. The fracture gradient however seems to be very commonly calculated with leak-off pressures and so in most cases should probably be called the leak-off pressure gradient.

**Equivalent mud weight (EMW):** The pressure in the borehole at some vertical depth, expressed in terms of the mud weight which would cause such a hydrostatic pressure at that depth if it were in the borehole. Hydrostatic pressure (psi) = EMW (pounds per gallon) x vertical depth (ft) x 0.0519.

### 4.6.2 Leak-off Test Procedure.

The leak-off test procedure varies between operators and drilling engineers. Some companies have a well defined leak-off test procedure (Daget and Parigot, 1979; Kunze and Steiger, 1991) although there seems to be a degree of subjectivity on the part of the drilling engineers as to how exactly to carry out each test.
Leak-off tests are performed routinely, particularly in poorly known or overpressured areas, by drilling engineers during the drilling of any boreholes that require casing to be set in different stages (mainly but not exclusively in the oil industry). The leak-off test is not performed for the purpose of measuring stress magnitudes. It is performed in order to test the integrity of the cement at the casing shoe and to determine the maximum mud weight which can be used to drill the next section of hole. This is sometimes referred to as the strength or mud holding capacity of the formation in the open hole below the casing shoe.

Tests are usually performed in around ten to fifty feet of open hole which has been drilled below the casing shoe (Figure 4.10). Fresh drilling mud should be circulated around the drill string and annulus so that it is free of cuttings which would affect the density. The drill bit is pulled back up into the casing and the annulus is closed at the top of the hole (Figure 4.10). Mud is then pumped down the hole to increase the bottom hole pressure. The mud is usually pumped using the cement pump which is a small volume, high pressure pump.

Mud is slowly pumped into the drillpipe (Dickey, 1979). The rate of pumping varies and is commonly at the discretion of the drilling engineer and limited by the capacity of the pump. Pumping rates are rarely recorded during leak-off tests, however, according to previously published references (Daget and Parigot, 1979; Enever et al., 1996) to pumping rates, or where the rate can be inferred from a leak-off test report, rates are between 0.25 and 1 barrel per minute. The total volume pumped can be anything from half a barrel to over ten barrels (the barrel is a common measure of volume in the oil industry and 1 US barrel is equal to 42 US gallons or approximately 160 litres).

The amount by which one barrel pumped raises the pressure depends on several factors: the compressibility of the mud, the compressibility of the pump tubing and cased borehole section and the permeability/poroelastic properties of the open hole formation. Indeed, it has been argued by Enever et al. (1996) that the most serious problem with leak-off tests as a stress determination technique is that pressure may be an artifact of compressibility and leakage of the tubing and casing system. A plot of the “stiffness”
(surface pressure increase per barrel pumped) against surface area of open hole during a leak-off test from over 100 North Sea wells drilled for Amerada Hess (Figure 4.11), indicates that the amount of formation exposed in the leak-off test strongly affects the "stiffness". There is a negative correlation between "stiffness" and open hole surface area, indicating that it is mainly the rock properties which control the pressure build up during a leak-off test.

Figure 4.11 Leak-off test system "stiffness". Pressure increase per barrel pumped against the surface area of the open hole below the casing shoe.

Pumping of the mud is sometimes continuous and sometimes incremental. During continuous pumping the surface pump pressure is noted every half or quarter barrel. When pumping is incremental, the pump is stopped after every half or quarter barrel and the pressure recorded. Sometimes two pressures are recorded, a dynamic pressure as the pump is turned off, and a static pressure one or two minutes after the pumping has stopped and the system has equilibrated. The static pressure is lower, usually by a few tens of pounds per square inch (psi) and is likely to more accurately reflect the bottomhole pressure.

Some operators decide whether to pump incrementally or continuously depending on the type of formation being tested. For example, permeable formations can yield ambiguous
results when incremental pumping is used whereas it is possible to pump against pressure loses when pumping is continuous, giving a more linear pressure build up. The choice of test type can also depend on the degree of accuracy that is thought necessary, the incremental type in most cases being considered more accurate (Pers. com - Drilling Engineers, Amerada Hess, Aberdeen).

A limit test is often performed instead of a leak-off test. If the maximum mudweight needed to drill the next section has already been estimated (this can only be done in cases where other wells have been drilled in the vicinity), the leak-off test can be stopped when the bottom hole pressure reaches the equivalent of this mud weight.

4.6.3 Leak-off Test Records and the Leak-off Pressure.

The leak-off test record is made during and shortly after the test. The amount of information which is recorded varies greatly although it seems that the amount and quality of data recorded is generally improving as operators become increasingly aware of the potential amount of useful and essentially cost free information available from such a test.

The main part of the leak-off test record is a plot of surface pump pressure against volume of mud pumped. This relationship is ideally linear after the first barrel or so has been pumped and air has been expelled from the system (Figure 4.12). The pumping continues until the pressure/volume plot deviates from linearity. The pressure at which this happens is the leak-off pressure. There appears to be a degree of subjectivity involved in picking the leak-off pressure. Engineers who perform these tests sometimes observe a sudden decrease in the rate of pressure build up, or a pressure drop, at which point the test can be stopped. Often however, there is a gradual decrease in the in the rate of pressure build up which can be seen when the pressures are plotted against volume. In these cases the pumping may well continue for several more increments or until the pressure does suddenly drop, and sometimes beyond this point. Leak-off pressures reported seem to vary between the initial point of deviation from linearity and the maximum pressure. There is clearly a need for the terms used to describe the leak-off test to be more rigorously defined.
Leak-off test records also record a variable amount of further information which can include: diameter of the well casing, length of the open hole, number of barrels pumped during the test, number of barrels returned to the surface at the end of the test, lithology in the open hole, some of the mud properties, the driller's interpretation of whether leak-off occurred and if it did, the leak-off pressure picked by the driller.

4.6.4 Previous Interpretations of Leak-off Pressure.
Although the standard leak-off test is not performed for the purposes of measuring stress magnitudes, information from leak-off tests has often been used in a variety of
ways, albeit somewhat inconsistently, to estimate the magnitude of the minimum stress. It is the similarity of the leak-off test to the first part, or the re-opening part, of a hydrofrac stress measurement test that invites comparison of the leak-off pressure to some function of stress magnitude.

The minimum injection pressure was the first recorded pressure to be equated with the minimum stress magnitude (Hubbert and Willis, 1957). The minimum injection pressure was the pressure required to hold open and extend a fracture, and was known to be less than the breakdown pressure for intact rock. Hubbert and Willis (1957) did not mention leak-off tests or fracture tests as described above, they were talking about hydro-fracs. Since that time, numerous other papers have been published which discuss prediction of fracture pressures, where fracture pressure can mean leak-off pressure or breakdown pressure, (Mathews and Kelly, 1967; Eaton, 1969; Anderson et al.; 1973, Daines, 1982). The fracture pressure in some models (Eaton, 1969; Daines 1982) is simply equated with the minimum horizontal stress magnitude ($\sigma_h$).

The “Zero-tensile-strength-concept” was discussed by Daines (1982) to actually give a rationale for equating leak-off pressure with $\sigma_h$. It is assumed that in any interval of wellbore, there will be natural joints and partings across which the tensile strength is effectively zero and that at least one of these will be oriented near perpendicular to the minimum stress. It must further be assumed that the crack is permeable to the pressurising fluid in the wellbore and that the crack extends far enough away from the wellbore wall so that the pressure required to open it is not influenced greatly by the stress concentration at the wellbore wall. Figure 4.13 shows the idealised leak-off test plot discussed by Daines and interprets the various points on the plot in terms of stress magnitude in the following way:

Assuming that the minimum stress is horizontal, then the wellbore pressure at point B (the leak-off pressure) is equal to $\sigma_h$, and any vertical fractures in the borehole wall which are perpendicular to that stress will have no compressional forces holding them closed. The pressure difference between B and C is that pressure necessary to push fluid into the cracks and to apply pressure to the walls and leading edge of the crack (pressure
needed to overcome frictional losses in the crack). At point D, the pressure at the crack tip is great enough to cause rapid propagation of the fracture, the fracture volume increases, and the wellbore pressure drops rapidly to the fracture propagation pressure (equivalent to the injection pressure of Hubbert and Willis (1957)) which will be equal to, or slightly higher than $\sigma_h$.

![Figure 4.13 An idealised leak-off pressure/volume plot interpreted in the framework of the zero tensile strength concept: point B (leak-off pressure) = $\sigma_h$, point c (crack extension pressure) = leak-off pressure + crack pressure losses, point d (shut-in pressure) = $\sigma_h$. (After Daines, 1982).](image)

The zero tensile strength concept is broadly equivalent to the case of hydro-frac reopening (section 4.5.1) and to the fracture mechanics analysis of hydro-frac breakdown in the case of a large pre-existing permeable crack oriented perpendicular to $\sigma_h$ (section 4.5.2.4).

Published in the same year as the Daines paper is a study of stress with depth in sedimentary basins from around the world by Breckels and van Eekelen (1982) who state that the pressure at which the formation starts taking on fluid (leak-off pressure) can theoretically be anywhere between $2\sigma_h - p$ (the classical breakdown equation when $\sigma_h = \sigma_h$ and tensile strength is zero) and $\sigma_h$ (the case of pre-existing cracks) and that it is therefore the lower end of the leak-off pressure range that is of interest for estimating $\sigma_h$. They go on to compare approximately 180 leak-off pressures from wells around the US Gulf Coast area with approximately 120 $\sigma_h$ values determined from hydro-fracs in the same area. There turns out to be a very close correspondence between the leak-off
pressures and hydro-fract determined $\sigma_h$ in the US Gulf Coast area, with leak-off pressures being only slightly in excess of hydro-fract determined $\sigma_h$ (Figure 4.14).

Breckels and van Eekelen also discuss data from Brunei which has been used to compare leak-off pressures with $\sigma_h$ determined from hydro-fracs. In this data set, the leak-off pressures come from the same tests as the $\sigma_h$ values. There are only 14 tests in which Leak-off pressure can be compared to $\sigma_h$, and on average the leak-off pressure in these tests is 11% higher than $\sigma_h$. This value of 11% appears to be somewhat higher than the difference shown for the US Gulf Coast data. However, if leak-off pressure is taken from the pressure/time plot of the mini-frac test, as is implied, it may, due to the
difference in pressurisation rate, be higher than in a standard leak-off test. Wellbore pressurisation rates, which are generally higher in mini-fracs than leak-off tests, have shown a positive correlation with breakdown pressures in numerous studies (see section 4.5.2.4).

Leak-off pressures have been used on previous occasions as estimates of $\sigma_h$ (Bell and Ervine, 1987; Bell, 1990). Bell (1990) states that from an understanding of hydraulic fracturing, it is reasonable to interpret the leak-off pressure as approximating a fracture opening pressure and that fracture opening pressures exhibit values close to $\sigma_h$, so that leak-off pressures can be considered broadly equivalent to $\sigma_h$.

More recently further empirical relationships have been proposed by comparing leak-off pressure with $\sigma_h$ determined from extended leak-off tests in the Norwegian North Sea. During an extended leak-off test, a pressure decline analysis is performed to determine the fracture closure pressure as an estimate of $\sigma_h$. It has been found that the leak-off pressure is between 6-11% higher than closure pressure in these tests (Amundsen, 1995). Estimates of $\sigma_h$ from closure pressures and re-opening pressures determined during extended leak-off tests performed in Australia have been shown to lie on the lower bound of leak-off pressures from standard leak-off tests (Enever et al., 1996).

Qualitative observations, relevant to the interpretation of leak-off pressures, which describe the form of the pressure/volume plot during a leak-off test and the behavior of the formation being tested have been made by several authors. In particular, tests performed in highly permeable formations have been discussed as well as time dependent effects. Leak-off test “performance” is better in low permeability formations (Daget and Parigot, 1979; Rabia, 1985; Amundsen, 1995). This means that a linear plot of pressure/volume is generally obtained during pressurisation of an impermeable formation during a leak-off test, as little or no fluid enters the formation prior to breakdown.

The shape of the pressure volume plot from a typical test performed in a highly permeable formation is shown in Figure 4.15. The pressure tends towards a horizontal asymptote, indicating that the drilling mud is largely being lost to the formation,
presumably into natural openings of some sort, making a leak-off pressure very difficult
to determine. Often these tests can improve (become more linear) if a second or third
pressurisation cycle is performed. A more impermeable mud cake layer (mud plug) is
built up at the wellbore wall as filtrate from the mud flows into the formation
concentrating the solid component (Daget and Parigot, 1979). Water based muds are
more effective at forming mud plugs than oil based muds as the water more readily
flows into the formation (Aadnoy, 1995).

Leak-off pressures can be higher when an "old" mud is used during the leak-off test
which is thought to indicate that cuttings present in the old, or well used mud, can fill
cracks and gaps in the wellbore wall which prevents fluid penetration prior to
breakdown (Aadnoy, 1995). Similarly, the leak-off pressure can increase if the formation
is exposed to the drilling mud for a long time prior to the test being performed (Aadnoy,
1995).

4.6.5 Theoretical Interpretation of Leak-off Pressures.
The leak-off pressure can now be considered within the various frameworks which
describe the initiation and propagation of tensile fractures at a wellbore wall (section
4.5.2). In this section the leak-off tests are considered to be performed in vertical
boreholes where the vertical stress is a principal stress and the minimum principal stress is horizontal. It should also be noted that where the leak-off pressure is equal to or higher than the vertical stress, it can not be used as an estimate of $\sigma_h$. This can arise if horizontal fractures are opened during the leak-off test such that the vertical stress is being sampled. Similarly, fracture re-opening and closure pressure estimates from hydrofracs are clipped at the value of the vertical stress.

Hydro-frac breakdown pressure is the maximum pressure on the pressure/time plot during a hydro-frac and it corresponds to the point where the fracture is growing in volume at a rate faster than fluid is being pumped into the hole. In the cases where there is no pre-existing crack, and the stress anisotropy is such that the maximum circumferential stress is at the borehole wall (see section 4.5.2.2), there is no fracture opening before breakdown, and the leak-off pressure is equivalent to the breakdown pressure.

Therefore the leak-off pressures (LOP) in porous permeable rock for the case of no pre-existing cracks, are (i) The case of an impermeable wellbore wall:

$$ LOP = 3\sigma_h - \sigma_H - p + T $$

and (ii) the case of a permeable wellbore wall:

$$ LOP = \frac{3\sigma_h - \sigma_H + T - \alpha \frac{1 - 2\nu}{1 - \nu} p}{2 - \alpha \frac{1 - 2\nu}{1 - \nu}} $$

Clearly neither of these relationships are useful for uniquely defining $\sigma_h$, as they both also contain $\sigma_H$. The tensile strength is also an unknown in both equations which can vary a lot between lithologies and is also very much size dependent within any particular lithology (section 6.2.3).

In the case of pre-existing permeable cracks at the wellbore wall, the leak-off pressure may not be equivalent to the breakdown pressure. The breakdown pressure is the pressure at which the fracture propagates very rapidly, however, the crack may begin to
open as soon as the pressure in the crack is equal to the stress perpendicular to it. A small amount of additional pressure may be necessary to apply fluid pressure to the crack tip and to increase the stress intensity at the crack tip to the fracture toughness of the material in order to cause breakdown. In this case, the pressure/time plot would deviate from linearity as the crack begins to open, as it does in the case of crack re-opening (see section 4.5.1), so that the leak-off pressure is equal to the crack opening pressure.

In the case of crack opening before breakdown, the crack opening pressure (leak-off pressure) depends on the permeable length of the crack. If the permeable length of the crack is short, less than the radius of the borehole, the opening pressure will be dependent on $\sigma_H$ and $\sigma_N$, due to the stress concentration near the wellbore wall. For the case of a very short crack the problem is equivalent to either equation 4.25 or 4.26 in which the tensile strength is zero. The maximum leak-off pressure in this case is $2\sigma_N - p$.

As the permeable length of the crack increases to become of the order of the wellbore diameter, the situation converges towards the case of crack re-opening where the crack opening pressure is close to the stress acting across the crack. Further, the breakdown pressure predicted by the fracture mechanics breakdown equation (equation 4.24) given by Rummel (1987), for the case where the permeable crack length is equal to the wellbore diameter, and is perpendicular to $\sigma_N$, can be shown to be only slightly higher than $\sigma_N$. Taking values for the stress intensity functions from Figure 4.9, where the crack length is equal to the wellbore diameter so that $b=3$: $f(b) = -0.75$, $g(b) = 3$ and $h(b) = 3$ assuming pressure at the crack tip is equal to $p_w$. Using these values in equation 4.24 gives:

$$p_b = \sigma_N + \frac{K_{IC}}{3\sqrt{a}} - \frac{1}{4\sqrt{a}}\sigma_H$$

It can be seen from inspection of equation 4.27 that the breakdown pressure in the case of oilwell diameter boreholes is essentially equal to the stress acting across the crack plus the fracture toughness of the rock. This can be illustrated by taking some typical values of stress magnitude, wellbore diameter and fracture toughness. Taking $\sigma_N = 30$
MPa, $\sigma_H = 40$ MPa, $K_{ic} = 1.5$ MPa.m$^{0.5}$ and $a = 0.08$ m, it can be seen from equation 4.27 that the breakdown pressure is 30.77 MPa which is only slightly higher than $\sigma_h$.

If it is assumed that in the case of pre-existing permeable cracks, the leak-off pressure is the crack opening pressure and at least one crack is perpendicular to $\sigma_h$, the leak-off pressure will be between $\sigma_h$ and $2\sigma_h - p$.

The fact that empirical studies (Breckels and van Eekelen, 1982; Amundsen, 1995; Enever, 1996) have shown the leak-off pressure to generally be close to $\sigma_h$ implies that in the relatively large (both in length and diameter) open hole sections in which leak-off tests are conducted, and under the slow pressurisation rates used, there generally exist cracks which are permeable to the drilling mud, and which are of a length approaching the diameter of the borehole as proposed by Daines (1982) in the zero tensile strength concept.

Additional information, which will help the interpretation of how the wellbore wall and surrounding rock responds to pressurisation of the wellbore, can be obtained from examination of the pressure/volume plots recorded during the leak-off test. These are discussed in the next section. This discussion is also continued in section 6.3.2 with regard to the hydro-frac and leak-off datasets from Nirex.

**4.6.6 Interpretation of Leak-off Test Pressure/Volume Plots.**

When pressure/volume plots are available from leak-off tests, they can be used to interpret the behavior of the borehole wall during wellbore pressurisation. In this section the shapes of real leak-off pressure/volume records (from 150 standard leak-off tests from Amerada Hess drilling records and 12 leak-off test from boreholes drilled by UK Nirex of which 8 are double leak-off tests, all of which are described in detail in Chapter 5) are interpreted in terms of crack opening behavior. From such interpretations it is possible to apply the most suitable breakdown or leak-off equation (see section 4.6.5) and thus in some cases to estimate the magnitude of $\sigma_h$.

**4.6.6.1 Standard Leak-off Test Plots (Amerada Hess Drilling Records)**
Figure 4.16 shows 6 types of leak-off pressure/volume plot shapes which are seen in the drilling records. They are labeled (a)-(f) in order of how frequently they occur, (a) being the most frequent.

Figure 4.16 Examples of the various shapes of leak-off test pressure/volume plots found in Amerada Hess drilling records (1989-1995). The plots (a) to (f) are in order of frequency with which they occur in the records (a) being the most frequent and (f) the least.
The "simplest" result is shown in plot type (e) where no leak-off occurs before breakdown. The pressure build up is linear all the way until the formation breaks down. This type of pressure response is more typical of a hydro-frac pressure/time response during breakdown, in which the wellbore section being pressurised has been selected for being free of cracks, the pressurisation rate is high and the fracturing fluid viscous so as to avoid fluid penetration into any cracks that might be present. Leak-off test pressure/volume plots of this type are not common. They are interpreted here as being tests performed in competent, effectively fracture free formations. The leak-off pressure is either that in equation 4.25 or 4.26

By contrast, plot types (a), (b), (c) and (d) all show a linear pressure build up followed by leak-off prior to formation breakdown (or maximum pressure). Plot types (a) and (b) are essentially similar except that in plot (b), more frequent measurements were taken around the leak-off point enabling a curve to be defined rather than just two straight lines. If pumping had continued in either plot (a) or plot (b), the plots would have looked something like either plot (c) or plot (d). As the plot deviates from linearity before the maximum pressure in plots (a) to (d), the leak-off pressure is here interpreted as the pressure at which pre-existing cracks in the wellbore wall begin to open. The increase in pressure above the LOP represents the additional pressure required to force fluid to the crack tip to cause fracture extension. The difference between these pressures has been discussed in terms of hydro-frac re-opening tests in section 4.5.1. The only other possible explanation for this behavior is that the stress anisotropy is such that the hoop stress is not a maximum at the borehole wall (this is true if equation 4.14 is satisfied). Assuming that the horizontal stress field in the North Sea is not that anisotropic, the leak-off pressure in plot types (a) - (d) can be taken as being between $\sigma_h$ and $2\sigma_h-p$, and tending towards $\sigma_h$ for crack lengths approaching the wellbore diameter.

Plot (d) shows a leak-off pressure followed by a constant pressure slightly above the leak-off pressure as the pumping continues. This type of plot is here interpreted as representing the case of fracture opening (the LOP) followed by propagation. As the fracture is likely to propagate perpendicular to $\sigma_h$, the propagation pressure can be taken
as being slightly above $\sigma_h$, around 50-200 psi (Economides and Nolte, 1989). The leak-off pressure in this case is very close to $\sigma_h$.

Plots (a) - (c) would also be expected to look similar (but with a small peak before a plateau propagation pressure) to plot (d) if pumping had continued. The fact that these plot types show a pressure peak indicates that the pressure required to initiate rapid fracture propagation from the borehole wall is slightly greater than the pressure required to continue fracture propagation. Again this could be due to frictional effects encountered in an initially narrow crack.

The interpretation of these plots in terms of crack opening behavior and therefore stress magnitudes, on the basis of the information discussed so far, would be very tentative. However, in section 6.2, similar plots from leak-off test records are examined for cases where the stress magnitudes have been obtained in the same boreholes as hydro-frac stress measurements. It is seen that the above interpretations of leak-off test pressure/volume or pressure/time plots (i.e. that the LOP in plots (a) to (d) represent fracture opening pressures, which for cracks of permeable length that is of the order of the borehole diameter, are slightly higher than $\sigma_h$) can be used to estimate $\sigma_h$ in the rocks of the North Sea.

The type of leak-off test represented by plot (f) is relatively rare. There is no linear portion to the plot and so no leak-off pressure can be picked. This type of response has been interpreted as being indicative of a highly permeable zone into which a lot of fluid is being lost (Daget and Parigot, 1979).

4.6.6.2 Double Leak-off Tests Records (Nirex Drilling Records)
Leak-off tests performed in boreholes drilled by UK Nirex at Sellafield to depths of up to 6000 ft follow essentially the same procedures as those outlined in section 4.6.2. However, in eight out of twelve cases a double leak-off test has been performed. In a double leak-off test, two pressurisation cycles are performed in the same interval. It appears that this is done simply to double check the cementing job and formation strength at the casing shoe. The pressure/volume plot is recorded during both cycles.
During the first cycle, the formation may breakdown (by initiating a new fracture or re-open a pre-existing crack). This can be determined by interpreting if a breakdown pressure is shown on the pressure/volume plot. If the formation does breakdown during the first cycle, then it should be possible to identify a re-opening pressure during the second cycle which can be used as a good approximation of $\sigma_n$ (it is known from hydrofracs at Sellafield that the minimum principal stress is essentially horizontal, therefore, in this case it can be confidently assumed that fractures induced or opened during the leak-off tests will be near vertical).

The double leak-off test pressure/time or pressure/volume plots from the Nirex boreholes show a variety of shapes. The general form of the hydro-frac pressure/time plot shown in Figure 4.6 can be seen in some of the double leak-off tests, indicating either that the wall rock is initially intact, or that any pre-existing cracks are not long enough or permeable enough to approximate zero tensile strength. In other cases, both cycles resemble the re-opening portion of the curve in Figure 4.6 and there is little difference between the two opening pressures, indicating the existence of pre-existing cracks. In either case, the existence of the second pressurisation cycle helps interpretation of the plots greatly. Figure 4.17 shows an idealised pressure/volume plot from a double leak-off test which is interpreted as having been performed in rock with pre-existing cracks. The maximum pressure in the first cycle is slightly higher than in the second. This difference may represent the fracture toughness of the rock in the first pressurisation cycle, or the increase in crack permeability in the second pressurisation cycle.

Other leak-off test plots from the Nirex boreholes show fracture propagation pressures such as in plot type (d) in section 4.6.6.1. The fracture propagation pressure can be considered an upper limit to (but close to) $\sigma_n$. 

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Figure 4.17 Idealised pressure/volume or pressure/time plot (constant pumping rate) from a double leak-off test of the type performed by UK Nirex. In this case, the leak-off pressure in both cases could be interpreted as being close to the re-opening pressure of existing cracks.

The Nirex leak-off test reports contain some set instructions to the drilling engineer on the procedure to follow during the test (section 5.5.1). One of the instructions states that after the peak pressure is reached, the system should be shut in and the pressure decline monitored until it becomes stable. This implies two things. Firstly, that the test is taken to peak pressure as opposed to leak-off pressure means that the intention is to actually cause the formation to break down (section 4.5.2) which obviously has implications for trying to detect a re-opening pressure. Secondly, that the pressure decline is monitored after shut-in means that in theory a closure pressure could also be determined (section 4.5.1). In practice closure pressures can only be guessed at in a few cases as the resolution of the pressure decline data is not high enough to pick subtle changes in pressure gradient. In some cases however it is clear that a fracture has propagated and the pressure decline is then recorded until it becomes constant. If it is assumed that this constant pressure indicates that the fracture has completely closed, it follows that the fracture closure pressure must lie somewhere between this lower bound pressure and the peak pressure. Two of the plots, where pressure vs. time is recorded after breakdown appear to show an inflection point during shut-in. These inflection points might represent the $p_{usip}$ (see section 4.5.1) which is close to $\sigma_h$.  

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Interpreting leak-off test records in terms of $\sigma_h$ values as has been done for the leak-off tests from Sellafield, could normally be considered somewhat speculative. In five of the Sellafield boreholes however, sets of hydro-fracs have been conducted specifically for the purpose of making stress measurements. The interpretations of the leak-off data can therefore be made within the framework of a well defined stress field. The results of the Nirex leak-off test interpretations are compared to the hydro-frac stress measurements in section 6.2.3.

4.7 Estimation of Vertical Stress from Density and Sonic Logs
This section describes how either density logs or a combination of density and sonic logs can be used to estimate the vertical stress due to the weight of rocks in logged sections of borehole. These estimates are used in conjunction with density estimates of the unlogged sections (described below) to determine the total vertical stress at the points where leak-off, or hydro-frac tests have been conducted.

4.7.1 Vertical Stress as a Principal Stress.
The Earth's surface is a free surface and as such can support no shear stress, therefore at the surface the vertical stress must be a principal stress. In most cases, the vertical stress will remain a principal stress at depth. However, as discussed in section 3.3.6.1, in areas where there is significant topography, compressional tectonics or non-vertical fault planes acting as free surfaces, the vertical stress below the surface will not always be a principal stress. In the North Sea basin these effects can be considered as only acting locally, so that generally the vertical stress can be considered a principal stress.

In the absence of the above effects and large local contrasts in elastic properties of the rock (see section 3.3.6.1) the vertical stress will be equal to the weight of the overlying rock per unit area, that is the overburden pressure. The overburden pressure is equal to the product of density gravity and vertical depth below the surface. As the density varies with depth, the vertical stress can be obtained by integrating density with respect to depth from surface level (depth = 0) to the depth (z) of interest ($z = z_i$) then multiplying by gravity ($g$):

$$
\sigma_z = \rho g \int_0^{z_i} z \, dz
$$
The density logs obtained in this study were downloaded from the Amerada Hess wells database where they were stored as measurements at one foot intervals and measured in g/cm³. For ease of calculation, this combination of SI and imperial units has been adopted for the integration of the density logs. The vertical stress is conveniently calculated from a density log which has a density reading in g/cm³ every foot, by simply summing the density readings to the depth of interest and multiplying by a constant, 0.433:

\[ \sigma_v = g \int_{z=0}^{z_1} \rho dz \]

(4.28)

\[ \sigma_v = \sum_{z=0}^{z_1} \rho \cdot 0.433 \]

(4.29)

Where \( \rho \) is measured in g/cm³ at one foot intervals in a vertical hole and \( \sigma_v \) is measured in pounds per square inch (psi). Corrections for deviated holes are straightforward (section 4.7.6).

4.7.2 Well Logs

Geophysical logs (well logs) are run in boreholes by the oil industry, mainly for the purpose of formation evaluation. Formation evaluation means determining the physical properties (such as porosity, interval transit time of sonic waves, density etc.) of rocks by using a variety of geophysical measurements. A measuring tool is lowered into the hole and drawn back to the surface, taking whatever measurement it is designed for as it passes through the rock. Often a whole string of tools are lowered at once so that many measurements can be made in one run.

4.7.2.1 The Density Log

The density log is a continuous record of a formation's bulk density. This is the overall density of the rock including the solid matrix and the fluid in the pores. Quantitatively the density log is used to calculate porosity and indirectly, hydrocarbon density. It is also used to calculate acoustic impedance. Qualitatively, it is a useful lithology indicator, can be used to identify certain minerals, can help to assess organic matter content and may
help to identify over pressure and fracture porosity (Rider, 1986). The logging tool bombards the formation with medium energy, collimated gamma rays. The attenuation between the tool source and detectors is measured. The attenuation is a function of the electron density of the formation (electrons/cm$^3$) which is closely related to the common density of the formation (g/cm$^3$). The change in counts with change in density is exponential over the average logging density range of 2.0 to 3.0 g/cm$^3$. The source and detectors are mounted on a plough shaped pad which is pressed hard against the borehole wall during logging. Modern tools have two detectors, near and far (with respect to the source), which compensate for borehole effects when their readings are combined.

Detector counts in modern tools are converted directly to density. The logging tool is lowered to the bottom of the open hole section and the hole is logged from bottom to top. Sometimes the log remains running into the cased section above the open hole. Log readings from cased holes are spurious as they are in part a measure of the density of the steel casing. As the depth of the casing shoes is known, it should be easy to splice sections of the log together to get a continuous log throughout the length of the logged hole (sometimes this is only one section anyway). However care must be taken when integrating the density logs as in many cases readings which are obviously spurious are still present. It is not difficult to find such readings if the depths around the casing shoes are inspected. These sections are typically of lengths around 50 to 100 feet. The spurious values should be replaced by an average of the density in the 100 feet or so either side of the casing shoe.

The 30 and 20 inch hole sections, which are the first sections to be drilled are nearly always cased directly after drilling. They typically extend to around 1000 feet below sea bed in the central North Sea. The density log is sometimes run (presumably by accident) to the surface through this section, but its readings do not reflect the density of the formation. Inspection of logs run in such sections quite clearly show the dubious nature of the readings. Typically they will give low readings as the tool is sampling the mud density but often suddenly jump up to very high readings as the tool brushes past the steel casing. The bulk density of the top 1000 ft or so of sediment must be estimated by other means (see section 4.7.5).
4.7.2.2 The Sonic Log.

The sonic log records a formation's interval transit time ($\Delta T$) which is the reciprocal of the sonic velocity. It is a measure of the formation's capacity to transmit sound waves. The sonic log is used to help evaluate porosity. It is also used as an aid to seismic interpretation as it gives interval velocities and velocity profiles, and can be calibrated with the seismic section. Qualitatively the log indicates lithology, overpressure and to some extent fractures. It is frequently used in correlation (Rider, 1986). The sonic tool measures the time it takes for a P-wave to travel from the emitter at one end of the logging tool to the receiver at the other end. The emission frequency of the tool is between 20-40 KHz. The tool takes an average of one reading per 8 cm. Modern sonic tools consist of a double array of sonic pulse emitters and receivers, each array consisting of one emitter and two receivers. This arrangement compensates for borehole effects. The tools are run hole centred so the pulse radiates symmetrically about the tool and the measurements come from all sides of the hole simultaneously. The interval transit times are generally recorded in micro-seconds per foot, with values usually ranging from about 40 to 140 $\mu$s/ft.

The sonic log cannot be run through casing, therefore the same care must be taken as with the density logs to avoid spurious readings above casing shoes.


For any particular rock type, there is a linear relationship between interval transit time and density. Therefore, if both sonic and density logs are available for a particular formation in a few wells, a relationship between interval transit time and density can be established. This relationship can then be applied to other wells where only the sonic log has been run and the formation tops and bottoms are known.

A formation is a lithostratigraphic unit. There are numerous formations of fairly local extent (see section 2.3), although formations may be much more widespread such as the Kimmeridge Clay Formation. However, within one formation, the lithologies will be similar from well to well and so the density/sonic relationship established in wells where both density and sonic logs exist, can be used to estimate density where only the sonic log has been run.
Figure 4.18 shows the density/interval transit time for the Kimmeridge Clay Formation from well 15/21a-44. The line of best fit which defines the density/interval transit time relationship from 15/22-d3 and 15/21a-43 are also shown on this plot.

The equation of the best fit line is simply of the form density = mΔT + c where m is the gradient of the line and c is the intercept with the density axis. In the case of the Kimmeridge Clay the relationship is density = 3.376 - ΔT/98. This relationship is applied to the sonic log (which is in spread sheet form) to conveniently convert interval transit times into densities for that particular formation.

Several thousand feet of borehole may well consist of a dozen or so formations. Finding other wells from which to establish sonic/density relationships, plotting several of these and establishing reasonably what the relationship is, applying this to the sonic log of the well in question and also making corrections for well deviation (where the well is not
vertical) can be time consuming. However once the relationships between sonic and density readings have been established, they can be applied to more than one well, and the process speeds up.

4.7.4 Estimating Density Using Offset Well Data.
Sometimes neither the sonic or the density log has been run. This is often the case in the shallower sediments and in development wells drilled in areas where the lithostratigraphy is well known, and a particular target is being drilled for, the depth of which is fairly confidently known. In such cases the only information available may be the formation tops and bottoms. The best way of estimating density in such wells is to find an offset well (one that has been drilled in close proximity) that does have a density log. If the formations in the wells are the same (which they often would be) and the tops and bottoms are at roughly the same depths (this must be checked as a large fault between the two wells could mean that there are significant differences) then an average density for the formations in the well that has been logged can be calculated and used for the unlogged well.

The top few thousand feet of many wells in the North Sea is composed of Cenozoic sediments which are neither logged nor differentiated into formations. In these cases the only thing that can be done is to look at the average density of the shallow sediments in offset wells which are likely to have a similar stratigraphy. Good candidates are wells which are geographically close and that have deeper horizons at the same depth as the well in question. For example if the top of the Cretaceous is at the same depth in both wells and they are not far (perhaps a few kilometres) apart, it is reasonable to assume that their Tertiary and Quaternary stratigraphy are similar.

4.7.5 Estimating Density of Shallow Unlogged Sediments.
As the first casing strings are set immediately after drilling, the top 2000 ft MD (measured depth below rotary table) is not logged and so the density of the sediments in this top most section is not known in any of the wells. The method used to estimate the density of these sediments applies to the central and northern North Sea where there are large thicknesses of Cenozoic sediments. The method involves firstly estimating the
density of typical lithologies (sand and shale) at a vertical depth of 1500 ft below sea floor (BSF), which is the shallowest depth for which reliable logged information is available. Secondly, with a knowledge of the density of mineral grains and pore fluid, the porosity of the rock is calculated. A model of compaction (Sclater and Christie, 1980) can then be applied to extrapolate to the porosity to any depth up to the sea floor where the effective stress is zero and zero compaction has occurred. From these porosities, the bulk densities of the lithology in question can be calculated.

The shallowest density logs in this study indicate a density of between 2.01 and 2.07 for sediments at 1500 ft BSF in the central North Sea.

Sclater and Christie (1980) present curves of porosity with depth for the North Sea and other areas which have been constructed using log derived estimates and surface measurements. Their North Sea curves show the porosity of shale at the surface and down to below 1500 ft BSF to be higher than that of sand, although they show data from other areas which has previously been compiled by other workers which does not agree with this. Using the relationship:

\[ \rho(\text{bulk}) = \phi_p \rho(\text{fluid}) + (1-\phi_p) \rho(\text{grains}) \]  

where \( \phi_p \) is porosity, the porosities shown at 1500 ft BSF give values of bulk density for shale of 1.885 g/cm\(^3\) and 1.978 g/cm\(^3\) for sand. These are low compared to the values obtained from the few density logs obtained in this study. However, on the basis that the shale density is predicted to be lower at this depth, the densities measured on the density logs at the depth of 1500 ft BSF are interpreted here as \( \rho(\text{shale}) = 2.01 \) g/cm\(^3\) and \( \rho(\text{sand}) = 2.07 \) g/cm\(^3\).

Using the exponential porosity/depth relationship (Sclater and Christie, 1980):

\[ \phi = \phi_m e^{-cy} \]  

where \( \phi_m \) is the surface porosity, \( y \) is the depth below sea floor and \( c \) is a coefficient which varies depending on the sediment \( c = 0.51 \) for shale and \( c = 0.27 \) for sand, porosities and therefore densities of sand and shale can be estimated at any depth.
between 1500 ft BSF and sea floor. Assuming the sediments to be 50% sand and 50% shale, and taking values for the grain and fluid densities: \( \rho(\text{shale grain}) = 2.72 \ \text{g/cm}^3 \), \( \rho(\text{sand grain}) = 2.65 \ \text{g/cm}^3 \), \( \rho(\text{fluid}) = 1.05 \ \text{g/cm}^3 \), the average density of sediment in the top 1500 ft BSF in the North Sea is 1.98 g/cm³.

4.7.6 Correcting for Well Deviation.

When the well is deviated, that is when it is not vertical, it is not possible to simply sum the density readings. For most of the deviated wells, deviation surveys are available. These have also been down-loaded from the wells database. They list the measured depth, which corresponds to the depth in the logs, and the angle of deviation from vertical (\( \varphi \)).

The density readings are all multiplied by the cosine of the deviation at that point. Then they are summed and multiplied by 0.433 to give the vertical stress.

\[
\sigma_v = 0.433 \sum_{z=1}^{z=Z} \rho \cos \varphi 
\]

Where the vertical stress is measured in psi and the density in g/cm³.

When deviation surveys are not available, the summed density for a particular interval, the top and bottom depths of which are known both in measured depth and true vertical depth, can be corrected by multiplying by the cosine of the average angle of deviation for that interval which is equal to:

True bottom depth - true top depth / measured bottom depth - the measured top depth.

This method is not as accurate as correcting each foot individually, as the angle of deviation is not always constant through a given formation. In a given section that has been corrected as one unit, the parts that are deviated by an angle less than the average angle for the whole section will be "over corrected" and those that are deviated by an angle more than the average will be "under corrected".

4.7.7 Final Calculation of Vertical Stress
A schematic representation of a vertical stress estimate in a vertical borehole, which summarises the methods described above and shows how they are used in conjunction to obtain the total vertical stress, is shown in Figure 4.19.

The vertical stress is often expressed as a kind of gradient, usually called the overburden gradient. The overburden gradient is simply the vertical stress, or overburden pressure, divided by the vertical depth at the point of interest. The overburden gradient, because of compaction, obviously increases with depth. The rocks in this study all have overburden gradients of between 0.95 and 1.05 psi/ft.

In the northern and central North Sea, there are thick layers of Cenozoic sediments, many of which are as yet probably fairly unconsolidated and which therefore compact rapidly with depth. The overburden gradient in these areas is seen to increase rapidly with depth. In the southern North Sea, there are generally no such thick sequences of unconsolidated sediments and so the overburden gradient does not increase as rapidly with depth. Jurassic and Cretaceous sediments can be found forming the sea floor and shallow rocks in the southern North Sea. For this reason, care must be taken if the methodology described above to estimate the densities of shallow rocks is used in the southern North Sea.
Figure 4.19 Schematic representation of some of the elements used to calculate the overburden pressure. The calculation uses whatever information is available for each section and then sums the sections to obtain the total overburden.

Overburden Pressure at 13850 ft (below sea surface) = 13699 psi
Overburden gradient = 0.99 psi/ft
Chapter Five - Data Used in this Study

5.1 Introduction
This chapter describes the type of data used in this study and the various sources of the data. The data falls into three groups:

1. North Sea Leak-off test data from Petroleum Information (Erico) Ltd.
2. North Sea leak-off test data from Amerada Hess Ltd.
3. Combined leak-off test and hydraulic fracturing data from the North Sea (Dowell Schlumberger data) and from Sellafield (UK Nirex data).

5.2 Erico North Sea Pressure Studies
Petroleum Information (Erico) Ltd. is a private company which compiles a variety of data from the petroleum industry. The data comes from so called "released wells", that is wells for which the information has been released into the public sector by the relevant government body. In the case of wells in the British sector of the North Sea, this is the Department of Trade and Industry. The data has been compiled from the records of approximately 50 oil companies operating in the North Sea. It is assembled in a format useful to the oil industry and then sold back to individual oil companies.

The main part of the Erico pressure studies from the North Sea consists of formation pore pressure information. The wells included in the study are therefore wells in which formation pore pressure measurements have been made and where this pressure information has been released in a format of sufficiently high quality to be useful.

Information pertaining to leak-off tests and occurrences of lost circulation is also included in the data for most wells. This information has been compiled by Erico and is included in the pressure database.

Erico's compilation of oil industry pressure data is an on-going job. The data discussed in this chapter are those from the pressure studies which were available for analysis within the time scale of this project.
Other released information such as (i) depths of formation tops (ii) casing depths and (iii) lithologies, are also included in the pressure database. All the depths in the database are true vertical depths (TVDs) and are measured below KB (Kelly Bushing - part of the drilling apparatus which is found on the floor of the drilling platform). The information is all available in several large bound volumes and also in digital form. The digital copy of the data which was obtained from Erico took the form of ASCII files. These files were imported into a relational database (Paradox). The data had already been grouped by a well number which was assigned by Erico. It was then further grouped into the following tables as part of this study:

*Well header information:* This includes the well name (North Sea quadrant, block and well number), operator (which oil company operated the well), location, spud (start of drilling) and completion dates, total depth of the well, KB elevation, water depth, units used to record depths (feet or metres) and mud weights (lbs/gallon or grams/cc) and whether or not the well is deviated (drilled non-vertically).

*Formation tops and comments:* The depths to the tops of the major lithostratigraphic and chronostratigraphic units are recorded together with the depths to the tops of other locally important horizons.

*Formation pressures:* These are static formation pressures recorded in psi (pounds per square inch) from FITs (formation interval tests), RFTs (repeat formation tests), DSTs (drill stem tests), production tests and kicks.

*Leak-off test and lost circulation pressures:* Leak-off test data is expressed as a mud weight equivalent and the depth to the casing shoe. The pressures are assigned as leak-off pressures (LO, or LOP) if there is some evidence in either the drilling report or the final well report to indicate that leak-off occurred (pers. com., S. Thomas, Erico). The pressures are assigned as leak-off to or limit test pressures (LT, or LTP) if the available data does not indicate that leak-off occurred. If the data does not indicate whether leak-off occurred or not, the pressures are assumed to be limit test pressures.
Occurrences of open hole lost circulation during drilling are also recorded in the table. The depth to which the hole had been drilled (assumed to be the depth at which the lost circulation occurred) is recorded together with the mud weight being used for drilling at the time, so that the lost circulation pressure (LCP) can be calculated.

The lithology in which the leak-off test was performed or the lost circulation occurred is also recorded where such information is available (on average, the lithologies are listed for approximately half of all the leak-offs and lost circulations listed in the database). Erico has assigned a number to each of the main lithologies:

1- Shale, 2- Silt, 3- Sand, 4- Limestone, 5- Chalk, 6- Marl, 7- Dolomite, 8- Anhydrite, 9- Salt.

In many of the cases where lithology is recorded for the leak-off test, the lithology is mixed, such as sandy shale, which would be recorded as 31, with the main lithology being the second number. Silty marl would therefore be recorded as: 26.

The leak-off test tables also contain a comment column which usually just contains the name of the formation in which the test was performed. Occasionally however, some further information might be included such as the number of barrels of mud that were pumped during the test.

*Casing*: The casing size is recorded in decimal inches. The top and bottom depth of the casing is also recorded.

*Mud weights*: The mud weights used to drill each record section of hole are recorded. The top and bottom depth of each section is also recorded.

*Salinities*: The salinity of the formation pore fluid is recorded in parts per million NaCl. This data is only recorded when a good formation water flow was established and is only available for about 10% of the wells.
5.3 Areas Covered by the Erico Pressure Studies

There are 3 Erico Pressure Studies which cover the North Sea, Pressure studies 1, 2 and 5. However, for the purpose of this study, the data contained in the Erico Pressure Studies has been divided into three structural/geographic domains. Norwegian

From a basic understanding of crustal stress (see chapter 3) and from previous work in the North Sea (Cowgill et al., 1993; Cowgill, 1994; Brereton and Muller, 1991) it seems that the stress field can be influenced by structural setting. For this reason, the domains outlined below, are defined on the basis of the dominant structural elements of the North Sea (Figure 5.1). Figure 5.1 has been adapted from Figure 2.2 to show more clearly the division of the North Sea into national sectors (e.g. British Sector, Norwegian Sector) and the subdivision of these sectors into quadrants. Each national sector has its own quadrant numbering system. As the main structural elements of the North Sea are largely contained within 3 broad geographic regions, a first order division of the data can be defined by the following 3 geographic domains:

- The northern North Sea (includes the northern and southern Viking Graben, the East Shetland Basin) and is covered by Erico Pressure Studies 1 and 2.

- The central North Sea (includes British and Norwegian Central Graben, Witch Ground Graben and Inner Moray Firth) is also covered by Pressure Studies 1 and 2.

- The British sector of the southern North Sea is covered by Pressure Study 5.

Table 5.1 summarises the regions covered and the relevant volumes of Erico Pressure Study. This includes the number of wells that contain leak-off or lost circulation data and the number of leak-off test or lost circulation data.
The next sections describes the data obtained for each of the geographic domains listed in Table 5.1. Chapters 6 and 7 investigate some of the differences in the stress field between these geographic domains.

### 5.3.1 The Erico Central North Sea Data.

The area described here as the central North Sea can be largely defined in a structural context by the graben boundaries of the Inner Moray Firth, the Witch Ground Graben, the South Halibut Basin and the Central Graben (Figure 5.1). Wells which lie within these grabens are grouped together here as the central North Sea geographic domain. There is no data in the Erico Pressure Studies listed above for the Central Graben south of 56° latitude. There is little drilling activity in the Central North Sea outside of the major grabens. A few wells however are found on the graben flanks and in smaller basins such as the Forth Approaches Basin, and data from these wells has been included in the central North Sea geographic domain.

Table 5.2 shows the number of leak-off, limit test and lost circulation pressures from the Erico Pressure Studies in each quadrant of the central North Sea. The position of each of these quadrants with respect to the major structural features of the North Sea are shown in Figure 5.1.

<table>
<thead>
<tr>
<th>Geographic Domain</th>
<th>Pressure Study</th>
<th>Number of Wells Containing LOPs, LTPs or LCPs.</th>
<th>Number of LOPs, LTPs and LCPs.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Northern North Sea</td>
<td>Pressure Studies 1 and 2</td>
<td>371</td>
<td>975</td>
</tr>
<tr>
<td>Central North Sea</td>
<td>Pressure Studies 1 and 2</td>
<td>496</td>
<td>1270</td>
</tr>
<tr>
<td>Southern North Sea (British Sector)</td>
<td>Pressure study 5</td>
<td>232</td>
<td>676</td>
</tr>
</tbody>
</table>

Table 5.1 Erico Data from the North Sea.
Note that the Norwegian North Sea has its own quadrant numbering system (Figure 5.1). Norwegian quadrants listed in Table 5.2 are therefore suffixed (NOR).

<table>
<thead>
<tr>
<th>North Sea Quadrant</th>
<th>Erico Pressure Study</th>
<th>Number of Leak-off, Limit test and Lost Circulation Data</th>
</tr>
</thead>
<tbody>
<tr>
<td>11</td>
<td>1</td>
<td>6</td>
</tr>
<tr>
<td>12</td>
<td>1</td>
<td>23</td>
</tr>
<tr>
<td>13</td>
<td>1</td>
<td>28</td>
</tr>
<tr>
<td>14</td>
<td>1</td>
<td>74</td>
</tr>
<tr>
<td>15</td>
<td>1</td>
<td>162</td>
</tr>
<tr>
<td>16/26</td>
<td>1</td>
<td>20</td>
</tr>
<tr>
<td>20</td>
<td>1</td>
<td>80</td>
</tr>
<tr>
<td>21</td>
<td>1</td>
<td>324</td>
</tr>
<tr>
<td>22</td>
<td>1</td>
<td>90</td>
</tr>
<tr>
<td>23</td>
<td>1</td>
<td>40</td>
</tr>
<tr>
<td>26</td>
<td>1</td>
<td>7</td>
</tr>
<tr>
<td>28</td>
<td>1</td>
<td>9</td>
</tr>
<tr>
<td>29</td>
<td>1</td>
<td>88</td>
</tr>
<tr>
<td>30</td>
<td>1</td>
<td>160</td>
</tr>
<tr>
<td>31</td>
<td>1</td>
<td>20</td>
</tr>
<tr>
<td>1 (NOR)</td>
<td>2</td>
<td>30</td>
</tr>
<tr>
<td>2 (NOR)</td>
<td>2</td>
<td>72</td>
</tr>
<tr>
<td>3 (NOR)</td>
<td>2</td>
<td>3</td>
</tr>
<tr>
<td>7 (NOR)</td>
<td>2</td>
<td>31</td>
</tr>
<tr>
<td>8 (NOR)</td>
<td>2</td>
<td>2</td>
</tr>
<tr>
<td>9 (NOR)</td>
<td>2</td>
<td>1</td>
</tr>
</tbody>
</table>

Total = 1270

Table 5.2 Erico Pressure Studies data from the central North Sea. Norwegian wells are denoted by the suffix (NOR).
Figure 5.1 An outline of the main structural elements of the North Sea and the division of the North Sea into national sectors and quadrants. Each quadrant is one degree longitude by one degree latitude. The quadrant numbers in the British and Norwegian sectors (the sectors included in this study) have been highlighted for clarity, as have the sector boundaries (thick dashed line). (Adapted from Glennie, 1990)
5.3.2 The Erico Northern North Sea data.

The area described here as the northern North Sea geographic domain is largely defined in a structural context by the graben boundaries of the southern and northern Viking Graben and the East Shetland Basin. The vast majority of the northern North Sea wells have been drilled within these grabens, and they are grouped together here as the northern north Sea geographic domain. Erico Pressure Study 2 also contains a much smaller amount of data from wells which have been drilled on the flanks of these grabens. This data has been included here in northern North Sea geographic domain.

Table 5.3 shows the number of leak-off, limit test and lost circulation pressures from the Erico Pressure Studies in each quadrant of the northern North Sea. The position of each of these quadrants with respect to the major structural features of the North Sea are shown in Figure 5.1.
<table>
<thead>
<tr>
<th>North Sea Quadrant</th>
<th>Erico Pressure Study</th>
<th>Number of LOPs, LTPs or LCPs.</th>
</tr>
</thead>
<tbody>
<tr>
<td>2</td>
<td>2</td>
<td>28</td>
</tr>
<tr>
<td>3</td>
<td>2</td>
<td>107</td>
</tr>
<tr>
<td>9</td>
<td>2</td>
<td>50</td>
</tr>
<tr>
<td>16 (excluding 16/26)</td>
<td>1</td>
<td>307</td>
</tr>
<tr>
<td>210</td>
<td>2</td>
<td>9</td>
</tr>
<tr>
<td>211</td>
<td>2</td>
<td>112</td>
</tr>
<tr>
<td>15 (NOR)</td>
<td>2</td>
<td>57</td>
</tr>
<tr>
<td>16 (NOR)</td>
<td>2</td>
<td>14</td>
</tr>
<tr>
<td>17 (NOR)</td>
<td>2</td>
<td>3</td>
</tr>
<tr>
<td>24 (NOR)</td>
<td>2</td>
<td>5</td>
</tr>
<tr>
<td>25 (NOR)</td>
<td>2</td>
<td>11</td>
</tr>
<tr>
<td>29 (NOR)</td>
<td>2</td>
<td>5</td>
</tr>
<tr>
<td>30 (NOR)</td>
<td>2</td>
<td>66</td>
</tr>
<tr>
<td>31 (NOR)</td>
<td>2</td>
<td>60</td>
</tr>
<tr>
<td>33 (NOR)</td>
<td>2</td>
<td>5</td>
</tr>
<tr>
<td>34 (NOR)</td>
<td>2</td>
<td>80</td>
</tr>
<tr>
<td>35 (NOR)</td>
<td>2</td>
<td>17</td>
</tr>
</tbody>
</table>

Total = 975

Table 5.3 Erico Pressure Studies data from the northern North Sea. Norwegian wells are denoted by the abbreviation (NOR).

5.3.4 The Erico southern North Sea data.

All the Erico pressure data from the southern North Sea is contained within Pressure Study 5. It consists only of data from the British sector of the southern North sea. The area described here as the southern North Sea geographic domain is not so clearly defined in a structural context as the domains in the northern and central North Sea. The southern North Sea, unlike the northern and central North sea, is not characterised by deep, incisive, Mesozoic grabens. Instead, the hydrocarbon deposits lie in a broadly
NW-SE trending belt, which is part of the southern Permian basin (see section 2.3.4). As a consequence, the majority of the data comes from this belt which includes the Silver Pit Basin, the Sole Pit High and the Sole Pit Basin.

Table 5.4 shows the number of leak-off, limit test and lost circulation pressures from the Erico Pressure Studies in each quadrant of the northern North Sea. The position of each of these quadrants with respect to the major structural features of the North Sea are shown in Figure 5.1.

<table>
<thead>
<tr>
<th>North Sea Quadrant</th>
<th>Erico Pressure Study</th>
<th>Number of LOPs, LTPs or LCPs. 29/5b-2</th>
</tr>
</thead>
<tbody>
<tr>
<td>36</td>
<td>5</td>
<td>2</td>
</tr>
<tr>
<td>37</td>
<td>5</td>
<td>3</td>
</tr>
<tr>
<td>38</td>
<td>5</td>
<td>7</td>
</tr>
<tr>
<td>41</td>
<td>5</td>
<td>2</td>
</tr>
<tr>
<td>42</td>
<td>5</td>
<td>41</td>
</tr>
<tr>
<td>43</td>
<td>5</td>
<td>25</td>
</tr>
<tr>
<td>44</td>
<td>5</td>
<td>93</td>
</tr>
<tr>
<td>47</td>
<td>5</td>
<td>64</td>
</tr>
<tr>
<td>48</td>
<td>5</td>
<td>247</td>
</tr>
<tr>
<td>49</td>
<td>5</td>
<td>163</td>
</tr>
<tr>
<td>50</td>
<td>5</td>
<td>6</td>
</tr>
<tr>
<td>53</td>
<td>5</td>
<td>22</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Total = 675</td>
</tr>
</tbody>
</table>

Table 5.4 The Erico pressure studies data from the southern North Sea.

5.4 Leak-off Test Records, Drilling Reports, Final Well Reports and Well Logs
The information pertaining to leak-off tests in The Erico Pressure Studies consists of (i) single values for the leak-off or limit test pressure (ii) the depth of the test (iii) the lithology in which it was performed and (iv) whether or not leak-off was achieved. Due to the nature of the leak-off test procedure, as discussed in section 4.6, there can often be a degree of subjectivity involved in determining firstly, whether or not leak-off has
occurred, and secondly what the leak-off pressure actually is. For these reasons the actual drilling records (which contain the leak-off test records) have been obtained for as many of the Erico Pressure Study wells as possible. Drilling records were made available by Amerada Hess and BP. Given the difficulties involved in obtaining leak-off test records and associated well information (discussed below), and the limited time available for this project, it was not practical to spend time searching for further records.

In the vast majority of cases, leak-off test records, particularly those which contain the pressure/volume plot, were not available. There are two reasons for this. Firstly, for the majority of North Sea operators, it has only been a recent development in drilling practice to make a systematic record of events during a leak-off test. Secondly, once the well is drilled, the leak-off test records have usually been archived or discarded. Obtaining leak-off test records for wells of a few years old, is therefore not a trivial task.

Where leak-off test records have been available for wells which are also in the Erico Pressure Studies, a comparison has been made of the values of leak-off pressure listed in the pressure studies, and those which have been picked from examination of the leak-off test records. This is essentially an exercise in quality control of the Erico data.

There are a total of 52 leak-off tests from 26 wells in the Erico Pressure studies for which drilling records have been obtained. However, only 25 of these tests have pressure/volume plots included in the drilling report. From these pressure/volume plots, it has been determined on the basis of the leak-off test description set out in section 4.6.3, whether or not leak-off has occurred (whether the test is a leak-off test or a limit test), and what the leak-off or limit test pressure is. Table 5.5 summarises the comparison of data from leak-off test records and Erico data.
<table>
<thead>
<tr>
<th>Well Name</th>
<th>LOP picked from record (EMW/ppg)</th>
<th>LO or LT (picked from record)</th>
<th>LOP from Erico (EMW/ppg)</th>
<th>LO or LT (Erico Listed)</th>
<th>Difference in LOP (% of Erico Value)</th>
</tr>
</thead>
<tbody>
<tr>
<td>15/21a-7</td>
<td>15.03</td>
<td>LT</td>
<td>15.00</td>
<td>LT</td>
<td>+0.2%</td>
</tr>
<tr>
<td>15/21a-7</td>
<td>10.10</td>
<td>LO</td>
<td>10.10</td>
<td>LO</td>
<td>0</td>
</tr>
<tr>
<td>15/21a-7</td>
<td>13.20</td>
<td>LO</td>
<td>11.60</td>
<td>LO</td>
<td>+13.8%</td>
</tr>
<tr>
<td>15/21a-12</td>
<td>12.86</td>
<td>LO</td>
<td>13.7</td>
<td>LO</td>
<td>-6.5%</td>
</tr>
<tr>
<td>15/21a-8</td>
<td>13.90</td>
<td>LO</td>
<td>13.90</td>
<td>LO</td>
<td>0</td>
</tr>
<tr>
<td>15/21a-9a</td>
<td>13.96</td>
<td>LO</td>
<td>13.20</td>
<td>LT</td>
<td>+5.7%</td>
</tr>
<tr>
<td>15/21a-10</td>
<td>14.00</td>
<td>LO</td>
<td>14.20</td>
<td>LO</td>
<td>-1.4%</td>
</tr>
<tr>
<td>15/21a-11</td>
<td>13.90</td>
<td>LT</td>
<td>13.60</td>
<td>LO</td>
<td>+2.2%</td>
</tr>
<tr>
<td>47/14a-6</td>
<td>13.20</td>
<td>LT</td>
<td>13.20</td>
<td>LT</td>
<td>0</td>
</tr>
<tr>
<td>47/14a-6</td>
<td>17.23</td>
<td>LT</td>
<td>17.20</td>
<td>LT</td>
<td>+0.17%</td>
</tr>
<tr>
<td>47/14a-9</td>
<td>12.90</td>
<td>LT</td>
<td>12.90</td>
<td>LT</td>
<td>0</td>
</tr>
<tr>
<td>31/27-1</td>
<td>14.26</td>
<td>LO</td>
<td>14.80</td>
<td>LO</td>
<td>-0.8%</td>
</tr>
<tr>
<td>31/27-1</td>
<td>14.00</td>
<td>LO</td>
<td>14.30</td>
<td>LO</td>
<td>0</td>
</tr>
<tr>
<td>31/27-1</td>
<td>17.04</td>
<td>LO</td>
<td>17.14</td>
<td>LT</td>
<td>-0.6%</td>
</tr>
<tr>
<td>29/5b-2</td>
<td>15.29</td>
<td>LO</td>
<td>15.26</td>
<td>LO</td>
<td>+0.2%</td>
</tr>
<tr>
<td>29/5b-2</td>
<td>14.00</td>
<td>LT</td>
<td>14.00</td>
<td>LT</td>
<td>0</td>
</tr>
<tr>
<td>29/5b-2</td>
<td>10.87</td>
<td>LO</td>
<td>10.90</td>
<td>LO</td>
<td>-0.3%</td>
</tr>
<tr>
<td>26/12-1</td>
<td>15.03</td>
<td>LT</td>
<td>15.02</td>
<td>LT</td>
<td>+0.07%</td>
</tr>
<tr>
<td>26/12-1</td>
<td>15.06</td>
<td>LT</td>
<td>15.01</td>
<td>LT</td>
<td>+0.3%</td>
</tr>
</tbody>
</table>

Table 5.5. Comparison of LOPs and LTP obtained directly from drilling records and from Erico Pressure Studies.
It can be seen from Table 5.5 that the LOPs and LTPs reported by Erico are generally in very good agreement with those which have been picked from the leak-off test records. In only one case, that of the leak-off test in well 15/21a-7, is there any significant disagreement. In 4 cases, Erico list tests as limit tests, whereas from examination of the leak-off test records, it appears that leak-off has occurred. In these cases, the amount of leak-off is small, i.e. the test has been stopped very soon after deviation from linearity of the pressure/volume plot. Erico state that they tests as limit tests, if there is nothing to suggest that leak-off occurred. It appears that they do not consider a small amount of leak-off apparent in these tests as being sufficient evidence that leak-off occurred. These tests serve to illustrate the subjective nature of interpreting leak-off tests. In only one case do Erico list a test as a leak-off test, where I have found no clear evidence of leak-off. The generally good agreement between the Erico data and the original “raw” data, adds confidence to the use of the Erico dataset for estimating stress magnitudes in the North Sea (Chapters 6 and 7).

During the process of searching for the Amerada Hess drilling records, it was discovered that since 1989, Amerada Hess had been recording leak-off tests in a much more thorough and systematic way. It was decided to obtain as many of these recent drilling records as possible. Although these wells, having been drilled fairly recently, were not contained in the Erico Pressure Studies, the quality of the data, and the availability of other well data from the same wells (see below), was sufficient to justify the creation of a smaller but more detailed dataset consisting of all wells drilled by Amerada Hess, in the North Sea, between 1989 and 1995.

Further information about each Amerada Hess well (drilled between 1989 and 1995) has been obtained from Final Well Reports (FWRs). Geophysical logs (well logs: gamma ray, sonic and density), have also been obtained from the Hess Wells Database for the wells where leak-off test records are available. The data obtained directly from Amerada Hess is described in the following sections.
5.4.1 Amerada Hess Drilling Records/Leak-off Test Reports.

The essentials of the leak-off test procedure have been discussed in section 4.6. In 1989, Amerada Hess introduced a standard leak-off test report form to be completed by the supervising drilling engineer during a leak-off test. Prior to this, the information obtained during the leak-off test seems only to have been occasionally recorded and even then the amount of information for each test was sparse. The introduction of the leak-off test pro-forma has encouraged the recording of more information during the test. This enables a more objective decision to be made about whether or not leak-off has occurred, and what the leak-off pressure might be.

Figure 5.2 shows an example of the Amerada Hess leak-off test pro-forma. The most important information contained in the report is the plot of pressure against volume pumped (pressure/volume plot) which can be used to determine if leak-off has occurred and if so what the leak-off pressure is. In some cases the pressure/volume plot can be interpreted within the framework outlined in section 4.6 to make a more informed estimate of minimum horizontal stress magnitude. The pressure on the pressure/volume plot is the surface pump pressure. To obtain the actual pressure in the open hole at leak-off, the surface pump pressure must be added to the pressure due to the weight of drilling mud in the borehole. This is simply the mud weight in pounds per gallon multiplied by the vertical depth from the drilling platform to the open hole section multiplied by a constant. In most cases this calculation has been performed by the drilling engineer, and is recorded in the leak-off test record as an equivalent mud weight.

The Amerada Hess leak-off test pro-forma also records the volume of fluid pumped and the volume returned after pumping has stopped. The difference between the two is the volume of fluid that is lost to the formation. It would be expected that for the same volume of open hole being tested in the same lithology, more fluid would be lost in the case where leak-off occurs and a tensile fracture is propagated into the formation. In many cases, it can be seen from the pressure/volume plot that no leak-off has occurred and no fluid been lost to the formation (Figure 5.3). In other cases it can be seen from the pressure volume plot that leak-off has occurred and pumping has continued for some time after leak-off. In these cases, nearly all the fluid can be lost (Figure 5.4). The
amount of fluid lost to the formation can therefore add confidence to the interpretation of the leak-off test record.
Figure 5.2 A typical leak-off test, recorded on the Amerada Hess leak-off test pro-forma.
Figure 5.3 An example of a leak-off test record in which leak-off has not occurred. The pressure/volume plot does not deviate from linearity, and no fluid is lost to the formation.
Figure 5.4 An example of a leak-off test record in which leak-off has occurred. The pressure/volume plot clearly deviates from linearity, and in this case pumping has continued beyond leak-off resulting in a substantial loss of fluid to the formation.
Information relating to the open hole section is also recorded in the Amerada Hess leak-off test record. The measured depths and true vertical depths of both the casing shoe and the bottom of the open hole are recorded as is the diameter of the casing shoe. This enables both the volume and surface area of the open hole to be calculated. If the open hole section is very large, it might affect the amount of fluid leak-off and the rate of pressure build up during the leak-off test. This is also dependent on the lithology of the formation in the open hole.

In many cases, the lithology in the open hole section is included in the Amerada Hess leak-off test record. It is clear however from a consideration of the length of the open hole section (typically several tens of feet) that it might often not be a single lithology in the open hole. This can be seen from the gamma ray logs of some of the Amerada Hess wells and is discussed below.

Pumping rates are not included in the Amerada Hess pro-forma although they are sometimes jotted down anyway by the drilling engineer. However, pumping rates during leak-off tests are known to be between 0.25 to 1 barrel per minute which is approximately 0.7 to 1.4 litres per second (see section 4.6). With the large volume of open hole generally being pressurised during a leak-off test, these rates can be considered as being at the low end of the spectrum of rates encountered in hydraulic fracturing.

The approximate location of the Amerada Hess wells, for which leak-off test records have been obtained, is given by their quadrant and block numbers which are listed in Table 5.6. Also listed here are the number of wells in these quadrants, and the number of leak-off tests performed. Good quality leak-off test records are available in most cases, as is most of the further well information described in the next sections (5.4.2 and 5.4.3).
Table 5.6 A summary of Amerada Hess wells and leak-off test records from the North Sea.

Table 5.6 shows that Amerada Hess have been drilling in a wide range of locations within the North Sea between 1989 and 1995. With reference to Figure 5.1, it can be seen that the data covers the West of Shetland Basin and northern North Sea, the central North Sea and the southern North Sea. The majority of the data falls within the central North Sea. In fact within only 2 blocks, 15/21 and 15/22, there are over 100 leak-off tests.

5.4.2 Amerada Hess Final Well Reports.

Final Well Reports (FWRs), or sections of FWRs have been obtained for most of the wells where Amerada Hess drilling records have been studied. The Amerada Hess FWRs summarise most of the drilling, testing and completion data from each well. For this study FWRs have provided the following information:
**Formation pore pressure profile:** Changes in the pore pressure of the formations being drilled through are estimated by monitoring a variety of drilling parameters. These parameters, which are given below, are combined to give an equation (equation 5.1) containing the drilling exponent or d-exponent (Mills, 1984). The theory behind the d-exponent method relies on the fact that, all other things remaining unchanged, the rate of penetration will increase when entering an over pressured formation. The increase in drilling rate is due to the increased porosity of the over pressured formation. The d-exponent is found by solving the following equation that is given by Mills (1984):

\[ R = N \left( \frac{W}{D_b} \right)^d \]  

(5.1)

Where \( R \) is the rate of penetration of the drill bit (ft/hour), \( N \) is the rotary speed (per minute), \( W \) is the weight on the drill bit (lbs), \( D_b \) is the diameter of the drill bit (inches) and \( d \) is the d-exponent. When the d-exponent is plotted against depth, it increases when drilling through normally pressured formations. Increases in formation pore pressure are indicated by a reversal of the trend of the d-exponent with depth. Estimates of over pressure using the d-exponent have considerable limitations, but provide the most practical indicators generally available (Mills, 1984).

Although the d-exponent is not a direct measurement of pore pressure it gives an indication of relative changes in pore pressure and thus overpressured zones can be identified. When an over pressured zone is detected, the mud weight is increased to control the well, that is to stop formation pore fluid entering the wellbore. Estimated pore pressure, mud weight and leak-off test results are generally plotted on a pore pressure profile diagram (Figure 5.5). Other well data such as lithostratigraphy and casing shoe positions are also included on the diagram but are not intended for accurate correlation. The diagrams are simply a schematic summary of well information.

**Lithostratigraphic Summary:** An accurate lithostratigraphic summary is included in most FWRs. Chronostratigraphic information is often also included. This type of data tends to become more detailed towards the bottom of the well, nearer the target. Often the top several thousand feet, in the central and northern North sea is undifferentiated as it is
generally a thick, undeformed blanket of Cenozoic sediment with no hydrocarbon trapping or source potential. Where good lithostratigraphic information is available it can be used to corroborate any other lithology information available for the formations in which the leak-off tests have been conducted.

**Deviation Survey:** Many of the wells drilled by Amerada Hess are not vertical wells. Development wells often deviate several tens of degrees from vertical. A deviation survey keeps an accurate track of the degree of deviation and the azimuth in which it is deviated. Deviation surveys are often included in the FWR. In an anisotropic stress field, the stresses at the wall of a deviated borehole are different to those at the wall of a vertical borehole. The well deviation should therefore be considered in the analysis of leak-off test data.

**Leak-off Test Data:** Leak-off test data in final well reports usually consists of a single pressure value or equivalent mud weight. In approximately half of the reports there is some indication as to whether leak-off was achieved or not during the test, usually a simple yes or no, which is presumably the opinion of the drilling engineer. Some wells have had a FWR compiled by Geoservices, as opposed to, or in addition to the in-house Amerada Hess FWR. Geoservices is a service company which undertakes a variety of activities during the drilling of the well. Some of the Geoservices FWRs contain pressure volume plots from the leak-off tests. Geoservices pressure volume plots are similar to those in the leak-off test reports except that they are computer generated, they list an "intake pressure" which is equivalent to the leak-off pressure, and a maximum pressure which is usually only slightly higher than the leak-off pressure (Figure 5.6). The Geoservices pressure/volume plots essentially duplicate the leak-off test report, however they are useful when the leak-off test report is missing or incomplete. They also serve to corroborate the pressures derived from the leak-off test reports.
Figure 5.5 An example of a pressure profile diagram from an Amerada Hess final well report. The line labeled “fracture gradient” represents the trend of the anticipated leak-off pressure (in equivalent mud weight - EMW) against depth, based on knowledge gained from previously drilled wells in the same area.
Figure 5.6 An example of leak-off test data included in some final well reports by GeoServices. The intake pressure is equivalent to the leak-off pressure.

**BASIC DATA**

- RKB-SL = 86 ft
- RKB-SB = 556 ft
- SHOE at = 5288 ft
- TVD' = 4999 ft
- MUD WT = 10.4 ppg

**MAXIMUM PRESSURE**
- = 1185 psi

**INTAKE PRESSURE**
- = 1130 psi

- PUMPED 4.00 bbls
- RETURN 3.00 bbls
- LOST 1.00 bbls

**FINAL RESULTS:**

- EQUIVALENT MUD WEIGHT GRADIENT = 14.73 ppg
- MAXIMUM SAFE MUD WEIGHT = 13.96 ppg
5.4.3 Amerada Hess Well Logs.

Geophysical logs, or well logs, are run in all oil wells drilled in the North Sea. Which logs are run and over what depth range varies from well to well. The gamma ray, sonic and density logs are run in almost all the Amerada Hess wells for which leak-off test records have been obtained. The gamma ray log is often run throughout almost the entire depth of the well. The depth range of the sonic and density logs is more varied.

The sonic log and density log have already been described in section 4.7 under vertical stress estimation. The gamma ray log is a record of a formations radioactivity. The radiation emanates from naturally occurring uranium, thorium and potassium which occur in the rock forming minerals. Amongst the sedimentary rocks, shale has by far the strongest radiation (Rider, 1986). The simple gamma ray log simply measures the total amount of radioactivity and is thus often used at least qualitatively as a guide to the proportion of shale in the section of hole being logged. However, not all shales are radioactive, and not all radioactive minerals form shale. Despite this, the simple gamma ray log is still used to determine the degree of shaliness. Further, certain lithologies are thought to have typical gamma ray values, for example a gamma ray value of under approximately 35 API (units standardised by the American Petroleum Institute) will often be classed as a sand. Gamma ray values of over 100 API would almost certainly be shales. The majority of the gamma ray readings fall somewhere between the two and probably represent intermediate lithologies.

Before gamma ray logs are useful, they must be prepared for interpretation. Various corrections have to be performed to counter the effects of such factors as borehole washouts and drilling mud chemistry on the readings. Corrected gamma ray logs were obtained from the Hess wells database. In some cases, the gamma ray log has been useful for identifying the lithology in the open hole section in which a leak-off test has been performed. In these cases it is clear that the readings are representing a particular lithology. For example, the Kimmeridge clay formation which is encountered in many of the Amerada Hess wells has a very distinctive gamma ray signature with counts often well over 100 API. Reading this high are clearly representative of shales.
The gamma ray log has generally only been used to corroborate other lithological information such as might be included in the drilling report. It can also be useful for giving an indication of the degree of lithological variation within a section of borehole. Only in a few cases however, such as the case of the Kimmeridge clay, has it been used in this study as the sole lithological indicator.

5.5 Hydraulic Fracturing and Leak-off Test Data

In order to make a comparison of leak-off pressures and stress magnitudes from hydraulic fracturing (hydro-frac) tests, data was sought out where both leak-offs and hydro-fracs had been performed in the same holes. Two sources of such data were discovered, UK Nirex’s data from Sellafield, and data from the southern North Sea where hydro-fracing has been performed by Dowell Schlumberger in wells operated by BP and Hamilton Oil/BHP.

In the case of the hydro-frac data, from both Sellafield and the North Sea, the numerical raw data was not available. However, the data was available in graphical format, where it had been analysed to extract (amongst other things) minimum horizontal stress magnitude. This analysis has been performed by Geoscience, in the case of the Nirex data, and Dowell Schlumberger in the case of the North Sea data. In both cases, detailed reports which present the raw data in pressure/time format and show how the data was analysed, have been obtained. This section describes the hydro-frac and leak-off data obtained from the above two sources.

5.5.1 Nirex Data

The job of UK Nirex was to assess the geology and hydrogeology of the Sellafield area in order to determine if it would make a suitable site for the construction of an underground repository for nuclear waste. During the course of this assessment at least 16 boreholes were drilled to depths of up to 2000m within an area of approximately 7 km by 7 km for the purpose of determining the geology and performing an extensive range of geophysical and hydrogeological tests. Close to half the boreholes are concentrated within an area of about 1 km² which had been selected after initial studies.
as the best potential site for the repository. In fact, the concentration of wireline
geophysical (well logs) and other borehole measurements made at Sellafield, make this
area probably one of the best geophysically characterised areas in the world. This
extensive range of tests included approximately 5 or 6 hydraulic fracturing stress
measurement tests in each of 5 boreholes. During the drilling of at least 9 of these
boreholes, leak-off tests were performed. Figure 5.7 shows the location of the potential
repository zone and the boreholes. Table 5.7 summarises those boreholes from which
hydro-frac and leak-off data has been obtained for this study.

<table>
<thead>
<tr>
<th>Nirex Borehole Number</th>
<th>Number of Hydro-fracs</th>
<th>Number of Double Leak-off Tests</th>
<th>Number of Single Leak-off Tests</th>
</tr>
</thead>
<tbody>
<tr>
<td>2</td>
<td>6</td>
<td>2</td>
<td>0</td>
</tr>
<tr>
<td>3</td>
<td>6</td>
<td>0</td>
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</tr>
<tr>
<td>5</td>
<td>7</td>
<td>1</td>
<td>0</td>
</tr>
<tr>
<td>8a</td>
<td>3</td>
<td>1</td>
<td>0</td>
</tr>
<tr>
<td>10a</td>
<td>6</td>
<td>1</td>
<td>0</td>
</tr>
<tr>
<td>11a</td>
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<td>2</td>
<td>0</td>
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<td>0</td>
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<td>13a</td>
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<td>0</td>
<td>1</td>
</tr>
<tr>
<td>14a</td>
<td>0</td>
<td>1</td>
<td>0</td>
</tr>
</tbody>
</table>

Table 5.7 Nirex hydro-frac and leak-off test data.
Figure 5.7 Plan of the Sellafield site area showing the location of the potential repository zone and the Nirex boreholes.
5.5.1.1 Nirex Leak-off Tests

As the Nirex leak-off tests were not performed for scientific purposes, the leak-off test records do not form part of any official Nirex report. Fortunately however, leak-off tests records had been kept by the geological consultants, Sir Alexander Gibb and Partners, who were involved in the actual drilling operations. The Nirex leak-off tests were performed at casing shoes for the same reasons that they are performed in the oil industry, i.e. to test cement integrity and the strength of the next formation to be drilled. The standard leak-off test procedure and the double leak-off test procedure (used for most of the Nirex leak-off tests) have been described in section 4.6 and the results are presented in Chapter Six. The leak-off test records themselves are described in this section.

A standard leak-off test report pro-forma was used for all Nirex leak-off tests. The format of the leak-off test report is very similar to those used by Amerada Hess. However, unlike the Amerada Hess drilling records, the Nirex leak-off test reports list a series of precise instructions to the drilling engineer as to how to carry out the test. These instructions can be seen in Figure 5.8 which is an example of a Nirex leak-off test record. These instructions, assuming that they have been followed, are useful for the interpretation of the leak-off test record as they remove some of the subjectivity involved on the part of the drilling engineer as to how to conduct the test.

Again, the essential part of the report is the pressure/volume plot, or pressure/time plot with constant pumping rate. In most Nirex records it is actually a pressure/time plot which has been recorded and it is stipulated in the instructions that the tests are conducted at a constant pumping rate of between a quarter and a half barrel per minute (approximately 0.7 to 1.4 litres per second).

The volume pumped, and volume returned after the test, is recorded in most of the leak-off test reports. As described in section 5.4.1 for the Amerada Hess records, this information can be useful for determining whether or not leak-off occurred.
1. After setting casing and waiting on cement, clean out casing and drill down of new hole.
2. Pull back inside casing.
3. With the pipe gate closed, use the cementing unit to pump mud down the drillpipe at a slow, constant rate (40 ft per hour). Record and plot pressure run versus time, and volume injected versus time. The pump rate should be continuous. DO NOT PUMP IN STAGES.
4. Shut well immediately after peak pressure is ascended (see diagram) and continue to monitor pressure versus time until pressure is almost constant. Do not bleed off the pressure.
5. Release pressure and record bleed back volume.
6. Repeat stages 3 to 5 (see diagram).
7. Record mud weights and casing shoe TVD.

Figure 5.8 An example of a leak-off test record from the Nirex boreholes. Precise instructions to the drilling engineer are listed as part of the pro-forma. Stage 6, i.e. the second pressurisation cycle, is recorded on a separate sheet.
The instructions to the drilling engineer state that after the peak pressure is reached, the system should be shut in and the pressure decline monitored until it becomes stable. This seems to imply that an attempt is being made to determine some kind of shut-in pressure, however the limitations of this technique with regards to the Nirex leak-off data have been discussed in section 4.6.6.2.

The whole leak-off test procedure is repeated after the first pressurisation cycle. In most cases the second leak-off test is recorded on a new leak-off test pro-forma. The positive implications for determining stress magnitudes from the second cycle have been discussed in section 4.6.6.2.

A point of some interest about the data from Nirex is that although all the scientific data is reported using SI units, the leak-off test data is all in imperial units (e.g. psi, ppg, and ft) which indicates that the drilling and leak-off test procedures are firmly rooted in the oil industry.

5.5.1.2 Nirex-Hydraulic Fracturing Stress Measurements.

Hydraulic fracturing tests were performed in the Nirex wells at Sellafield specifically for the purpose of measuring crustal stress magnitudes. Crustal stress magnitudes are considered crucial to the design of the proposed underground waste repository and for hydrogeological models of fluid flow through fractured rock networks.

Test intervals were selected on the basis of the intact nature of the rock; the ideal interval being without any major fractures or discontinuities. The intervals were selected by examination of the core logs and the borehole imaging logs (e.g. Formation MicroImager, Acoustic Telescaner).
The tests were carried out in accordance with the ISRM Suggested methods for stress determination (Kim and Franklin, 1987). The test zones were isolated using a straddle packer assembly (section 4.4) with a spacing of 4m between the upper and lower inflatable packers. The system, which included a down hole pressure gauge, was lowered into the hole on the end of a 5 inch drill pipe. The surface system included a positive displacement high pressure-low flowrate pump, two pressure transducers and two flowmeters.

Each hydro-frac test comprised three injection stages; (i) breakdown (the initial pressurisation cycle in which the fracture is initiated and propagated) (ii) fast re-frac (the system is re-pressurised rapidly to re-open the fracture) (iii) slow re-frac (the fracture is re-opened again, this time using a slow pumping rate which increases in a step like fashion). Figure 5.9 shows an example of a complete pressure/time record from borehole 5 hydro-frac 1.

The tensile strength of the formations tested has been determined using two methods. Firstly, tensile strength has been inferred from the difference between the breakdown pressure and the fast re-frac pressure (section 4.5). The value of tensile strength determined from this technique is sometimes referred to as the hydro-frac tensile strength. Tensile strength has also been measured in the laboratory on core samples taken from the intervals where hydro-fracs have been performed. These tensile strength values can be called the laboratory tensile strengths. The values of tensile strength obtained from the two methods are quite different and are discussed further in section 6.2.3.2.

The step rate re-opening test was used during the slow re-frac for all the hydro-frac tests in order to estimate the minimum horizontal stress magnitude ($\sigma_h$) from the re-opening pressure. Generally the step rate re-opening tests were considered adequate to estimate $\sigma_h$. However, in some cases, problems with the flow rate controlling choke, during the step rate re-opening tests, led to uncertainty in the values of $\sigma_h$ determined by these tests. In these cases, in addition to the step rate re-opening test, pressure decline data after shut-in, following either the breakdown or any of the re-frac cycles, has also been
used to determine $\sigma_h$. To check measurement consistency, both corrected surface pressures and downhole pressures were used. The theory behind these methods has been discussed in section 4.5.1. In most cases, where both methods of $\sigma_h$ determination were used, the re-opening pressures were close to the closure pressures. A comparison of results from the two methods, in the cases where both methods have been used, is shown in Table 5.8.
Figure 5.9 An example of a complete pressure/time record from a hydro-frac test performed in a Nirex borehole. The flow rate is also recorded in this diagram. It can be seen from the five main peaks in this plot that five pressurisation cycles were performed as part of this hydro-frac. Firstly there is breakdown pressurisation cycle, this is followed by two fast re-fracs and then two slow re-fracs.
<table>
<thead>
<tr>
<th>Nirex Borehole/Hydrofrac number.</th>
<th>σ_h determined from pressure decline analysis (p_c) /MPa</th>
<th>σ_h determined from step rate re-opening analysis (p_c) /MPa</th>
<th>Difference between p_c and p_c as % of p_c.</th>
</tr>
</thead>
<tbody>
<tr>
<td>BH3/HF1</td>
<td>33.0</td>
<td>33.6</td>
<td>+1.8%</td>
</tr>
<tr>
<td>BH3/HF2</td>
<td>31.2</td>
<td>30.6*</td>
<td>-1.9%</td>
</tr>
<tr>
<td>BH3/HF3</td>
<td>31.4</td>
<td>37.8*</td>
<td>+20%</td>
</tr>
<tr>
<td>BH3/HF4</td>
<td>27.0</td>
<td>32.8</td>
<td>+21%</td>
</tr>
<tr>
<td>BH3/HF5</td>
<td>33.7</td>
<td>35.6*</td>
<td>+5.6%</td>
</tr>
<tr>
<td>BH3/HF6</td>
<td>32.4</td>
<td>36.8*</td>
<td>+13.5%</td>
</tr>
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<td>BH5/HF1</td>
<td>23.1</td>
<td>23.8</td>
<td>+3%</td>
</tr>
<tr>
<td>BH5/HF2</td>
<td>26.5</td>
<td>26.9</td>
<td>+1.5%</td>
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<td>27.3</td>
<td>+13.3%</td>
</tr>
<tr>
<td>BH5/HF4</td>
<td>20.0</td>
<td>22.2</td>
<td>+11%</td>
</tr>
<tr>
<td>BH5/HF5</td>
<td>23.8</td>
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<td>BH5/HF6</td>
<td>17.2</td>
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<tr>
<td>BH5/HF7</td>
<td>13.7</td>
<td>13.0</td>
<td>-5.1%</td>
</tr>
</tbody>
</table>

Table 5.8 Comparison of σ_h determined from pressure decline analysis and step rate re-opening analysis for Nirex hydro-frac stress measurements. * considered unreliable (see below).

In some cases, lack of equipment availability or poor equipment combinations meant that unreliable results were obtained from the step rate tests. Where it is indicated in the Nirex factual reports that the results are not considered reliable, they have been marked with an asterisk in Table 5.8. For example, in BH3/HF3 and HF5, the flow meters which were used did not function properly and were not able to give accurate readings at low flow rates, resulting in the step rate re-opening data being incomplete. In such cases the value of σ_h determined from pressure decline analysis is considered more reliable.
Figure 5.10 An example of flow rate against plateau pressure achieved at that flow rate during a step rate re-opening test. This data comes from the test shown in Figure 5.9. The step rate re-opening test has been performed twice, i.e. two slow re-fracs have been performed. The results from both tests are in good agreement and indicate fracture extension at 23.8 MPa, the point of maximum curvature.
Figure 5.10 shows the results of a step rate test from borehole number 5 where the plot is equivalent to the schematic representation in Chapter Four (Figure 4.8b). The value of $\sigma_h$ picked from this plot (23.8 MPa) is seen to be the point of maximum curvature. This is the fracture extension pressure, and as explained in section 4.5.1, can be slightly higher than $\sigma_h$, partly because of frictional effects. With the low flow rates used in this test, frictional effects are likely to be small.

Figure 5.11 shows the results of a pressure decline analysis that has been performed to determine the fracture closure pressure from the same borehole interval. A plot is made of the rate of pressure decline in the fractured interval against the square root of time since shut-in. The point at which the gradient of this curve becomes zero is taken as the point of fracture closure (section 4.5.1). This type of pressure decline analysis has also been used to determine the fracture closure pressures in the other intervals listed in Table 5.8.

The formation imaging tools were run through the fractured sections a second time, after the hydro-frac stress tests. The hydraulically induced fractures can clearly be seen in most cases. These before and after observations can be used to make sure that a fracture was induced and that the results did not reflect the opening of a pre-existing fracture. The orientation (both azimuth and inclination) of the induced fracture can be determined from the formation imaging logs. In most cases the induced fractures were vertical indicating that the minimum stress is horizontal. The azimuth of the fracture trace around the borehole was generally seen to be parallel to the maximum horizontal stress orientation which has been determined from borehole breakout analysis in many of the Nirex boreholes.

5.5.2 Southern North Sea Data

This data set from the southern North Sea consists of leak-off test data and estimates of the minimum horizontal stress magnitude ($\sigma_h$) made from hydraulic fracturing tests performed in the same wells. The wells in question are operated by BP and Hamilton Brothers Oil (now BHP). All bar one of the North Sea hydro-frac tests that were obtained came from blocks 42/30 and 43/26a. These blocks contain the Ravenspurn
fields (Ravenspurn North and Ravenspurn South) which lie in the Sole Pit Basin (Figure 5.1). The hydro-fracs are all performed in the Permian, Rotliegend Sandstone gas reservoir. It seems that in the North Sea, hydro-fracing is not a particularly common practice, especially outside of the southern North Sea. Obtaining hydro-frac data from the North Sea for this study has not been a particularly easy process. After approaching various oil companies with little success, the hydro-frac service company (Dowell Schlumberger) was approached. Dowell Schlumberger directed the search towards researchers in Delft Technical University who had already obtained the relevant information (copies of the hydro-frac reports) for another purpose. The North Sea hydro-frac data presented in this study is a combination of data from Dowell Schlumberger hydro-frac reports and data analysed by workers from TU Delft.

Table 5.9 summarises the BP and BHP wells in blocks 42/30 and 43/26a in which either hydro-frac or leak-off data has been obtained. Most of the wells have both hydro-frac and leak-off data.
<table>
<thead>
<tr>
<th>Well Name</th>
<th>Company</th>
<th>Number of leak-off tests</th>
<th>Number of hydro-fracs</th>
</tr>
</thead>
<tbody>
<tr>
<td>42/30-RA01</td>
<td>BP</td>
<td>3</td>
<td>1</td>
</tr>
<tr>
<td>42/30-RA02</td>
<td>BP</td>
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<td>2</td>
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<td>42/30-RA05</td>
<td>BP</td>
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<td>1</td>
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<td>BP</td>
<td>3</td>
<td>1</td>
</tr>
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</tr>
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<tr>
<td>42/26a-E8</td>
<td>BHP</td>
<td>2</td>
<td>1</td>
</tr>
</tbody>
</table>

Table 5.9 Southern North Sea hydro-frac and leak-off data.
Figure 5.11 An example of the pressure decline rate against time method of determining fracture closure pressure used in Nirex hydro-frac tests. This pressure decline data comes from the test shown in Figure 5.9. The closure pressure is picked as the point at which the pressure decline rate becomes constant.
5.5.2.1 Southern North Sea Leak-off Test Data
The leak-off test data was sought out once the hydro-frac data had been obtained. BP and BHP were approached and asked if they still kept leak-off test records for the wells in question. BP had some drilling records archived, some of which contained some basic leak-off test information. Only in a couple of cases were leak-off pressure/volume records present. Most of the leak-off test data has had to come from single leak-off pressure values either from drilling reports or final well reports. In most cases these reports noted whether the test was taken to leak-off or was a limit test. The formation in which the leak-off test was performed was generally also available or could be inferred from the final well report stratigraphy.

BHP provided a list of leak-off and limit test pressures with depths and borehole names. Lithologies in which these leak-off tests were performed were obtained for some of the tests by cross reference to the lithologies contained within hydro-frac reports prepared by TU Delft.

Other leak-off test data also from blocks 42/30 and 43/26a were contained in Erico Pressure Study 5. Although not in the same holes as the hydro-fracs, these leak-off tests, being in close proximity, in wells targeted at the same gas fields and the same structural setting, can still be useful for this study, provided they are used with caution.

5.5.2.2 Southern North Sea Hydro-frac Data
All the hydro-frac data were obtained via TU Delft. In most cases the hydro-frac report (frac-report), prepared by Dowell Schlumberger (DS) who performed the fracturing job (frac job), was available. In the case of wells from 43/26a, data from DS frac reports had been summarised by TU Delft researchers in data base format, but the original reports were not available.

DS perform hydro-fracs in the North Sea for the purpose of stimulating low permeability reservoirs. The “frac-job” consists of two parts. The first part is the so called mini-frac (or data frac) which is a scaled down version of the main frac and is performed for the
purposes of determining certain reservoir parameters which are needed for the main frac's design. These parameters include the minimum horizontal stress magnitude ($\sigma_h$).

The second part is the main frac itself in which the fracture is propagated into the reservoir and then held open by being pumped with proppant.

The DS frac reports detail both the mini frac and the main frac. Although the emphasis of these reports is on the logistical and economic aspects of the frac job (i.e. length of time taken, what equipment was used and how many barrels of fracturing fluid and proppant were pumped etc.) technical details, such as the pressure/time records (Figure 5.12) are also included.

The fractures discussed in this section are performed in completed wells which have been fully cased. To perform a fracture in a fully cased well, there is generally no need to set packers around the interval to be fractured. Instead, the casing is perforated at the desired depth. The rest of the formation is protected by the casing, so it is only interval where the perforations exist which is fractured. The perforations in these wells are multiple perforations which have been made over a depth interval comparable to (but less than) the thickness of the reservoir formation. In some cases a packer is set as well to isolate the section if there is concern about the strength of cement jobs in the higher parts of the hole.

The value of $\sigma_h$ is obtained from analysis of the mini frac data. An example of a complete pressure/time record from a DS mini frac is given in Figure 5.12. A typical mini frac procedure is as follows:

1. The formation is broken down by pumping fracturing fluid at rates of up to 50 barrels per minute (BPM). The fracture is propagated by continuing to pump fracturing fluid for several minutes. During the mini fracs performed in the Ravenspurn fields, several hundred barrels of fracturing fluid were typically pumped.

2. The system is shut in and the pressure decline is monitored.
3. The system is repressurised by pumping a similar amount of fracturing fluid as in the first cycle.

4. Again the system is shut in and the pressure decline is monitored.

5. A step rate test is performed. The pumping rate is increased in increments of around 0.1 BPM from 0.1 BPM to approximately 2.0 BPM. Each increment is pumped until a pressure plateau is reached. The pressure at this plateau is recorded against the relevant pumping rate. A plot of pumping rate against pressure attained is made to determine the fracture re-opening pressure (section 4.5.1). Figure 5.13 shows an example of a plot made in this type of step rate (re-opening) test with the fracture re-opening pressure indicated. The fracture re-opening pressure is thought to be close to (but an upper limit of) $\sigma_b$.

6. The pressure decline after shut in is analysed using the shut in decline test to determine fracture closure pressure. The theory of the shut in decline test has been described in section 4.5.1. As the pressure declines, the data points fall on a straight line until the point of fracture closure, thus from the deviation of the data from linearity, the fracture closure pressure is picked. Figure 5.14 shows an example of pressure decline data and interpreted closure pressure from a DS Ravenspurn mini-frac.

It is thought that fracture re-opening pressures from the step rate test are generally more reliable, as mistakes are possible in picking the right point on the pressure decline plot (Economides and Nolte, 1989). However fracture re-opening pressures do tend to give an upper bound to $\sigma_b$. If both fracture closure pressure and fracture re-opening pressure are successfully determined, a good estimate of $\sigma_b$ can be made. Table 5.10 shows the values of DS determined closure pressures and re-opening pressures where both are available in wells from the southern North Sea.
Figure 5.12 An example (slightly modified for clarity) of the pressure time record made during a mini-frac (or data-frac) performed by Dowell Schlumberger in the Ravenspurn Field, southern North Sea. The plot shows the pressure measured at the wellhead and at the annulus (psi), the flow rate (barrels/min) and the density of the fracturing fluid (g/cm³). The time is in hours and minutes.
Figure 5.13 An example of flow rate against plateau pressure attained at that flow rate from a step rate re-opening test performed by Dowell Schlumberger. The fracture extension pressure has been picked at the point of maximum curvature as 6580 psi.
Figure 5.14 An example of pressure against square root of time data from a pressure decline analysis performed during a hydro-frac by Dowell Schlumberger in the southern North Sea. The fracture closure pressure has been picked as the point at which the pressure against square root of time plot deviates from linearity.
<table>
<thead>
<tr>
<th>Company and Well Name</th>
<th>Depth of Fracture/ft</th>
<th>Frac Closure Pressure ($p_c$/psi)</th>
<th>Frac Re-opening Pressure ($p_r$/psi)</th>
<th>Difference between $p_r$ and $p_c$ as % of $p_c$.</th>
</tr>
</thead>
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<td>+3.4%</td>
</tr>
<tr>
<td>BP 42/30-RB02</td>
<td>9630</td>
<td>7057</td>
<td>7200</td>
<td>+2.0%</td>
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<tr>
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<td>6580</td>
<td>7300</td>
<td>+10.9%</td>
</tr>
<tr>
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<td>6100</td>
<td>+2.5%</td>
</tr>
<tr>
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<td>5829</td>
<td>6606</td>
<td>+11.8%</td>
</tr>
<tr>
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<td>5700</td>
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<tr>
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</tr>
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</tr>
<tr>
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<td>6996</td>
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<tr>
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<td>6162</td>
<td>6625</td>
<td>+7.5%</td>
</tr>
</tbody>
</table>

Table 5.10 Fracture closure and opening pressures from the southern North Sea.

The orientation of the fractures in these hydro-fracs is not determined directly. It is assumed to propagate perpendicular to the direction of $\sigma_h$. These fractures, despite the name mini frac, are large compared to the type of hydro-frac that is performed specifically to measure $\sigma_h$ such as the Nirex hydro-fracs. Even if the fracture initiates at the well bore wall by opening a pre-existing weakness, it is likely, due to its size, to rotate into the direction perpendicular to $\sigma_h$ as it propagates into the formation. It is a reasonable assumption then that the fracture pressures do indeed sample $\sigma_h$. 

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Chapter Six - Analysis and Interpretation of Leak-off Test and Hydrofrac Data.

6.1 Introduction

The leak-off test and hydro-frac test data is analysed in this chapter. The datasets have been analysed by plotting the appropriate pressures derived from these tests (e.g. leak-off pressures, hydro-frac closure and re-opening pressures etc.) against depth. The leak-off pressure data has also been analysed by plotting functions of the leak-off pressure (e.g. the leak-off pressure gradient) against depth. The results of these analyses are presented below.

For the purpose of comparing LOPs with values of minimum horizontal stress ($\sigma_h$), section 6.2 analyses the leak-off test data and the hydro-frac derived estimates of $\sigma_h$, where both tests have been performed in the same boreholes. Also, the trends of LOPs and LTPs with depth, within a particular geographical/structural setting, are compared to the trends of hydro-frac determined $\sigma_h$ with depth. With the aid of these comparisons the relationship between the leak-off pressure and $\sigma_h$ is discussed within the theoretical framework of stress, rock properties, crack behavior and borehole fluids etc. outlined in Chapter 4. It is concluded that, where leak-off test pressure/volume plots are available, a good estimate of $\sigma_h$ can be made in rocks which contain pre-existing permeable cracks.

Section 6.3 presents the trends of the Erico leak-off test data with depth in the North Sea, and analyses how these trends vary according to parameters such as geographical/structural domain, lithology in which the tests are performed, and formation pore pressure in which the tests are performed. Although the evidence presented in section 6.2 shows that leak-off pressures generally correspond closely with $\sigma_h$, for the purposes of this section, the data is presented simply in terms of LOPs and LTPs.

The more detailed Amerada Hess dataset is analysed in section 6.4. The data is analysed in the same way as the data in section 6.3. In addition, the pressure/volume plots from the leak-off test records have been examined and on this basis, estimates of $\sigma_h$ have been
made. Density logs have also been obtained for these wells and thus vertical stress has also been calculated.

Throughout this chapter, the pressures discussed (leak-off pressure, limit test pressure, hydro-frac derived pressures etc.) are all “down hole” pressures. That is, they are always the sum of the surface pump (or pressure gauge) pressure, plus the hydrostatic pressure of the fluid in the wellbore down to the vertical depth of the test. The corresponding depths used in this chapter are always vertical depths measured from ground level (TVDGL) in the case of onshore boreholes, and sea level (TVDSS) in the case of offshore boreholes.

In several of the plots presented in this chapter, a line of best fit has been added to highlight the trend of the dataset. The lines of best fit have been added by the plotting program, Microsoft Excel 5.0. They are linear regression lines fitted to the data using the least squares method.

6.2 Comparison of Leak-off Pressures with Hydro-frac Determined $\sigma_h$
This section presents the results of a comparison of the leak-off pressures (LOPs) and estimates of the magnitude of the minimum horizontal stress ($\sigma_h$) based on hydraulic fracture closure ($p_c$) and re-opening pressures ($p_r$), where leak-off tests and hydro-fracs have been performed in the same boreholes.

The methodologies and procedures used within the oil industry and by UK Nirex to obtain this data have been described in Chapter 4. The data presented in this section, and the specific analyses used to derive stress estimates from the hydro-frac test data, have been described in Chapter 5.

In the final part of this section, some conclusions are drawn regarding the use of LOPs as estimates of $\sigma_h$.

6.2.1 Nirex Hydro-frac and Leak-off Data
The data presented in this section has been summarised in Table 5.7. In the first part of this section the data is firstly presented, borehole by borehole, where both hydro-frac tests and leak-offs have been performed (Boreholes 2, 3, 5, 8A and 10A). The data is then all presented together to display the trends of the leak-off pressures and hydro-frac closure and re-opening pressures with depth. The general trends of these data sets are discussed. Exceptions to the leak-off pressure trend are discussed in detail with reference to the leak-off test records.

The second part of this section is focused on the 2nd LOP. The leak-off test records are examined for the purpose of determining the inferred crack opening and closing behavior. The trend with depth of the fracture opening and closing pressures estimated from the leak-off test records is compared with the trend of the hydro-frac determined opening and closing pressures.

6.2.1.1 Comparison of Nirex Leak-off Test Pressures and Hydro-frac Determined $\sigma_h$

Figures 6.1 to 6.5 show the values of hydro-frac re-opening and closure pressures as determined in the Nirex reports (Nirex, 1991, 1991a, 1992, 1994, 1994a) plotted with LOP against vertical depth from the ground surface (depth TVDGL), for boreholes where both hydro-frac tests and leak-offs have been performed (Boreholes 2, 3, 5, 8A and 10A). In boreholes 2, 8A and 10A, only the step rate re-opening pressure (see section 5.5) has been determined and for these tests this is considered to be a good estimate of $\sigma_h$ (section 5.5.1.2). In the boreholes where both hydro-frac closure pressures and hydro-frac re-opening pressures have been determined reliably (boreholes 3 and 5), these pressures are both displayed. Four of the re-opening pressures in borehole 3 were considered unreliable (section 5.5.1.2) and have not been displayed. In these cases the closure pressure must be used as the best estimate of $\sigma_h$. Where reliable re-opening pressures have been obtained in boreholes 3 and 5, it is these pressures which are thought to provide the best estimate of $\sigma_h$ (Nirex 1996).

In the cases where double leak-off tests have been performed, LOPs from both 1st and 2nd pressurisation cycles are plotted.
Figure 6.1 Sellafield borehole 2. Hydro-frac re-opening pressures from step rate re-opening tests plotted with leak-off pressures, against depth.
Figure 6.2 Sellafield borehole 3. Hydro-frac re-opening pressures from step rate re-opening tests and hydro-frac closure pressures from pressure decline analysis, plotted with leak-off pressures, against depth.
Figure 6.3 Sellafield borehole 5. Hydro-frac re-opening pressures from step rate re-opening tests and hydro-frac closure pressures from pressure decline analysis, plotted with leak-off pressures, against depth.
Figure 6.4 Sellafield borehole 8A. Hydro-frac re-opening pressures from step rate re-opening tests plotted with leak-off pressures, against depth.
Figure 6.5 Sellafield borehole 10A. Hydro-frac re-opening pressures from step rate re-opening tests plotted with leak-off pressures, against depth.
It can be seen in Figures 6.1 to 6.5 that most of the leak-off pressures plot fairly close to the trend of hydro-frac determined re-opening and closure pressures. The most notable exception to this occurs in borehole 2 at 487m depth, where both 1st and 2nd LOPs plot considerably above the trend of the hydro-frac data. This test is discussed below.

Figure 6.6 summarises the geology of the Nirex boreholes. These borehole sections can be used in conjunction with the borehole location map in Figure 5.7 to define the geological setting of this region. It can be seen that the Triassic Permian and Carboniferous sediments increases in thickness from east to west, and that the whole area is underlain by the Ordovician Borrowdale Volcanics. As the boreholes all sample essentially the same geological setting, it seems reasonable to group the data from these boreholes to examine the regional trends with depth of LOPs and hydro-frac determined re-opening and closure pressures.

Figure 6.7 presents the Nirex interpreted $\sigma_h$ values from all the hydro-frac tests and LOP data from boreholes 2, 3, 5, 8A and 10A plotted against depth (TVDGL). The Nirex interpreted $\sigma_h$ values are the re-opening pressures in all but 4 cases. These 4 cases are from borehole 3, where closure pressures are considered more reliable (section 5.5.1.2). There is broadly a correspondence between the two sets of data although clearly the LOPs, taken as a whole, are higher than the hydro-frac derived $\sigma_h$ values. In some cases the 1st LOP (and the 2nd if breakdown did not occur during the 1st cycle) is considerably higher than the trend of hydro-frac determined $\sigma_h$. This occurs mainly where tests have been conducted in the Borrowdale Volcanic rocks (Table 6.1). The influence on the LOP of the tensile strength and the absence/presence of pre-existing permeable cracks in these rocks is discussed in section 6.2.3.2. The possibility that pre-existing cracks which are not oriented perpendicular to $\sigma_h$, are opened during a leak-off test, might also explain why some of the LOPs are higher than the hydro-frac determined $\sigma_h$ trend. These ideas are discussed further in section 6.2.3.2.
Figure 6.6 Summary of the Geology sampled by the Nirex boreholes, shown against depth in meters from Ordnance Datum (maOD). (Adapted from Nirex, 1993).
Figure 6.7 Sellafield boreholes where hydro-frac and leak-off test data is available: 2, 3, 5, 8A, 10A. Nirex interpreted $\sigma_h$ (determined from hydro-frac data) plotted with leak-off pressures, against depth.
Figure 6.8 Sellafield boreholes where hydro-frac and leak-off test data is available (2, 3, 5, 8A, 10A) plus leak-off pressures from other nearby Sellafield boreholes where hydrofracs have not been performed (11A, 12A, 13A, 14A). Nirex interpreted $\sigma_0$ (determined from hydro-frac data) plotted with leak-off pressures, against depth.
In addition to the leak-off tests from the wells where hydro-fracs were conducted, leak-off test records from a further 4 Nirex boreholes, listed in Table 6.1 and shown on the location map (Figure 5.7), have been obtained. These wells, 11A, 12A, 13A and 14A lie within 2 or 3 km of the hydro-frac tested wells (Figure 5.7), and are also within essentially the same geological setting (Figure 6.6). Table 6.1 summarises all the LOPs obtained from the Nirex boreholes and shows the lithology in which the tests were performed.

Figure 6.8 shows all the Nirex LOPs from both first and second pressurisation cycles together with the Nirex interpreted values of $\sigma_h$, plotted against vertical depth (TVDGL). It can be seen from Figure 6.8 that the general correspondence between the trend of LOP and hydro-frac determined $\sigma_h$, which was seen in Figure 6.7, is still apparent. The only exception to this trend comes from the test conducted in borehole 13a. However, on examination of the pressure volume record for this test, there is no evidence that leak-off occurred, and so this test would appear to be a limit test.

It can be seen from Figure 6.1 to 6.5 and Figures 6.7 and 6.8, that the 2nd LOP is always lower than the 1st leak-off pressure. This is discussed in more detail in the following sections.

As mentioned above, with reference to Figure 6.1, the test conducted in borehole 2 at 487m depth stands out as being an exception to the trend of the leak-off test data. The first LOP plots considerably above the trend of the hydro-frac determined $\sigma_h$ as do the 1st LOPs from several other tests. However, the 2nd LOP is also very high, which is in contrast to all the other tests. This leak-off test was conducted in the Borrowdale Volcanic Group. Examination of the pressure/time plots for this leak-off test indicates that no leak-off occurred in the 1st pressurisation cycle (Figure 6.9a). Although the pressure/time curve in Figure 6.9a shows a small degree of deviation from linearity towards the maximum pressure, the fact that there is no rapid pressure drop after peak pressure and that the pressure decreases very slowly after the pump is stopped, indicates that no fracturing occurred. This is in sharp contrast to the 2nd pressurisation cycle (Figure 6.9b) where the pressure drops very rapidly after peak pressure, indicating
formation breakdown, at the point that the maximum pressure is reached. In the 2nd pressurisation cycle, the formation breaks down at a slightly lower pressure than was attained in the 1st cycle. However, no leak-off occurs before this breakdown, so the 2nd LOP is the breakdown pressure. The rapid and large drop in pressure at breakdown, is indicative of an intact rock with a significant tensile strength such that the breakdown pressure is influenced by the tensile strength and the hoop stress (see background in section 4.5.2 and discussion in section 6.2.3.2).

The fact that the formation breaks down in the 2nd cycle, at a pressure lower than that reached with no breakdown in the 1st cycle, indicates that the formation was weakened during the first cycle. The pressure/time plot from the 1st cycle (Figure 6.9a) does show a small degree of deviation from linearity close to the maximum pressure, which may be indicative of cracks beginning to open. As pumping was stopped very soon after this small deviation from linearity, the cracks may not have had a chance to propagate rapidly into the formation. This initiation of crack opening however, may have weakened the formation, so that during the 2nd cycle the formation breaks down at a slightly lower pressure. The other possibility, is that the difference in pressure of 190 psi between two
cycles, represents the limit of accuracy of the pressure gauges used during the leak-off test.

Figure 6.9b Sellafield borehole 2. Leak-off test at 490m. 2nd pressurisation cycle. Surface pressure/time plot from Nirex records indicating formation breakdown (leak-off pressure) at approximately 1800 psi surface pressure.

6.2.1.2 Using Double Leak-off test Records to Estimate $\sigma_b$

It can be seen from Figures 6.1 to 6.5 that the leak-off pressure in the 1st pressurisation cycle, in all the boreholes where double leak-off tests have been performed, is higher than the leak-off pressure in the 2nd pressurisation cycle. Generally during the 1st cycle, fractures will be initiated from the borehole wall, or previously existing fracture will be extended into the formation. On closing, the fractures are likely to remain hydraulically open, or at least be more permeable than before the 1st cycle (section 4.5.1). The pressurisation rate in leak-off tests is low and therefore the 2nd leak-off cycle is likely be equivalent to a slow re-frac. The fractures opened and extended during the 1st cycle are therefore likely to be pressurised along their length so that the 2nd LOP will generally be a fracture re-opening pressure. The exception to this is the double leak-off in borehole 2, where as discussed above, the formation does not appear to have been fractured during the 1st pressurisation cycle.
Figure 6.10a Sellafield borehole 10A. Leak-off test at 1028m. 1st pressurisation cycle. Surface pressure/time plot from Nirex records indicating formation breakdown (1st leak-off pressure) at 2250 psi followed by a pressure drop and then a stable pressure of 1550 psi.

Figure 6.10b Sellafield borehole 10A. Leak-off test at 1028m. 2nd pressurisation cycle. Surface pressure/time plot indicating fracture re-opening (2nd leak-off pressure) at approximately 1500 psi surface pressure.
Fracture re-opening seems to be clearly illustrated by the pressure/time plots from borehole 10A. Figure 6.10a shows the 1st pressurisation cycle from the leak-off test in borehole 10A. This test is performed in Borrowdale Volcanics. The pressure/time plot shows a peak surface pressure of 2250 psi. Shortly after the peak pressure is reached, the pump is stopped. The record of pressure decline is only made once a minute, so that an analysis of the pressure decline data to determine a closure pressure is not possible. However, it can be seen that the pressure after approximately 2 minutes has stabilised at 1550 psi indicating that the $p_{salp}$ is somewhere between the peak pressure of 2250 psi and 1550 psi. The second pressurisation cycle is more conclusive (Figure 6.10b). It shows a LOP at 1400 psi, which probably represents the pressure at which the fracture begins to re-open. The maximum pressure of 1750 psi could be the fracture extension pressure (typically slightly higher than fracture re-opening pressure - see section 4.5). After pumping is stopped, the pressure declines to 1500 psi which is slightly higher than the 2nd LOP. This difference in pressure may represent the limits of accuracy of the pressure gauges used during the leak-off test. Another explanation is that in both 1st and 2nd pressurisation cycles, the permeability of the Borrowdale volcanics is so low, that the fracture remains propped open at a pressure of around 1500 psi. The pressure required to prop open a fracture would be very close to the stress acting across it.

It can be seen (Figure 6.5) that the 2nd LOP plots very close to the trend of hydro-frac determined re-opening pressures, and so it seems that the data from borehole 10A provides good evidence that the double leak-off tests can be used to give a good estimate of re-opening pressures and thus $\sigma_h$. Where double leak-offs have been performed, and fracturing has occurred during the 1st cycle, a re-opening pressure (the 2nd LOP) can be picked and taken as an estimate of $\sigma_h$. This analysis has been performed on all the leak-off test records except the three single tests and the test in borehole 2 at 487m depth, which is discussed in the previous section. These re-opening pressures (2nd LOPs) are plotted together with the Nirex interpreted $\sigma_h$ values in Figure 6.11.

In the case of the leak-off test at 487m in borehole 2, it can be seen in Figure 6.9b that directly after the peak pressure is reached and the pump is stopped, the surface pressure
drops by 400 psi to 1400 psi. The pressure then slowly drops off by a further 100 psi over the next 2 minutes. The record of pressure decline after shut-in is not recorded accurately enough to be used to determine the fracture closure pressure. However, it can be assumed that the pressure to which the fracturing fluid drops immediately after shut-in (1400 psi in this case) is close to the \( p_{aip} \), which in a small fracture will be very close to the closure pressure. This value can therefore be used as an upper limit to the fracture closure pressure, and thus provide an estimate of \( \sigma_h \). This type of fracture behavior is also seen in two other cases, and estimates of \( p_{aip} \) from these tests are listed in Table 6.1. The estimated values of \( p_{aip} \) from the Nirex leak-off test records have also been compared with the values of \( \sigma_h \) determined from the Nirex hydro-frac data in Figure 6.11.

The single leak-off test in borehole 13A is a limit test and so has not been used here for estimating \( \sigma_h \). The single leak-off test at 1286m in borehole 3 has not been recorded beyond leak-off, and is therefore not used here to estimate \( \sigma_h \).

It can be seen from Figure 6.11 that there is a close correspondence between the hydro-frac determined values of \( \sigma_h \) and the values of \( p_r \) and \( p_{aip} \) determined from the leak-off test records. As there are only three estimates of \( p_{aip} \) made from the leak-off test records, it is not concluded here that this method is necessarily useful. However, in eight cases where a double leak-off has been performed, and the formation fractured in the 1st cycle, the 2nd LOP, interpreted here as the re-opening pressure, is shown to be a good estimate of \( \sigma_h \). Linear best fit lines have been added to the 2nd LOP and the Nirex interpreted \( \sigma_h \) datasets. The relationship between the 2nd LOPs and the hydro-frac determined \( \sigma_h \) values is examined further in section 6.2.3.
Figure 6.11 All Sellafield boreholes where data is available. Nirex interpreted $\sigma_b$ (determined from hydro-frac data) plotted with 2nd leak-off pressures (from tests in which leak-off occurred in the 1st pressurisation cycle) and $p_{up}$ estimated from leak-off test records, against depth.
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<th>LOP (MPa)</th>
<th>Lithology</th>
<th>( p_r ) (MPa)</th>
<th>( p_{isip} ) (MPa)</th>
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<td>14.9</td>
<td>St. Bees Sandstone</td>
<td>14.9</td>
<td></td>
</tr>
<tr>
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<td>1st</td>
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<td>10.1</td>
<td>St. Bees Sandstone</td>
<td>9.4</td>
<td></td>
</tr>
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<td>9.4</td>
<td>St. Bees Sandstone</td>
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</tbody>
</table>

Table 6.1 Nirex leak-off test pressures and estimates of \( p_r \) and \( p_{isip} \) made from the leak-off test pressure/time plots.

6.2.2 Southern North Sea Leak-off Test and Hydro-frac Data

The data presented in this section has been described in section 5.5.2 and summarised in Table 5.9. The data all come from boreholes drilled in the Ravenspurn fields in the southern North Sea. The size and location of these fields, and their structural setting
within the Southern Permian Gas Basin is shown in Figure 6.12. The Ravenspurn North and South fields lie adjacent to each other and have almost identical lithostratigraphies. The Leman sandstone forms the reservoir rock in both fields with depths to the top of the reservoir of between 9,000 and 10,000 ft varying little between the two fields, and throughout each field. A generalised lithostratigraphy of the Ravenspurn South field is shown in Figure 6.13.
The data from this area differs from the Nirex data in that all the hydro-frac data obtained from the southern North Sea comes from hydro-fracs performed in the Rotliegend Sandstone gas reservoir, and is thus confined to a small depth range (Figure 6.14). This data does therefore not lend itself to making comparisons between the trends with depth of the leak-off and hydro-frac pressures. Due to the larger number of boreholes in this data set, and because the hydro-frac data all lie in a narrow depth range, borehole by borehole plots of the data have not been included here. Instead, the
data are simply all plotted together in Figure 6.14. Units of feet (ft) and pounds per square inch (psi) are the most commonly used units of depth and pressure in the oil industry and as all the data for this part of the study is oil industry data, these units are used here. Using these units facilitates quick and easy comparisons with other oil industry data. The depths used here are true vertical depths measured from sea surface (TVDSS). (1m approximately equal 3.28 ft, 1 MPa approximately equals 145 psi).

Figure 6.14 shows the fracture closure pressures and fracture re-opening pressures obtained from pressure decline analysis and step rate re-opening tests made during mini-fracs (see section 5.5.2). Fracture closure pressures have been obtained for all the wells, and re-opening pressures for approximately half the wells. Also plotted in Figure 6.14 is all the leak-off test data available from the wells in which hydro-fracs have been performed (see section 5.5.2).

There are a number of features of this data which are immediately apparent in Figure 6.14:

- The leak-off test data down to approximately 8,000 ft (TVDSS) consist almost entirely of limit test pressures (LTPs) which increase linearly with depth.

- In the few cases where actual leak-off pressures (LOPs) have been achieved, the LOPs are indistinguishable from the trend of the LTPs.

- The most noticeable feature of this dataset is the distinct drop in pressures below approximately 8000 ft. Both the LTPs and the hydro-frac closure and re-opening pressures plot considerably below the trend of the shallower leak-off and limit test pressures (LOLTPs).

- The LTPs from below 8000 ft plot close to the higher values of hydro-frac determined closure and re-opening pressures in this depth range. As these are only LTPs, the actual LOPs could be significantly higher.
As would be expected (see section 4.5.1) the hydro-frac determined re-opening pressures lie above the hydro-frac determined closure pressures. However, neither the closure pressures nor the re-opening pressures show any obvious correlation with depth, and the range of these pressures is large (compared to the LTPs) for the given depth interval. Some possible causes of this clearly somewhat unusual distribution of stress magnitudes are discussed in Chapter 7.

All the hydro-fracs were performed in the Leman Sandstone. The leak-off tests which are below 8000 ft are all performed in the lithologies which lie below the evaporites, and above the reservoir. These lithologies include the limestones and dolomites at the base of the Zechstein Supergroup, and the siltstones and claystones of the Silver Pit formation (Figure 6.13). The leak-off tests conducted below 8000 ft, and the hydro-fracs, are separated from the shallower leak-off tests by the massive halite of the Zechstein which is well over 1000 ft thick. The leak-off tests above 8000 ft occur in a variety of lithologies, including claystones, siltstones and sandstones.

Further leak-off test data from wells in the Ravenspurn fields has been obtained from Erico Pressure Study 5. As the structure and lithostratigraphy of the Ravenspurn fields are essentially uniform, it is assumed that leak-off test data from wells adjacent to those in which hydro-fracs have been performed, can be used to compare LOPs with hydro-frac pressures. The Erico pressure atlas contains a further 19 LTPs, 10 LOPs and 2 lost circulation pressures from wells drilled in the Ravenspurn fields. The leak-off test and lost circulation data from the Erico Pressure Study has been added to the Ravenspurn data already presented, and the whole dataset is now shown in Figure 6.15.

The Erico leak-off test data from the Ravenspurn fields shows the same trend as the data plotted in Figure 6.14. The LOLTPs from depths less than 8000 ft increase linearly with depth. Again, the majority of tests have not been taken all the way to leak-off, but where leak-off has occurred, the LOPs are close to the LTPs. The Erico data includes some LOPs from below the Zechstein halite. These LOPs, like the LTPs plotted in Figure 6.14, are considerably lower than the trend of LOLTPs from above the Zechstein halite. Again, similar to the LTPs in Figure 6.14, the LOPs below the Zechstein halite plot
close to the higher values of the hydro-frac determined pressures. The lost circulation pressures also plot close to the hydro-frac determined pressures.
6.14 Data from southern North Sea Ravenspurn Fields. Hydro-frac re-opening pressures and closure pressures from Dowell Schlumberger mini-frac reports plotted with leak-off and limit test pressures from BP and BHP drilling records from the same wells.
from the Ravespum Fields obtained from the Echo Pressure Alphas. The data from Figure 6.14 is plotted here together with further leak-off test data.
It appears that in this case, LOPs can be taken as an upper estimate of fracture re-opening pressure and thus of \( \sigma_h \). This relationship is discussed further in section 6.2.3. The LTPs in this data set are also close to the fracture re-opening pressures. Generally however, where only limit test pressures are available, it would not be possible to equate them to fracture re-opening pressures, as the difference between the LTP and the pressure at which leak-off would occur, is not known.

As the LOPs from below the Zechstein halite correspond reasonably well with hydrofrac determined \( \sigma_h \), and as the tests from above and below the halite are conducted in similar rock types, it seems reasonable to assume that the LOPs from above the halite would also correspond approximately with \( \sigma_h \). If this is the case, then there is clearly a large "stress drop" below the Zechstein halite of the Ravenspurn fields. A similar drop in pressure values from leak-off tests below the Zechstein has been noted by Breckels and van Eekelen (1981) in onshore data from the Netherlands. Breckels and van Eekelen suggested that movements in the overlying halite may be responsible for the stress reduction in the underlying rocks, however it is not clear exactly what mechanical process is being suggested. A similar type of stress drop has also been detected from detailed bed by bed hydro-frac stress measurements at a depth of around 900m below the Appalachian Plateau (Evans et al., 1989; Evans et al., 1989a; Evans, 1989). The possible explanations for the Appalachian stress drop proposed by Evans et al. (1989a) include (i) a decoupling between the overlying and underlying rocks such that the overlying rocks carry more present day tectonic stress, (ii) a remnant component of stress from a palaeo-tectonic event locked into the rocks below the stress drop and acting in a direction such as to counteract the present day tectonic stress, and (iii) the most plausible explanation according to Evans et al. (1989a); the beds which today show a low stress level once hosted overpressures, which on their release, caused a contraction of the rock, inducing a tensile component of stress which today acts to reduce the total compressive stress. It is concluded however (Evans et al., 1989a) that the origins of the measured stress drop remain enigmatic. The apparent stress drop in the southern North Sea is discussed further in Chapter 7.
6.2.3 The Comparison of Leak-off Pressures and Hydro-frac Derived $\sigma_h$ Values: A Discussion.

This section discusses the relationship between the leak-off pressure and $\sigma_h$ in the light of the data presented above (sections 6.2.1 and 6.2.2) and the theoretical framework outlined in Chapter 4.

As discussed in section 4.6.5 the LOP can theoretically represent a whole range of pressures between $\sigma_h$ (in the case of pre-existing permeable cracks perpendicular to $\sigma_h$) and $[2\sigma_h - p + T]$ (in the case of an initially smooth and fracture free wellbore where $\sigma_h = \tau_h$). Because in most cases, the condition of the borehole wall is unknown during a leak-off test, it is generally not possible to be certain which breakdown equation is applicable and thus how to relate LOP to $\sigma_h$. However, the data presented in the sections 6.2.1 and 6.2.2 has been studied, firstly in a statistical framework, to establish the differences between the LOPs and hydro-frac derived $\sigma_h$ values. These pressure differences have then been examined in conjunction with the pressure/time records from leak-off tests with a view to establishing a relationship between LOP and $\sigma_h$ in cases where rock properties such as tensile strength can be estimated and the behavior of cracks at the borehole wall can be determined. On this basis, an empirical relationship between LOP and $\sigma_h$ plus tensile strength is proposed for the case where pre-existing permeable cracks perpendicular to $\sigma_h$ are present at the borehole wall.

In the light of the above comparison of LOP and $\sigma_h$, some conclusions are drawn as to the most appropriate ways to use the leak-off test data sets from the North Sea to investigate stress magnitudes.

6.2.3.1 A Statistical Comparison of Leak-off Pressures with $\sigma_h$.

Previous empirical studies (Breckels and van Eekelen, 1981) see Figure 4.14, (Amundsen, 1995; Enever et al., 1996) have shown that the trends of leak-off pressures with depth lie close to (but commonly slightly above) the trends of hydro-frac determined $\sigma_h$ with depth. Where both LOPs and hydro-frac determined $\sigma_h$ have been obtained for the same boreholes or the same area, it is possible to apply statistical techniques to determine how closely related the two sets of measurements are and to
Previous empirical studies (Breckels and van Eekelen, 1981) see Figure 4.14, (Amundsen, 1995; Enever et al., 1996) have shown that the trends of leak-off pressures with depth lie close to (but commonly slightly above) the trends of hydro-frac determined \( \sigma_h \) with depth. Where both LOPs and hydro-frac determined \( \sigma_h \) have been obtained for the same boreholes or the same area, it is possible to apply statistical techniques to determine how closely related the two sets of measurements are and to quantify the difference between the LOP at the point where the leak-off test is conducted and the value of \( \sigma_h \) at that point predicted by the surrounding hydro-frac stress measurements.

**Nirex data:**

Figure 6.8 presents all the leak-off pressures (both 1st and 2nd) from the Nirex wells together with the hydro-frac determined values of \( \sigma_h \). It can be seen from inspection of this plot that many of the 1st leak-off pressures lie above the trend of \( \sigma_h \). The 2nd LOPs however, can be seen in Figure 6.8 and 6.11 to lie much closer to the \( \sigma_h \) trend. The lines of best fit in Figure 6.11 come from linear regression analysis. The difference between any of the LOPs and the line of best fit for the hydro-frac data can be determined to estimate by how much, at that depth, the LOP exceeds the predicted \( \sigma_h \) value. This approach assumes that there is no lateral variation of \( \sigma_h \) magnitudes, i.e. that \( \sigma_h \) varies only with depth. The differences in pressure between the LOP and the predicted \( \sigma_h \) at that point based on this 1D regression analysis are listed in Table 6.2 (below).

A more rigorous approach which takes account of possible lateral variations in \( \sigma_h \) has also been applied to compare the values of the LOPs with the predicted \( \sigma_h \) values at the point of leak-off. A statistical 3D regression analysis has been performed on the hydro-frac determined \( \sigma_h \) values.

The statistical analysis is as follows. The geographical co-ordinates and depth (TVDGL) of all hydro-frac data points have been analysed using geo-statistical 3D regression analysis computer software (Howarth, 1997). A regression analysis has been applied to the hydro-frac dataset. The regression equation which defines the best fit surface to the data has the form:
The measured values of hydro-frac determined $\sigma_h$ are compared to those predicted by the regression analysis, and the differences between the two (the residuals) are examined. The residuals are seen to be normally distributed, which implies that the model is a good fit to the data (Howarth, 1997). The values of the LOPs are then also compared with the values of $\sigma_h$ predicted from the regression equation at the point of LOP. Both 1st and 2nd LOPs tend to be higher than the predicted $\sigma_h$ values based on the hydro-frac regression. However, in half of the cases where a 2nd LOP has been obtained, the 2nd LOP falls within the range of $\sigma_h$ values that are predicted with 95% confidence for that point within the 3D framework (Figure 6.16). The differences in pressure between the 3D hydro-frac predicted $\sigma_h$ values and the LOPs are presented in Table 6.2.

![Image](6.16 Observed leak-off pressures plotted against predicted $\sigma_h$ values from 3D regression analysis of hydro-frac re-opening pressures. The 95% confidence limits to the predicted $\sigma_h$ value are also shown.)
Predicted $\sigma_h$ from 3D regression analysis (MPa)

6.16 Observed leak-off pressures plotted against predicted $\sigma_h$ values from 3D regression analysis of hydro-frac re-opening pressures. The 95% confidence limits to the predicted $\sigma_h$ value are also shown.

Whereas the 1st LOPs from the Nirex dataset clearly lie above the hydro-frac determined $\sigma_h$ values, this analysis has shown that, although the 2nd LOP also tend to be slightly higher, they are generally a good estimate of $\sigma_h$. The pressure differences between the hydro-frac predicted $\sigma_h$ values and the LOPs are examined in section 6.2.3.2 in conjunction with leak-off test pressure volume plots and rock tensile strength measurements.

Southern North Sea Data:
As the dataset from the Ravenspurn field in the southern north sea also contains both hydro-frac and leak-off data, the values obtained from the two procedures can be
compared. The possibility of performing the same kind of 3D regression analysis (above) on the data from the Ravenspurn field (Figure 6.15) is considered here firstly.

As the hydro-frac step rate re-opening pressures are considered the most reliable values for estimating $\sigma_h$ (sections 4.5.1 and 5.5) these have been compared with the leak-off pressures. As there is clearly a discontinuity in the stress trend with depth at around 8000 ft in the Ravenspurn field, only the data from below this depth are considered. Because LTPs can be below $\sigma_h$, clearly only LOPs can be used. Also, the latitude and longitude of the wells was only available for 10 of the hydro-fracs. Unfortunately, these restrictions make the dataset rather small. Further problems are encountered when the geographical location of the various data points is considered. All the hydro-fracs lie more or less in a straight line which is parallel to the crestal trend of the reservoir. This means that predictions of $\sigma_h$ away from the crestal line will be subject to large uncertainty. Two of the four leak-off tests lie approximately 2 km to the side of this line.

The kind of analysis that was used on the Nirex data can not therefore be used in a meaningful way on the Ravenspurn data because of the restrictions of size and geographic location on the dataset. The limits of uncertainty associated with the predicted value of $\sigma_h$ at the points of leak-off are too large (the order of 2000 psi) to be useful.

Unlike most datasets of stress magnitude and depth, there appears to be no systematic correlation between $\sigma_h$ and depth for the hydro-frac data from Ravenspurn. The causes of the stress distribution for this part of the North Sea are discussed further in Chapter 7. This obvious lack of correlation with depth prevents a simple 1D regression analysis of the data from being useful.

The data from the Ravenspurn field do not lend themselves well to a rigorous 3D geostatistical analysis or the more simple 1D regression analysis. However, it is clear from visual inspection of the data points below 8000 ft in Figure 6.15 that the LOPs are close to the hydro-frac determined re-opening pressures. It is not intuitively obvious how to compare the two values. For this reason, a comparison has been made between the two
values and also between the two values normalised with depth, i.e. LOP/depth and re-opening pressure/depth.

The mean of the re-opening pressure values is 6649 psi with a standard deviation of 451 psi. The mean of the LOP values is 6850 psi with a standard deviation of 844 psi. The mean of the LOP values is therefore 201 psi (3%) higher than the mean of the re-opening pressure values and lies well within one standard deviation of the mean of the re-opening pressure values.

The mean of the re-opening pressure/depth values is 0.6805 psi/ft with a standard deviation of 0.049 psi/ft. The mean of the LOP/depth values is 0.732 psi/ft with a standard deviation of 0.048 psi/ft. The mean of the LOP/depth values is therefore 0.0439 psi/ft (6.4%) higher than the mean of the re-opening pressure/depth values and lies just within one standard deviation of the mean of the re-opening pressure/depth values. A difference of 0.0439 psi/ft at the reservoir depth of the Ravenspurn field is equivalent to approximately 390 psi.

It is interesting to note that the mean of the hydro-frac closure pressures is 5931 (with mean of the hydro-frac closure pressure/depth values = 0.60197 psi/ft) with a standard deviation of 474 psi (standard deviation closure pressure/depth values of 0.0568 psi/ft). The difference between the mean closure pressure and the mean re-opening pressure is thus 718 psi (or using mean closure pressure/depth and mean re-opening pressure/depth is = 0.079 psi/ft, which is equivalent to approximately 700 psi at the reservoir depth). This pressure difference is much greater than between the LOPs and the hydro-frac re-opening pressures and lies well outside of one standard deviation from the mean re-opening pressure/depth.

This simple comparison of these datasets does not take into account the geographic location of the tests, and any possible systematic variation associated with geographic location, nor does it take into account the lithology in which the tests were performed. However, bearing in mind these limitations to the analysis, the fact that there is clearly less difference between LOPs and re-opening pressures than there is between closure
pressures and re-opening pressures strongly indicates that these LOPs are a reasonable estimate of $\sigma_h$. It should be noted that, as discussed in section 4.5.1, fracture re-opening and extension pressures are expected to be higher than closure pressures due to fluid friction entering the fracture and the fracture toughness of the rock when the fracture extends into previously unfractured rock. Fracture closure pressures determined during the data-frac stage of the frac job however, are generally not the preferred method of $\sigma_h$ determination due the ambiguities associated with the large volumes of fluid used and the difficulties in picking the closure pressure (section 4.5.1).

The pressure differences between the LOPs and hydro-frac determined $\sigma_h$ values seen in the Ravenspurn dataset are considered in section 6.2.3.2 with respect to the empirical relationship proposed between LOP and $\sigma_h$.

6.2.3.2 The Effect of Rock Properties (Tensile Strength and Pre-existing Crack Characteristics) on Leak-off Pressures
The Nirex data (section 6.2.1) has shown that although there is broadly an equivalence in the trend with depth of the 1st LOPs and the hydro-frac determined $\sigma_h$ values, individual 1st LOPs can be considerably higher than $\sigma_h$. This is particularly true in rocks such as those of the Borrowdale Volcanic Group. Laboratory tests have shown (Nirex, 1996) that these rocks can have tensile strengths of up to approximately 20 MPa, although hydro-frac derived tensile strengths are much lower. The Nirex data has shown that the 2nd LOP (assuming the formation was broken down in the 1st cycle) is a good estimate of $\sigma_h$ although it still tends to give values slightly higher than hydro-frac determined $\sigma_h$ (section 6.2.3.1). In section 4.5.2.4 and section 4.6.5 - 4.6.6, the influence of pre-existing cracks in the borehole wall on the breakdown or leak-off pressure has been considered. The difference between the 1st and 2nd leak-off pressure is clearly explained in these terms.

In this section, the differences between the 1st LOP and $\sigma_h$ (\(\Delta_{1\text{st LOP}}\)), the 1st and 2nd LOP (\(\Delta_{\text{LOP}}\)), and the 2nd LOP and $\sigma_h$ (\(\Delta_{2\text{nd LOP}}\)), are investigated and discussed in terms of the possible behavior of fractures in the borehole wall and the tensile strength of the formation.
The predicted value of $\sigma_h$ at some point can be based on either the simple 1D linear regression line ($\sigma_h$ 1D), the line of best fit in Figure 6.11, in which case $\sigma_h$ must be assumed to vary only with depth, or it can be based on the 3D hydro-frac regression model ($\sigma_h$ 3D), section 6.2.3.1, which takes into account geographic location as well as depth. Predicted values based on both of these models are shown in Table 6.2 and an average of the two predicted values is also shown. This average value of predicted $\sigma_h$ ($\sigma_h^{av}$) has been used to calculate the pressure differences $A_{1st}$ LOP, $A_{LOP}$ and $A_{2nd}$ LOP.

Table 6.2 summarises the above information, together with borehole numbers (Bh.no) and the lithology in which the leak-off test was conducted.

<table>
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<th>Bh.no</th>
<th>1st LOP MPa</th>
<th>2nd LOP MPa</th>
<th>$\sigma_h$ 1D MPa</th>
<th>$\sigma_h$ 3D MPa</th>
<th>$\sigma_h^{av}$ MPa</th>
<th>$A_{LOP}$ MPa</th>
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Table 6.2 Comparison of 1st and 2nd LOPs with predicted values of $\sigma_h$ based on average of 1D and 3D regression models. BVG = Borrowdale Volcanic Group, BB = Brockram Breccia, SBS = St. Bees Sandstone, CS = Calder Sandstone.

The tensile strengths of the formations have been tested in the laboratory using the Brazilian test on core samples taken from the hydro-frac intervals (Nirex, 1996). Within the commonly occurring rocks, the Borrowdale Volcanic Group (BVG) and the St. Bees Sandstone (SBS), a range of tensile strengths have been found. A very large range of tensile strength values have been measured for the rocks of the BVG. The BVG consists of a variety of rocks including tuffs, andesites and volcanic breccias, each of which
presumably have different tensile strengths. Unfortunately, the exact lithology at the
casing shoes where the leak-off tests were performed within the BVG is not available.
The range and mean of the laboratory measured tensile strengths (Nirex, 1996) for the
various lithologies is as follows:

- **Tensile Strength/MPa (BVG):** 3.1 - 24.9 (mean = 12.3)
- **Tensile Strength/MPa (SBS):** 0.7 - 6.1 (mean = 3.8)
- **Tensile Strength/MPa (BB):** 11.0 (only one value available)
- **Tensile Strength/MPa (CS):** no measurements.

Tensile strength has also been calculated from hydro-frac records by comparing the
breakdown pressure with the first fast re-frac re-opening pressure (section 4.5). The
values derived from this method are as follows:

- **Tensile Strength/MPa (BVG):** 0.6 - 6.4 (mean = 2.7)
- **Tensile Strength/MPa (SBS):** 2.1 - 6.8 (mean = 4.2)
- **Tensile Strength/MPa (BB):** 9.6 (only one value available)
- **Tensile Strength/MPa (CS):** no measurements.

The hydro-frac derived tensile strengths of the BVG rocks are clearly much less than the
lab derived values. The lab derived and hydro-frac derived values for the SBS and the
BB are very similar. The appropriate value of tensile strength is of crucial importance in
the determination of \( \sigma_H \) from the classical breakdown equation (section 4.5.2) and the
subject has received considerable attention in the hydro-frac literature which has been
summarised by Amadei and Stephansson (1997). The most likely explanation for the
difference in apparent tensile strengths in the BVG rocks is that the size of the samples
being tested in the lab are smaller than the volume of rock around the wellbore which is
stressed by the fracturing fluid during the build up to break down. This effect is known
as the size effect. The size effects is essentially that the apparent tensile strength
measured on a rock sample is very strongly dependent on the size of the sample. The
apparent tensile strength is related to the size of the largest, favorably oriented, pre-
existing crack, such that when large cracks are present the apparent tensile strength is
very low, and when small cracks are present the apparent tensile strength is very high.
The larger the sample tested, the greater the size of the largest favorably oriented crack is likely to be and thus the lower the apparent tensile strength.

The core samples retrieved from the Nirex boreholes were up to 5 inches in diameter and thus the maximum crack length in the core being tested would be less than 5 inches. The interval of borehole exposed to the fracturing fluid during hydro-frac testing in the Nirex boreholes was of the order of several metres and thus it is likely that some cracks of length greater than the maximum crack length in the core will exist. The hydro-frac intervals are chosen on the basis that they are intact (i.e. they contain no pre-existing fractures), however, this is simply done using borehole imaging techniques which are clearly only able to detect the larger cracks present where they are partly open at the borehole wall. It is not surprising that many small cracks, or large cracks which are mechanically closed at the borehole wall, remain undetected. On the other hand, when core is recovered, visual inspection and handling of the core can easily detect the presence of most cracks such that a very undamaged piece is likely to be chosen for laboratory testing.

The values of the tensile strength measured in the lab for the SBS are very similar to the hydro-frac measured values. Both values are much lower than the lab derived tensile strength for the BVG. The fact that the two methods yield similar values for the SBS can have two possible explanations: (a) there are no cracks in either the core or the borehole section tested, such that both methods sample the intrinsic tensile strength of the material (b) as the rock has a low intrinsic tensile strength, the measured tensile strength is only reduced by cracks up to a certain length, i.e. the larger cracks in the borehole section do not reduce the tensile strength any further than the smaller cracks present in the core, or in other words, the size effect only holds up to a certain crack length for this rock.

Explanation (a), that there are no cracks at all in the SBS is considered unlikely, on the basis of the geological history, the generally faulted nature of the rock at Sellafield, the core descriptions and borehole image descriptions contained in the Nirex reports (e.g. Nirex, 1996).
\[ \Delta_{1st\ LOP\ (BVG)/MPa} = 9.2 \ (borehole\ 8a),\ 8.0 \ (borehole\ 10a),\ 8.6 \ (average) \]

\[ \Delta_{LOP\ (BVG)/MPa} = 1.7 \ (borehole\ 8a),\ 5.1 \ (borehole\ 10a),\ 3.4 \ (average) \]

\[ \Delta_{2nd\ LOP\ (BVG)/MPa} = 7.5 \ (borehole\ 8a),\ 2.9 \ (borehole\ 10a),\ 5.2 \ (average) \]

The mean hydro-frac derived tensile strength of the BVG is 2.7 MPa.

All the results for the BB come from just one hydro-frac and one leak-off. In the BB, the pressure differences are:

\[ \Delta_{1st\ LOP\ (BB)/MPa} = 7.0 \ \text{MPa} \]

\[ \Delta_{LOP\ (BB)/MPa} = 5.2 \ \text{MPa}. \]

\[ \Delta_{2nd\ LOP\ (BB)/MPa} = 1.8 \ \text{MPa} \]

The hydro-frac derived tensile strength for the BB is 9.6 MPa.

The pressure difference \( \Delta_{1st\ LOP} \) in the BVG in both boreholes is considerably greater than the average or indeed any of the hydro-frac determined values of tensile strength for the BVG. Clearly then, the 1st LOP in the BVG tests is either being influenced by the hoop stress such that the LOP is equivalent to the breakdown pressure in the classical breakdown equation (section 4.6.5), or the LOP represents the pressure needed to open some pre-existing crack which is not oriented perpendicular to \( \sigma_h \), or a combination of both these effects.

In borehole 10a \( \Delta_{LOP} \) is larger than the average hydro-frac determined tensile strength, but \( \Delta_{2nd\ LOP} \) is small. This leak-off test may represent the classical behavior expected during a hydro-frac. That is, the fracture opens perpendicular to \( \sigma_h \) during the 1st cycle, overcoming the tensile strength of the formation and also the hoop stress, the 2nd LOP represents a slow re-frac and its pressure is thus close to \( \sigma_h \). The value of \( \Delta_{LOP} \) is thus higher than the tensile strength. The pressure/time record of this test has been presented earlier (Figure 6.10). The drop in pressure after the peak pressure in the 1st cycle is indicative of a classical breakdown, that is, a breakdown pressure that is higher than the
to open some pre-existing crack which is not oriented perpendicular to $\sigma_h$, or a combination of both these effects.

In borehole 10a $\Delta$LOP is larger than the average hydro-frac determined tensile strength, but $\Delta$2nd LOP is small. This leak-off test may represent the classical behavior expected during a hydro-frac. That is, the fracture opens perpendicular to $\sigma_h$ during the 1st cycle, overcoming the tensile strength of the formation and also the hoop stress, the 2nd LOP represents a slow re-frac and its pressure is thus close to $\sigma_h$. The value of $\Delta$LOP is thus higher than the tensile strength. The pressure/time record of this test has been presented earlier (Figure 6.10). The drop in pressure after the peak pressure in the 1st cycle is indicative of a classical breakdown, that is, a breakdown pressure that is higher than the pressure needed to propagate the fracture. The breakdown pressure is higher than the propagation pressure because it must overcome the tensile strength and the hoop stress.

In the BB, both values of tensile strength are higher than all pressure differences. However, $\Delta$1st LOP is only 2 MPa less than the hydro-frac derived tensile strength, and as $\Delta$LOP is also large and $\Delta$2nd LOP is small, it would seem that a fracture was propagated perpendicular to $\sigma_h$ in the 1st cycle and the tensile strength of the formation was mainly responsible for the large value of $\Delta$1st LOP. The 2nd LOP would then be equivalent to a slow re-frac and thus its pressure is close to $\sigma_h$. The pressure/time record for this leak-off test has the same form as that from borehole 10a.

In borehole 8a, $\Delta$1st LOP is considerably larger than the hydro-frac derived tensile strength of the rock. $\Delta$LOP is of the order of the hydro-frac determined tensile strength indicating that a fracture may have been extended during the 1st cycle, but $\Delta$2nd LOP is still large indicating that if a fracture had been extended during the 1st cycle, it either closed perfectly such that the 2nd LOP is equivalent to a fast re-frac (this is unlikely given the slow pump rate) or the fracture is not oriented perpendicular to $\sigma_h$. The pressure/time record for this test does not show a drop in pressure after the peak pressure, and thus this leak-off test can be interpreted as re-opening a pre-existing crack, in the manner of a slow re-frac. The 2nd LOP is lower than the 1st because the crack,
having been opened once, is more permeable during the 2nd cycle and the rock (out to
the limit of crack growth in the 1st cycle) has lost any tensile strength that it initially had.

Conclusions from the BVG and BB data are as follows: In two out of the three leak-off
tests in these rocks, the 1st LOP is influenced by the hydro-frac tensile strength and the
hoop stress, producing a sharp drop in pressure after the peak in the leak-off test
pressure/time record and a 1st LOP much higher than $\sigma_h$. The 2nd LOP from these tests
is a much better estimate of $\sigma_h$ and it is concluded that the 2nd LOP represents a re-
opening pressure where the fracture has been pressurized along its length such that that
it is not influenced by the hoop stress and the tensile strength effect is small.

The other leak-off test conducted in these rocks re-opens a pre-existing fracture that is
not oriented perpendicular to $\sigma_h$ and thus both 1st and 2nd LOPs are higher than $\sigma_h$. It is
thought that this is more likely to occur in rocks which have high intrinsic tensile
strengths.

**St. Bees and Calder Sandstone:**
As the pressure differences ($\Delta$LOPs, $\Delta$1st LOP, and $\Delta$2nd LOP) in the SBS and CS are
all less than 3.7 MPa, their mean values from the 5 leak-off tests in these lithologies are
presented below.

$\Delta$LOP (SBS) = 1.0 MPa (average)
$\Delta$1st LOP (SBS) = 1.8 MPa (average)
$\Delta$2nd LOP (SBS) = 0.9 MPa (average)

These pressure differences are generally considerably smaller than those seen in the
BVG and BB rocks. The difference in $\Delta$LOPs values between the BVG and the
sandstones can be explained easily if the lab derived tensile strengths of the different
rock types are used. Despite the fact that the hydro-frac derived tensile strength values
for BVG are actually lower than the hydro-frac derived values for sandstones, it seems
likely that the high intrinsic (lab derived) tensile strengths of the BVG had some effect
on the tensile strengths in the BVG rocks in which the leak-off tests were performed.
However, if the 1st LOPs in sandstones simply represent a classical breakdown but in a
formation with a low tensile strength, the hoop stress would still be expected to cause
the LOP to be considerably higher than $\sigma_h$. The classical breakdown equation with zero
tensile strength predicts a breakdown pressure of up to $2\sigma_h - p$, decreasing with
increasing horizontal stress anisotropy. Such a breakdown would be expected to
produce a pressure/time plot with a maximum pressure at breakdown and a sharp
pressure drop once the fracture was initiated. As discussed below, this type of
pressure/time plot is seen for two of the leak-off tests in the BVG but not for any of the
sandstones.

The other explanation for the lower values of $\Delta$LOP and $\Delta$1st LOP in the sandstones, is
either that pre-existing cracks in these rocks are more common than in the BVG and BB
or that they are more permeable to the drilling mud during the 1st leak-off test
pressurisation. These cracks become pressurised along their length during the leak-off
test and thus open when the pressure exceeds $\sigma_h$. The 1st LOP is therefore close to $\sigma_h$.
Once the crack has opened and closed in the first cycle, it will be more permeable in the
2nd cycle and thus the 2nd LOP will be even closer to $\sigma_h$.

The pressure/time plots of the leak-off tests performed in the sandstones are typically
rather different to the pressure time plot expected from a classical breakdown (such as
the two described above in the BVG and BB). The pressure/volume plots from leak-off
tests in the sandstones mostly show a linear build up in pressure but none show the sharp
peak and rapid drop indicative of classical formation breakdown. The pressure/time
plots from the leak-off tests in the Nirex sandstones look similar to typical pressure/time
plots such as those of Figure 4.12 and 4.16 (a) - (d). Figure 6.17 shows a copy of the
actual pressure/time record from the 1st leak-off test in borehole 11a at 200m depth.
The type of pressure/time plot shown for the leak-off test in the sandstones is consistent
with the opening of cracks due to pressurisation along their lengths. In such a case, the
hoop stress has only a small effect on the LOP, as does the tensile strength, and so the
pressure after leak-off does not drop sharply.
6.17 Sellafield borehole 11A. Leak-off test at 200m. 1st pressurisation cycle. Surface pressure/time plot from Nirex records showing a shape characteristic of fracture re-opening at approximately 600 psi.

It should be remembered that the pre-existing crack opened may not necessarily be perpendicular to $\sigma_h$ and thus the opening pressure may be between $\sigma_h$ and $\sigma_H$. However, in rocks with low intrinsic tensile strength, it is probable that smaller pre-existing cracks which are oriented perpendicular to $\sigma_h$, would open in preference to larger pre-existing cracks in an unfavorable orientation. This explains the difference in behaviour between the test in the BVG rock which seems to show re-opening of a crack not oriented perpendicular to $\sigma_h$ (borehole 8a) and the tests from the SBSs. Also, if the size of open hole section being tested is large (as is the case with leak-off tests - particularly in the North Sea, where the open hole section is typically 10m in length and generally between 0.3 and 0.1m in diameter), the probability of encountering a favorably oriented crack in all but the most intact rocks is high.

If it is true that leak-off tests in sandstones tend to re-open pre-existing cracks and thus display no sharp peaks in their pressure/time records, then it is implied that as peaks are seen in 2 out of 3 the pressure/time records from the leak-off tests in BVG and BB rocks, these tests have tended not to re-open pre-existing cracks or at least that if pre-existing cracks are opened in the BVG tests, then they have not been pressurised along their lengths prior to leak-off. This is somewhat contrary to the evidence from the
hydro-frac derived tensile strengths in the BVG, which imply that there must be cracks in the borehole wall in order to reduce the hydro-frac tensile strength so far below the lab derived value. It would appear then, that the cracks in the BVG which cause the low hydro-frac derived tensile strengths are often not permeable to the drilling mud during pressurisation. It would be interesting to devise some laboratory tests to assess the permeability of cracks in the BVG compared to those in the SBS under in-situ conditions, although such tests are somewhat beyond the scope of this study.

If it is assumed these LOPs in the SBS represent the opening pressures of fractures which have been pressurised along their lengths, such that the hoop stress can be neglected, it follows that the positive values of \( \Delta 1st \) LOP and \( \Delta 2nd \) LOP could be largely due to frictional losses incurred from forcing the fluid into the crack towards the crack tip, and the inherent resistance to fracturing at the crack tip. As the pumping rates used in these leak-off tests are low, the pressure differences from frictional losses are assumed to be small. The values of \( \Delta 1st \) LOP and \( \Delta 2nd \) LOP can therefore be compared to the fracturing resistance of the rock. This comparison would be best performed within a fracture mechanics framework (e.g. section 4.5.2.4) with values of fracture toughness. The only comparable measurement available however is the tensile strengths of the rocks.

The hydro-frac and lab derived tensile strengths of the SBS are close, and have a low value of approximately 4 MPa. The mean \( \Delta 1st \) LOP value for the sandstones is approximately half the value of the tensile strength, and the mean \( \Delta 2nd \) LOP for the sandstones is approximately one quarter of the tensile strength. The maximum \( \Delta 1st \) LOP is 2.7 MPa, which could be of the order of the tensile strength of the rock. On the basis of these comparisons, the following empirical relationship appears to hold for the SBS:

\[
1st \text{ LOP} = \sigma_t + \alpha_t T
\]

Where \( \alpha_t \) is a co-efficient of the tensile strength which can vary between 0 and 1.

This relationship will hold if there are pre-existing permeable cracks oriented perpendicular to \( \sigma_t \) and of sufficient length to prevent the LOP from being influenced by
the hoop stress, and if the intrinsic tensile strength of the formation can be considered an upper limit of the resistance to tensile fracturing inherent in the material. The co-efficient \( \alpha_i \) is used to describe the empirical relationships seen from this dataset. It has no strictly definable physical meaning but if the above assumptions are satisfied, it must depend on the permeable length of the pre-existing crack, the frictional resistance to fluid flow in the crack (which should be low at the low pressurisation rates used in leak-off tests), and what proportion of the inherent tensile strength is required to drive a tensile fracture through the material. The co-efficient \( \alpha_i \) would be close to zero in the case of a very long crack in which there was no frictional resistance to fluid flow such that all the pressure in the wellbore is transmitted to the crack tip.

The similarity between the relationship suggested above and a similar relationship defined in fracture mechanics terms and described by equation 4.27 should be noted. It is acknowledged that a fracture mechanics approach to defining a relationship such as the one shown above, would probably be more suitable. However, in the absence of any information on fracture toughness for these rocks, such an approach is not possible.

*Ravenspurn LOPs*

The LOPs shown in Figure 6.15 below the stress drop are all in siltstones of the Rotliegend group. Based on the statistical analysis of these values in section 6.2.3.1, it was concluded that on average, the LOPs in these were between 200 and 390 psi (1.4 and 2.7 MPa) higher than the \( \sigma_h \) values based on hydro-frac step rate re-opening tests in the adjacent reservoir. Assuming a tensile strength for siltstones that is of the order of the values measured for the sandstones above (this seems reasonable on the basis of many laboratory measurements reported by Lama and Vutukuri, 1978) these LOP values are in reasonable with the empirical relationship proposed above between LOP and \( \sigma_h \) seen in the sandstones of the Nirex boreholes. Assuming tensile strength values of MPa for these rocks, the data implies that the average value of \( \alpha_i \) would be approximately 0.5.

6.2.3.3 The Use of Leak-off Data to Estimate \( \sigma_h \): Conclusions
Rocks of the BVG and the BB appear to yield LOPs which are ambiguous to interpret in terms of $\sigma_h$. This is because pre-existing cracks in these rocks do not generally appear to be permeable to the drilling mud during pressurisation in the leak-off test. Also, the rocks have a high intrinsic tensile strength which favours the opening of pre-existing cracks, which are not oriented perpendicular to $\sigma_h$, over the opening of smaller pre-existing cracks, which are oriented perpendicular to $\sigma_h$.

The 1st LOPs from leak-off tests performed by Nirex in sandstones appear to be less ambiguous. Both the 1st and 2nd LOPs are seen to be close to the hydro-frac predicted value of $\sigma_h$, with the 2nd LOPs being the closest. The average difference between the 1st LOP and $\sigma_h$ in these tests is 1.8 MPa. The fact that this difference is so small implies that (unless the horizontal stresses are highly anisotropic) the hoop stress is probably not influencing the LOP. This is supported by the pressure/time plots from the 1st leak-off cycles which show no sharp drop in pressure after peak pressure. The type of pressure build up seen in these tests is indicative of fractures re-opening when pressurised along their length.

The following empirical relationship has been proposed for LOPs from the Nirex Sandstones:

$$1\text{st LOP} = \sigma_h + \alpha_1 T$$

where $\alpha_1$ can vary between 0 and 1.

This empirical relationship also holds for the data from the Ravenspurn field in the southern North Sea where the average value of $\alpha_1$ would be approximately 0.5. Clearly, the relationship given above is based on few measurements. Further measurements of a similar type, if they exist, would greatly enhance any further studies of this nature.

As mentioned at the start of this section, and as discussed in section 4.6.5 the LOP can theoretically represent a whole range of pressures between $\sigma_h$ (in the case of pre-existing permeable cracks perpendicular to $\sigma_h$) and $2\sigma_h - \rho + T$ (in the case on an initially intact wellbore where $\sigma_h = \sigma_h$). Therefore because in most cases, the condition of the borehole wall is unknown during a leak-off test, it is generally not possible to be certain which breakdown equation is applicable and thus how to relate LOP to $\sigma_h$. The ambiguities
inherent in the interpreting LOPs, particularly where just the 1st LOP is present (such as in a standard leak-off test) have been demonstrated above, and thus the empirical relationship proposed for the Nirex sandstone data presented above is not recommended for widespread use. However, where the pressure/time records from a leak-off test are available, considerably more confidence can be placed in the interpretation of 1st LOP in terms of $\sigma_h$. Where the pressure/time record indicates the opening of a pre-existing crack (e.g. plots (a)-(d) in Figure 4.16), in a rock in which the resistance to fracturing is not high (as would be the case for most sedimentary rocks) the empirical relationship presented above can be used to give a first estimate of $\sigma_h$.

In the light of the above conclusions, the most appropriate way to use the leak-off test datasets from the North Sea can now be decided. As the pressure/volume plots for the Erico data are not available, it is not possible to be confident that all of these pressure represent the type of crack re-opening that has been interpreted for the Nirex sandstone data. Although it is thought that most leak-off tests conducted in large open hole volumes of sedimentary rock will yield 1st LOPs close to $\sigma_h$, it is likely that some, probably in formations with high tensile strengths or few permeable cracks, will either open unfavorably oriented cracks, or be influenced by the hoop stress.

However, as most LOPs will be close to $\sigma_h$, it is concluded that the trends in the Erico LOPs will closely reflect trends in $\sigma_h$ magnitudes in the North Sea. There are therefore two ways to use the Erico dataset is to investigate the trends of $\sigma_h$ with depth for the North Sea. Firstly the general trends and how these trends vary according to parameters such as geographic location, lithology, and pore pressure can be investigated using the whole dataset such as in section 6.3. Secondly, the actual trend of $\sigma_h$ with depth for a particular lithology (where there are enough tests performed in that lithology) can obtained from the Erico dataset by selecting the lower bound to a plot of LOP vs. depth as it is the data points which define the lower bound which will be equal to $\sigma_h$. This is discussed further and results are presented in section 7.2. Such trends can be compared to trends with depth of $\sigma_h$ from other parts of the world (section 7.5).
The dataset from Amerada Hess on the other hand does contain leak-off pressure/time (actually mainly pressure/volume with constant pump rate) plots. These plots have already been discussed in section 4.6.6.1. They have been examined carefully and where the shape of the pressure/volume plot indicates fracture re-opening, absolute estimates of $\sigma_h$ have been made. This is discussed further in section 6.4.2. The data can then also be used to investigate the trends of $\sigma_h$ with depth. Section 6.4 presents the results from the Amerada Hess wells.

Where pressure/volume plots are available, such that absolute values of $\sigma_h$ can be obtained, and where borehole breakouts have also been detected in the same section of borehole, the stress analysis can be taken one step further. By assuming some reasonable values for rock strength properties and by adopting an appropriate rock failure criterion, the lower limit of the maximum horizontal stress ($\sigma_h$) can be determined. A preliminary investigation using this method is presented in Section 7.4.

Values of $\sigma_h$ determined from leak-off tests with pressure/volume records and trends of $\sigma_h$ with depth from the Erico dataset can be compared to values and trends predicted by models of stress in sedimentary basins (section 3.4) in order to test the applicability of these models. This is fully discussed in section 7.3.

### 6.3 Variation of North Sea Leak-off Pressures from the Erico Pressure Studies

This section presents the results of the main bulk of the data analysed in this study, that is the leak-off pressures from throughout the North Sea which have been obtained from the Erico Pressure Studies. The leak-off test data is firstly presented as a whole, plotted against depth. The data is then analysed in more detail by subdividing the data firstly according to its geographical/structural setting (as defined in section 5.3), and secondly on the basis of the lithology in which the tests were performed. The data is also presented on the basis of estimates (or measurements where available) of the pore pressure of the formation in which the leak-off test was conducted.

In view of the discussion above (section 6.2.3.3) the leak-off test data presented in this section is not interpreted in terms of $\sigma_h$ but rather presented just as “raw” data.
However, it can be assumed that the trends of these datasets with depth do (in most cases) reflect the trends of $\sigma_s$ with depth.

### 6.3.1 All Erico Leak-off, Limit Test and Lost Circulation Pressures

Over 2900 leak-off, limit test and lost circulation pressures have been compiled in the Erico North Sea Pressure Studies obtained for this study (see section 5.2). This data covers the northern, central and southern North Sea. Figure 6.18 shows all the data plotted against depth (TVDSS).

From inspection, the trend of the LOPs with depth appears the same as the trend of the LTPs. As LTPs are the pressures from tests which have not been taken to leak-off (section 4.6), it might be expected that they would on average show lower values than the LOPs. This would be the case if all the tests were conducted in the same lithology in a horizontally uniform stress field (i.e. a horizontal stress field that only increases with depth). On the other hand, given that the properties of formations are probably not homogeneous, and also that the stress field itself is probably not homogeneous, LTPs might represent those tests which have been conducted in formations with particularly high tensile strengths, or with few pre-existing cracks, such that the LOP would be high (section 6.2.3.2). Alternatively these tests may have been conducted in formations which, for whatever reason, have high horizontal stress magnitudes. In such formations, it is expected that the leak-off pressure would be higher than in other formations. In fact, as discussed in section 6.2.3, there are many combinations of conditions which can give rise to high LOPs. When conducting leak-off tests in these formations, once a pressure adequate to drill the next section is estimated to have been achieved, the test may often be stopped before leak-off has occurred (see section 4.6.2). These LTPs may then be as high as LOPs in other weaker formations or formations with lower stress magnitudes.

The LTPs present some ambiguity, however, because of their apparently close correspondence to LOPs, and moreover because they form the majority of the dataset, they have been included in most of the analyses presented in this section.

The average trend of the lost circulation pressures (LCPs) is significantly lower than the trend of the LOPs and LTPs. Lost circulation occurs accidentally while drilling. It can be caused by drilling into a highly permeable zone (perhaps where naturally open fractures,
or a vuggy limestone exists) or by accidentally inducing a hydraulic fracture in the formation being drilled. If the lost circulation is caused by accidentally inducing a hydraulic fracture, then the lost circulation pressure is equivalent to a leak-off pressure and it could be inferred that the very low values of lost circulation pressure indicate very low values of $\sigma_h$. The cause of lost circulation can sometimes be determined by the drilling engineers through examination of the cuttings from the formation being drilled, or by examination of logs run after drilling. However, such determinations can often remain somewhat speculative, and have generally not been included in the Erico Pressure Studies. The inference that LCPs are equivalent to LOPs can therefore not be made here. LCPs are considered further, in terms of the lithologies in which they occur, in section 6.3.2.2. but for most of the results presented here, LCPs have not been included.

The most obvious feature of Figure 6.18 is that the LOPs, LTPs and LCPs, as would be expected, increase with depth. This increase however, is non-linear. In fact, the data seem to fall along a curve, the gradient of which increases with depth. This increase in gradient with depth will be illustrated further in the following sections and the possible causes of this type of relationship will be discussed (Section 6.4.2 and Chapter 7).
6.3.2 Variation of Erico Leak-off Pressures with Depth and other Factors

The data displayed in Figure 6.18 increases in a non-linear way with depth. The change in gradient with depth can be shown by plotting the LOP divided by depth ("LOP gradient") plotted against depth. The term leak-off pressure gradient or fracture gradient is often used in oil industry drilling literature to mean simply leak-off pressure (or fracture pressure) divided by depth. In the case where the leak-off pressure does not increase linearly with depth, this is not the same as a gradient in the usual sense of the word (i.e. it is not equal to the tangent of the leak-off pressure vs. depth plot and is thus not a tangent gradient) but is actually a secant gradient. The secant is the line between the origin and the co-ordinate of the data point in question. The secant LOP gradient, referred to in the rest of this chapter as simply the LOP gradient is thus defined:

\[
\text{LOP gradient} = \frac{\text{LOP (psi)}}{\text{Depth (ft TVDSS)}} \tag{6.1}
\]

The use of the secant gradient is considered the most practical meaningful way of displaying this data. The secant gradient takes account of the shape of the curve at points between the origin and the point at which the secant gradient is calculated. Thus the secant gradient is equivalent to an average gradient for that section of curve. The tangent gradient on the other hand is an "instantaneous" gradient which only measures the gradient at that point.

The LTP gradient and LCP gradient can also be defined as the pressure divided by the depth below sea level, i.e. the secant gradients. Figure 6.19 uses the same data as Figure 6.18, only the data are now displayed as gradients. It is clear from Figure 6.19 that the gradients of the LOPs, LTPs and LCPs, when all the data from the North Sea Pressures Studies is plotted together, increase significantly with depth. Linear regression analysis has been applied to the data to produce the lines of best fit. For the purposes of a quantitative comparison between various trends displayed in this chapter, some of the statistical attributes of the data are summarised at the end of this chapter in Table 6.5. The statistics listed for the data are (a) the gradient of the linear best fit line (b) the intercept of this line with the depth = 0 line (c) the mean of the pressure gradients (d) the standard deviation of the pressure gradients.
6.19 Erco LOPs LTPs and LCPs displayed as gradients (e.g. LOP gradient = LOP/depth) against depth. Linear best fits (Table 6.5) have been added to highlight the comparative trends of the datasets.
As there is clearly a lot of scatter in the LOP, LTP and LCP data, variations of the pressure data with factors other than depth have been investigated. Based on an understanding of crustal stresses (Chapter 3) the factors which have been chosen here, to subdivide the pressure data by are:

Geographic/Structural Domain, Lithology, and Formation Pore Pressure. In these analyses, the LCP data has not been included, as generally the cause of lost circulation is not known.

6.3.2.1 Variation of Leak-off and Limit Test Pressures with Geographic/Structural Domain

This section examines the leak-off test data in order to investigate the influence of location and structural setting on the trend of the leak-off and limit test pressures with depth. The geographical/structural grouping of the Erico Pressures Studies data (described in section 5.3) simply represents a “first order” division of the dataset. The data has been divided into three groups: the northern North Sea, the central North Sea and the southern North Sea. Figure 6.20 shows the leak-off and limit test pressure (LOLTP) gradients for these domains. The lines of best fit indicate that the average LOLTP gradient increases with depth at approximately the same rate in the central and northern North Sea. The average LOLTP gradient does not increase significantly with depth in the southern North Sea. The amount of scatter is greater in the southern North Sea dataset.

Figure 6.21 shows the same plot as Figure 6.20 but with LOP gradients only. Although this dataset is considerably smaller than that in Figure 6.20, almost exactly the same trends in the data are apparent. The similarity between Figure 6.20 and 6.21 is evidenced by the equations of the lines of best fit (the gradients and intercepts are listed in Table 6.5 for comparison). This similarity indicates (as seen in the dataset as a whole, before division into geographic domains i.e. Figures 6.18 and 6.19) a close correspondence between the trends of the LTP and LOP gradient with depth.
Figure 6.20 Enco LOLTP Gradients from the North Sea displayed by geographic location of the dataset. Linear best fits (Table 6.5) have been added to highlight the comparable trends.
Figure 6.21 The dataset shown in Figure 6.20 is shown here without the LTP.
The term limestone simply means that the rock is predominately calcium carbonate. Calcium carbonate however, has a very wide range of rock forming modes e.g.: fossiliferous and shelly, fine grained brittle micritic, very porous and bedded oolitic. It seems likely then that limestones will have a very wide range of geomechanical properties, and for that reason, the trend of the limestone LOLTP gradient should not be over interpreted but should be treated with caution.

<table>
<thead>
<tr>
<th>Main Lithology in Open Hole Interval Tested</th>
<th>Number of Tests (Northern and Central North Sea)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Shale</td>
<td>849</td>
</tr>
<tr>
<td>Silt</td>
<td>74</td>
</tr>
<tr>
<td>Sand</td>
<td>228</td>
</tr>
<tr>
<td>Limestone</td>
<td>129</td>
</tr>
<tr>
<td>Chalk</td>
<td>80</td>
</tr>
<tr>
<td>Marl</td>
<td>54</td>
</tr>
<tr>
<td>Dolomite</td>
<td>9</td>
</tr>
<tr>
<td>Anhydrite</td>
<td>3</td>
</tr>
<tr>
<td>Salt</td>
<td>2</td>
</tr>
</tbody>
</table>

Table 6.3 Lithologies in which LOPs and LTPs are recorded in Erico Pressure Studies of the central and northern North Sea.
The term limestone simply means that the rock is predominately calcium carbonate. Calcium carbonate however, has a very wide range of rock forming modes e.g.: fossiliferous and shelly, fine grained brittle micritic, very porous and bedded oolitic. It seems likely then that limestones will have a very wide range of geomechanical properties, and for that reason, the trend of the limestone LOLTP gradient should not be over interpreted but should be treated with caution.

<table>
<thead>
<tr>
<th>Main Lithology in Open Hole Interval</th>
<th>Number of Tests (Northern and Central North Sea)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Shale</td>
<td>849</td>
</tr>
<tr>
<td>Silt</td>
<td>74</td>
</tr>
<tr>
<td>Sand</td>
<td>228</td>
</tr>
<tr>
<td>Limestone</td>
<td>129</td>
</tr>
<tr>
<td>Chalk</td>
<td>80</td>
</tr>
<tr>
<td>Marl</td>
<td>54</td>
</tr>
<tr>
<td>Dolomite</td>
<td>9</td>
</tr>
<tr>
<td>Anhydrite</td>
<td>3</td>
</tr>
<tr>
<td>Salt</td>
<td>2</td>
</tr>
</tbody>
</table>

Table 6.3 Lithologies in which LOPs and LTPs are recorded in Erico Pressure Studies of the central and northern North Sea.
Figure 6.22 EICO LOLTP gradients from the northern and central North Sea displayed by lithology.
Figure 6.23 The dataset shown in Figure 6.22 is shown here without the data from the minor lithologies. Linear best fits (Table 6.5) have been added to highlight the comparative trends of the datasets.
In fact the lithology groups in Figures 6.22 and 6.23, are all to some extent mixed lithologies (i.e. the sand group consists of pure sand, shaley sand, silty sand etc.). Therefore, any conclusions concerning the geomechanical properties of the rocks and the level of stress magnitude sampled by the leak-off test, may be somewhat clouded. For this reason, the datasets have been reduced to the results of tests conducted in "pure" lithologies, where "pure" lithologies are those which have been listed in the pressure atlas as just one lithology (section 5.2). Because of the ambiguities mentioned above concerning small datasets and the possible variety of geomechanical properties being represented by a lithological group, only the "pure sand" and "pure shale" data is displayed.

Figure 6.24 shows the LOLTP gradients from the central and northern North Sea from tests conducted in pure sand and pure shale where linear best fit trends have been added. There are results from 545 tests conducted in pure shales and from 179 tests conducted in pure sand. Again, Table 6.5 summarises the slope and intercepts of these lines for comparison with the other data presentations. The trends of the LOLTP gradients with depth are similar to those shown in Figure 6.23.

Although there is clearly some scatter in the LOLTP gradients within each lithological group, the trends indicate that generally the LOLTP gradients are higher in shales than in sands. This is most clearly illustrated by considering only the data from tests conducted in "pure" lithologies (Figure 6.24). There may also be systematic differences in LOLTP gradients within other lithologies. These could be illustrated more clearly if a larger number of better defined lithologies were available.
Figure 6.24 Erico LOLTP from the central and northern North Sea from tests performed in “pure” sands and shales. Linear best fits (Table 6.5) have been added to highlight the comparative trends of the datasets.
**LOLTP Gradients in the lithologies of the Southern North Sea.**

The southern North Sea dataset has been treated separately because of the differences in LOLTP trends shown in Figure 6.20. Table 6.4 shows the number of tests performed in each lithology, where in this case, only “pure lithologies” have been included. It can be seen that, as in the rest of the North Sea, the most common lithology in which tests have been performed is shale. However, unlike the rest of the North Sea, few tests have been performed in sand, but many more have been performed in anhydrite and salt.

<table>
<thead>
<tr>
<th>Lithology in Open Hole Interval Tested</th>
<th>Number of Tests (Southern North Sea)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Shale</td>
<td>165</td>
</tr>
<tr>
<td>Silt</td>
<td>18</td>
</tr>
<tr>
<td>Sand</td>
<td>43</td>
</tr>
<tr>
<td>Limestone</td>
<td>11</td>
</tr>
<tr>
<td>Chalk</td>
<td>40</td>
</tr>
<tr>
<td>Marl</td>
<td>4</td>
</tr>
<tr>
<td>Dolomite</td>
<td>15</td>
</tr>
<tr>
<td>Anhydrite</td>
<td>32</td>
</tr>
<tr>
<td>Salt</td>
<td>62</td>
</tr>
</tbody>
</table>

Table 6.4 Lithologies in which LOPs and LTPs are recorded in Erico Pressure Study of the southern North Sea.

The LOLTP gradients from the southern North Sea have been grouped by lithology and plotted against depth in Figure 6.25. As was seen in Figure 6.20, the trend of the dataset as a whole does not increase with depth, and there appears to be more scatter than in the data from the rest of the North Sea. Also, some of the lithologies appear to be largely bunched into depth ranges, e.g. nearly all the tests in chalk are at shallow depths and most of the tests in salt are at greater depths. This simply reflects the structure and stratigraphy of much of the southern North Sea gas basin, in which the Permian evaporites are overlain by up to a few thousand feet of Mesozoic sediments (see for example the lithostratigraphy of the Ravenspurn Field, Figure 6.13).
Figure 6.25: LOOTP gradients from the southern North Sea displayed by lithology, where only tests performed in "pure" lithologies have been included.
It can be seen from Figure 6.25 that LOLTP gradients of sands in the southern North Sea are generally lower than in shales. However, as there are few tests conducted in sands, a separate plot of the sand and shale trends has not been included. Although the tests performed in the salts and anhydrites are largely within a restricted depth range, such that the trends in these datasets do not stand out, it appears that the average gradients for these lithologies do not change noticeably with depth.

It has been shown in section 6.2.2, that in at least some parts of the southern North Sea, the LOLTPs seem to reflect a stress drop below the Zechstein evaporites. Simple trends in the LOLTP gradients with depth, for the whole dataset, or for individual lithologies, would therefore not be expected. For example, the LOLTP gradients in the sands below 8000 ft can be seen from Figure 6.25 to be generally lower than those above 8000 ft. Trends in the other lithologies are less clear, either because there are not enough data points, or because they are bunched in depth range.

Although there is more scatter in the data, it seems that in the southern North Sea, as in the rest of the North Sea, LOLTP gradients in sands are generally lower than in shales. However, the trends in the southern North Sea are relatively poorly defined due to the small number of data points and are further complicated by an apparent stress drop below the Zechstein evaporites.

**Lithologies in which Lost Circulation Occurs**

The lost circulation data has been treated separately because (as discussed above) the cause of lost circulation is generally not known. Figure 6.26 shows all the lost circulation pressure for the northern, central and southern North Sea. There is a lot of scatter in the data, and trends with depth for any of the lithologies are not clear. This probably reflects the fact that some cases of lost circulation occur in highly permeable zones and some occur due to accidental fracturing of the formation (e.g. due to pistonnage while running the drillpipe in hole). The majority of cases of lost circulation occur in sands. This may be because some sands have very high permeabilities, or it may be because (as implied above by the generally low LOP gradients in sands, Figure 6.24) \( \sigma_n \) is low in these formations and so the formations are being fractured.
Many of the cases of lost circulation which occur below about 7000 ft, are at pressures which are comparable to the LOLTPs displayed in Figures 6.18 to 6.25. The mud weight is generally only raised to these sorts of pressures when over pressured formations are being drilled. In such circumstances, it can be very difficult for the drillers to gauge the correct mud weight in order to control the well but not cause fracturing (see section 4.6) and as a consequence, lost circulation due to fracturing is not uncommon. These cases of high lost circulation pressure therefore probably do represent cases of formation fracturing.

Although it is interesting to speculate as to the causes of lost circulation, the inherent ambiguities prevent the data being used further in a quantitative way in this study.

6.3.2.3 The Effect of Formation Pore Pressure on Leak-off and Limit Test Pressures

The effect of formation pore pressure on the LOPs and LTPs is examined in this section. The formation pore pressures can either be measured or estimated from the mud weight used during drilling. If the formation pore pressure is hydrostatic, it is said to be normally pressured. If pore pressure is above hydrostatic, the formation is said to be over pressured. If the pore pressure is below hydrostatic, the formation is said to be under pressured.

Formation Pore Pressure Measurements

The Erico Pressure Studies contain formation pore pressure measurements where they have been made (see section 5.2) and are contained in the released well data. Such measurements tend to be over small depth ranges, as formation pore pressures are not generally measured throughout the entire depth of oilwells. By integrating the pore pressure measurements and the leak-off test measurements into a relational database (Paradox), it has been possible to select the subset of leak-off tests which fall within sections where pore pressure was measured. The depths of the leak-off test and the pore pressure test are never exactly the same, so interpolation between two pore pressure measurements (one above and one below the leak-off test) has been performed to estimate the pore pressure at the depth of the leak-off test. For the purposes of the
interpolation, a linear change in pore pressure is assumed between the two points of measurement. In almost all cases, pore pressure measurements are made at intervals of a few tens of feet or less, so interpolation does not generally introduce any significant error.

In the central and northern North Sea, there are 125 LOLTPs which fall within intervals which have been tested for formation pore pressure. Many of these intervals have pore pressures above hydrostatic. This may reflect a sampling bias, in that zones which are suspected of over pressure will be more likely to be tested, and where over pressured zones exist, the length of the section tested is likely to be longer. The proportion of normally pressured to over pressured zones for the basin as a whole is probably more accurately reflected by the mud weight data presented below.

The effect of over pressuring on the LOLTPs is examined using the following method which is based on that employed by Breckels and van Eekelen (1981):

1. The LOP or LTP at the depth in question is extracted. The pore pressure at this depth in the same well is interpolated from the pore pressure measurements above and below.

2. From the trend of the whole dataset of LOLTPs with depth for the central and northern North Sea (taken, for example, from a line of best fit equation) the difference (ΔLOLTP) between the LOLTP measured at the depth of interest, and the average LOLTP at that depth is calculated (ΔLOLTP = LOLTP measured - LOLTP average at that depth).

   The Difference (Δp) between the measured pore pressure and the hydrostatic pore pressure at the depth of interest is also calculated (Δp = p measured - p hydrostatic at that depth).

3. ΔLOLTP and Δp are then normalised by depth. The normalised ΔLOLTP (ΔLOLTP (norm.) = ΔLOLTP/depth), is plotted against normalised Δp (ΔP(norm.) = ΔP/depth).

   The units of both ΔLOLTP (norm.) and ΔP(norm.) are psi/ft.

Figure 6.27 shows ΔLOLTP (norm.) against ΔP(norm.) for the cases in the central and northern North Sea where leak-off tests fall within pressure tested intervals. The vertical
line ΔP(norm.) = 0 psi/ft defines the term normally pressured and separates the region of overpressure (ΔP(norm.) > 0) from the region of under pressure (ΔP(norm.) < 0). It can be seen that there are no formations tested which are significantly under pressured.

Points which plot along the line (ΔLOLTP = 0 psi/ft) have a LOLTP equal to the average LOLTP at that depth from the whole central and northern North Sea dataset.

Points which plot above ΔLOLTP = 0 psi/ft have LOLTPs above average, and those that plot below ΔLOLTP = 0 psi/ft have LOLTPs below average. A linear best fit trend shows clearly that there is a positive correlation between degree of over pressure and the magnitude of the LOLTP, i.e. LOLTPs are higher in over pressured formations. This observation is discussed further in Chapter 7.
Figure 6.27 Data from Erico Pressure Atlas, central and northern North Sea. The degree of overpressure (normalised by depth), $\Delta P$ norm., against the degree by which the LOLTP is above average (normalised by depth) $\Delta$LOLTP norm. The linear best fit shows the trend for LOLTPs from overpressured formations to be higher than those from normally pressured formations.
Formation Pore Pressures estimates from Mud Weights.

The procedure of drilling oilwells, and the use of drilling mud of a density (the mud "weight") greater than that of the formation pore fluids in order to control the well, has been described in section 4.6.1. When over pressured formations are drilled, the mud weight must be increased to prevent the formation pore fluid entering the well. Records of mud weights during drilling can therefore be used to identify over pressured sections.

Records of the mud weights used during drilling are available throughout the well, for all the wells in the Erico Pressure Studies. The mud weight used at the start of drilling is generally around 9 to 10 pounds per gallon (ppg) which is equivalent to a specific gravity of 1.08 to 1.2. The density of formation pore fluid with specific gravity of 1.05 (which can be taken as a typical value for formation fluid) is 8.76 ppg. The pressure in the well, due to the hydrostatic weight of the drilling mud is therefore greater than the pressure in the formation assuming the formation is normally pressured. In wells which have no over pressured sections, the mud weight is kept at around 9 to 10 ppg until the total depth is reached. In wells which are over pressured, the mud weight is increased, usually incrementally, up to densities maximum densities of around 18 ppg. A mud weight of 18 ppg causes a hydrostatic pressure in the well close to the weight of the overlying rock:

\[ \text{pressure (psi)} = \text{mud weight (ppg)} \times \text{depth (ft)} \times 0.0519 \]

Where leak-off tests have been performed, the mud weights used to drill those sections have been obtained from the Erico Pressure Studies. This data is analysed in a similar way to the pore pressure data in Figure 6.27:

1. The mud weight used to drill the section in which the leak-off test was performed is converted (using the above relationship) to the equivalent pressure (\(p_{mw}\)) at the depth of leak-off. The difference (\(\Delta p_{mw}\)) between this pressure and hydrostatic pressure of pore fluid is normalised by the depth of the leak-off (TVDSS):

\[ \Delta p_{mw} \text{ (norm.)} = (p_{mw} - \text{Hydrostatic pressure})/\text{Depth}. \]
2. ΔLOLTP (norm.) is calculated in the way described above for formation pore pressure measurements.

3. ΔLOLTP (norm.) is plotted against Δp<sub>mw</sub> (norm.) for all the LOLTPs for the central and northern North Sea (Figure 6.28).

Points which plot in Figure 6.28 on the line Δp<sub>mw</sub> (norm.) = 0 psi/ft represent cases where mud weights used produce a pressure at the bottom of the well equal to hydrostatic pore pressure. As would be expected from a consideration of well control (see section 4.6.1) it can be seen that there are very few such points, but very many that plot just above.

The linear best fit trend in Figure 6.28 has been added by the plotting program. It can be seen that there is a positive correlation between ΔLOLTP (norm.) and Δp<sub>mw</sub> (norm.). This shows that generally, the higher the mud weight used to drill a section of a well, the higher the LOLTP. The data presented in Figure 6.28 implies that LOLTPs are higher in formations with higher formation pore pressures.

It can be seen from the line of best fit in Figure 6.28 that the average value of ΔLOLTP (norm.) is above 0 psi/ft for values of Δp<sub>mw</sub> (norm.) greater than approximately 0.2 psi/ft. On this basis, the LOLTP dataset is divided into two.

- LOLTPs from sections of well that have Δp<sub>mw</sub> (norm.) less than 0.2 psi/ft are here classified as being from essentially normally pressured formations.

- LOLTPs from sections of well that have Δp<sub>mw</sub> (norm.) greater than 0.2 psi/ft are here classified as being from significantly over pressured formations.

Figure 6.29 shows LOLTP gradients (plotted against depth TVDSS) from both normally pressured and over pressured formations. As seen already in Figure 6.28, it is clear from Figure 6.29 that LOLTPs are higher in over pressured formations. Figure 6.29 also enables the distribution with depth of the over pressured formations to be examined.
Figure 6.28 Data from Erico Pressure Atlas, central and northern North Sea. The degree of overpressure (normalised by depth) inferred from the mud weight pressure used to drill ($\Delta p_{mw}$ norm.) against the degree by which the LOLTP is above average (normalised by depth) $\Delta$LOLTP norm. The linear best fit shows the trend for LOLTPs from formations inferred to be overpressured to be higher than those from normally pressured formations.
Figure 6.29 Data from Erico Pressure Atlas, central and northern North Sea. LOLTPs from “overpressured” formations plotted with LOLTPs from normally pressured formations. Linear best fits (Table 6.5) have been added to highlight the comparative trends of the datasets.
It can be seen that the proportion of tests in over pressured formations to normally pressured formations increases with depth. The very high LOLTP gradients at depths around 14,000 ft and below are almost all over pressured.

Although the mud weight used to drill a section of well is only an indication of the formation pore pressure in the well, it has been shown that when the dataset is studied as a whole, a relationship between pore pressure (implied by the mud weight used) and LOLTP is apparent. This correlation is in very good agreement with the correlation seen within the smaller dataset (Figure 6.27) between actual formation pore pressure measurements and LOLTPs. These observations are discussed further in Chapter 7.

6.4 Variation of North Sea Leak-off Pressures, \( \sigma_h \) and \( \sigma_v \) from Amerada Hess Data.

The dataset obtained directly from Amerada Hess constitutes a smaller but more detailed dataset than the Erico data (see section 5.4). It differs from the data in the Erico Pressure Studies in that actual leak-off test records are available (see sections 4.6.3 and 5.3). The leak-off pressure/volume records have been studied carefully and on the basis of the conclusions drawn in section 6.2.3.3 values of \( \sigma_h \) have been determined. Density and sonic logs, are also available and have been used to calculate the vertical stress (see sections 4.7 and 5.4).

The Amerada Hess dataset is firstly presented as LOLTP as has been done for the Erico data. The dataset is then presented in terms of estimated values of \( \sigma_h \) from the pressure/volume records.

6.4.1 Leak-off and Limit Test Pressures in Amerada Hess Dataset.

The availability of the leak-off test records has, for most of the tests, enabled the data to be divided into LO and LT pressures by examination of the pressure/volume plots. Where the pressure/volume plot is not available in the drilling record, and if it is not specifically stated that leak-off occurred, the test is assumed to be a limit test. Figure 6.30 shows the general trend of LOP and LTP gradients with depth for all the Amerada Hess data.
Figure 6.30 Amerada Hess LOP and LTP gradients from the North Sea. Linear best fits (Table 6.5) have been added to highlight the comparative trends of the datasets.
As in the data from the Erico Pressure Atlas, the LOLTP gradients are seen to increase with depth, and there is very little difference between the trend of the LOPs and LTPs with depth. The gradients and intercepts of the best fit lines have been listed in Table 6.5 from which it can be seen that the trends of the Amerada Hess LOPs and LTPs are slightly higher than but similar to the trends of the larger Erico dataset.

As shown in Table 5.6 (with reference to Figure 5.1) the Amerada Hess leak-off test data comes from the West of Shetland Basin, the Viking Graben, the central North Sea and the southern North Sea. The majority of the data however, comes from the central North Sea, and in particular blocks 15/21 and 15/22 which lie on the edge of the Witch Ground Graben. Figure 6.31 shows the Amerada Hess LOLTP data divided by geographic domain, where the geographic domains (central, northern and southern) are those set out for the Erico data in section 5.3. The statistical attributes of the datasets are listed in Table 6.5. Although the Amerada Hess datasets are much smaller, the trends of the LOLTP gradients for these geographic domains are again, slightly higher but very similar to those seen for the Erico data (Figures 6.18 and 6.19). As with the Erico data, the trend of the central and northern North Sea LOLTPs are very close, both increasing with depth. The trend of the southern North Sea data is distinct from the central and northern North sea in that the LOLTP gradients at shallow depths in the southern North sea are higher than the those in the central and northern North Sea, but do not increase with depth.
Figure 6.3: America Hess LOLT Gradient from the North Sea displayed by geographical domain. Linear best fits (Table 6.3) have been added to highlight the comparative trends of the datasets.

- LOLT Gradient (Central N Sea)
- LOLT Gradient (Northern N Sea)
- LOLT Gradient (Southern N Sea)
6.4.2 Interpreted $\sigma_h$ from the Amerada Hess Dataset

Values of $\sigma_h$ have been interpreted from the pressure/volume records of the Amerada Hess leak-off test reports in line with the conclusions drawn in section 6.2.3.3. The pressure volume records have been carefully examined and those which show the characteristics of fracture re-opening have been selected. The LOPs taken from these records are equated with $\sigma_h + 1.8$ MPa. The value of 1.8 MPa has been taken as it is the average difference between 1st LOP and $\sigma_h$ in the Nirex leak-off tests conducted in the sandstones. The leak-off tests analysed here from the Amerada Hess dataset have been performed in a variety of lithologies, including sandstone, shale, chalk and siltstone/claystone. It is acknowledged that the tensile strength will vary between lithologies and that it may also increase with compaction and thus be higher at depth. However, it is assumed that the variation in tensile strength between these lithologies and with depth will be very small compared to the values of $\sigma_h$, and thus an average value of 1.8 MPa to account for the resistance to fracturing during the leak-off test is sufficient. It is assumed that these leak-off tests have opened fractures perpendicular to $\sigma_h$, which is thought to be a reasonable assumption in rocks with low intrinsic tensile strength and in tests where a large surface area of borehole is pressurised such that numerous permeable cracks are likely to be encountered.

For ease of comparison with the other leak-off test data, these estimates of $\sigma_h$ are presented in the same way as the leak-off test data so far in sections 6.3 and 6.4, i.e. the values of $\sigma_h$ are presented as $\sigma_h$ gradients, where the $\sigma_h$ gradient is the secant gradient (section 6.3.2) defined as $\sigma_h$ (psi)/depth(ft TVDSS). The process of selecting the leak-off test records naturally reduces the dataset substantially. There are 49 tests where pressure/volume plots have been confidently interpreted as showing fracture opening characteristics. These are tests where the pressure/volume plot shows an unambiguous pressure build up followed by a shape which gradually rolls over such as shown in Figure 4.16, plots (a) - (d). Any plots which show characteristics similar to plot (e) or (f) in Figure 4.16 have been dropped.
Figure 6.32 Values of $\sigma_h$ (Shmin) gradients ($\sigma_h$/depth or Shmin/depth) determined from Amerada Hess leak-off test pressure/volume plots, and displayed by geographic domain. Linear best fits (Table 6.5) have been added to highlight the comparative trends of the datasets.
The interpreted values of $\sigma_h$ (called Shmin in Figure 6.32 and all subsequent figures displaying $\sigma_h$) are displayed in Figure 6.32 in which the data have been divided into geographic domain. The gradients and intercepts of the lines of best fit have been listed in Table 6.5. Clearly, there is a lack of data from the southern North Sea, such that negative gradient of the trend line is probably misleading.

The general trends of $\sigma_h$ with depth are, as would be expected, clearly similar to those of the LOLTP trends presented above for both the Erico dataset and the Amerada Hess dataset. The contrast between the southern and northern/central north sea is still very apparent. The most noticeable difference in this data set is that the intercepts of the trend lines with the depth = 0 line are much lower than the LOLTP data. This is largely due to the fact that the deduction of the tensile strength term lowers the stress gradient by a greater proportion at shallow depths. As a consequence of the relatively lowering of the values at shallow depths, the gradients of the trend lines are steeper.

6.4.3 Leak-off Pressures, Limit Test Pressures and $\sigma_v$ in Amerada Hess Dataset.
As it has been possible to calculate the vertical stress throughout each well for the Amerada Hess wells (see sections 4.7 and 5.4), it is now possible to examine how the vertical stress ($\sigma_v$) varies with depth, and also how the LOLTPs vary with vertical stress. The vertical stress could be calculated at any point in the well, but for the purposes of this leak-off test study, the vertical stress at each point of leak-off is presented here. Figure 6.33 shows how the $\sigma_v$ gradient varies with depth. The $\sigma_v$ gradient is defined in the same way as the LOP and LTP gradients: $\sigma_v$ gradient = $\sigma_v$ (psi) / depth (ft TVDSS). The data in Figure 6.33 have been grouped by geographic domain. The trend of the vertical stress gradient differs between the southern North Sea and the rest of the North Sea and correlates very well with the LOLTP data. The trends of the two datasets can be compared quantitatively with reference to the data listed in Table 6.5.

The northern and central North Sea basins contain thick sequences of Cenozoic sediments. Density log measurements of these sediments show that they have low densities at the surface, which increase quite rapidly with depth, due to compaction. Therefore, in the central and northern North Sea the vertical stress gradient increases.
quite rapidly with depth. In the southern North Sea, the sea floor is commonly composed of Mesozoic rocks which have been buried, compacted and then brought back to the surface (see section 2.3.7.3). These rocks therefore have relatively high densities at the surface, which do not increase rapidly (compared to uncompacted rocks) with depth, and so do not cause rapid increases in vertical stress gradient with depth.

It is also apparent from Figure 6.33 that the trends of the $\sigma_v$ gradients reflect the trends shown by the lines of best fit for the LOLTP gradients shown in Figure 6.31 and the $\sigma_h$ trends shown in Figure 6.32.

The relationship between the LOLTPs and $\sigma_v$ is also displayed in figure 6.34, where the ratio of the LOLTPs to $\sigma_v$ is plotted against depth. As there are relatively few points in the datasets for the northern and southern North Sea, these trends are probably not so reliable as the trends for the central North Sea. However, it is clear that at least in the central North Sea, the LOLTPs and the $\sigma_h$ values, taken as a whole, appear to be proportional to $\sigma_v$, with a LOLTP/$\sigma_v$ ratio of approximately 0.95.

Figure 6.35 show the ratio of $\sigma_v$/Lo plotted against depth. Again, the trend for the southern and perhaps the central North Sea should not be over interpreted as they are based on very few data points. The trend for the central North Sea however, is quite different to that of the LOLTP/$\sigma_v$ trend for the central North Sea. The LOLTP trend indicates that the ratio LOLTP/$\sigma_v$ is constant with depth, which would imply, if the LOLTP were simply assumed equal to $\sigma_h$, that $\sigma_h$ is due entirely to the overburden. It is clear however, from Figure 6.35, that the ratio of $\sigma_v$/Lo is not constant, but actually increases with depth. Thus, although the increase in $\sigma_v$ gradient with depth is a significant factor in the increase of $\sigma_h$ and LOLTP gradient, it is not entirely responsible, and it must be that some other factor also acts to increase $\sigma_h$ with depth. The implications of this observation, on the origin of $\sigma_h$ in the North Sea are discussed in the next chapter. The statistical attributes of the datasets presented in Figures 6.34 and 6.35 are summarised in Table 6.5.
Figure 6.33 Values of $\sigma_v$ gradients ($\sigma_v$/depth) calculated at the points of the Amerada Hess leak-off tests and displayed by geographic domain. Linear best fits (Table 6.5) have been added to highlight the comparative trends of the datasets.
Figure 6.34 The ratio of LOLTP/σ, from the Amerada Hess data displayed by geographic domain. Linear best fits (Table 6.5) have been added to highlight the comparative trends of the datasets.
Figure 6.35 The ratio of $\sigma_p/\sigma_v$ from the Amerada Hess dataset displayed by geographic domain. Linear best fits (Table 6.5) have been added to highlight the comparative trends of the datasets.
What is clear from the comparison between Figures 6.31 and 6.32 and the comparison between 6.34 and 6.35, is that, of course, the values of $\sigma_h$ selected from the pressure/volume plots and corrected for tensile strength show lower values than the LOLTPs, but this has had a considerable effect on the values taken from the shallower tests. Further, the amount of scatter in the values taken from the shallow tests is greatly reduced by using carefully selected LOPs corrected for tensile strength. This is particularly noticeable in the comparison between Figures 6.34 and 6.35.

**6.4.4 Lithology and Formation Pore Pressure in Amerada Hess Dataset.**

Information on the lithologies in which the tests were performed is available from a combination of the drilling records, final well reports and the and the well logs (see section 5.4). As with the data contained in the Erico Pressure Studies, the majority of tests are performed in shales. There are therefore only a few tests for each of the other lithologies, which do not constitute datasets large enough to examine variations in trend with lithology. The Amerada Hess LOLTP gradients, grouped by lithology, are plotted against depth in Figure 6.36. The LOLTP gradients have been used because of the small size of $\sigma_h$ dataset. Even so, because of the lack of data points, only one line of best fit has been included. This is the line of best fit for the data from tests in shales. The gradient and intercept of this line are included in Table 6.5. In common with other comparisons between the Amerada dataset and the larger Erico dataset, the trend of the Amerada data is very similar but slightly higher than the Erico data.

It can be seen from Figure 6.36 that several of the tests conducted in sands plot higher than the average shale value. This appears to be contrary to the lithological contrast displayed by the Erico data. However, when the pore pressures in the Amerada dataset are examined (Figure 6.37) it is seen that several of these high values in sands are from overpressured formations.

Formation pore pressure information is available from final well reports for most of the wells in the Amerada Hess dataset. Only a few of the wells show any sign of overpressured formations. The dataset of over pressures is not large enough to examine the trends in the LOLTP with formation pressure.
Figure 6.30: Amerada Hess LOLTP gradients displayed by lithology.
Figure 6.37 American Hess LOLTP gradients from tests conducted in normally pressured and overpressured formations.
The Amerada Hess LOLTP gradients, grouped by overpressured and under pressured formations, are plotted against depth in Figure 6.37. The separation of data points from overpressured and normally pressured formations is not as clear as for the Erico dataset, although it can be seen that the data from over pressured formations generally plot on the high side of the data.

6.5 Summary

Comparison of LOPs, from a variety of rock types, with hydro-frac derived $\sigma_h$ values shows that the 1st LOP can significantly overestimate $\sigma_h$. LOPs from 2nd leak-off test pressurisation cycles provide better estimates of $\sigma_h$ than LOPs from 1st pressurisation cycles. However, comparison of 1st LOPs from tests performed in sedimentary rocks (where it is assumed that permeable cracks oriented perpendicular to $\sigma_h$ exist, and the leak-off test pressure/volume plot indicates that such fractures have opened along their length) with hydro-frac derived $\sigma_h$ values has shown that such LOPs are only marginally higher than $\sigma_h$. An empirical relationship has been proposed to relate the LOP to $\sigma_h$ plus some measure of the rocks resistance to fracturing which, in the absence of other data, is a function, $\alpha_T$, (where $0 < \alpha_T < 1$) of the tensile strength of the rock: $1st \text{LOP} = \sigma_h + \alpha_T T$. This relationship holds for the data from the southern North Sea, where leak-off tests have been performed in siltstones.

Where leak-off test pressure volume plots are not available, the single leak-off pressure cannot be used as an estimate of $\sigma_h$. However, as LOPs are clearly closely related to $\sigma_h$, it is considered that trends of LOP with depth do indeed reflect trends of $\sigma_h$ with depth. LOPs for which pressure/volume plots are not available have therefore been presented, generally in the form of LOP and LOLTP gradients against depth to investigate how these trends vary with depth, geographic domain, lithology and formation pore pressure (e.g. Figures 6.18 - 6.31).

Where leak-off test pressure/volume plots are available and the rock types are known, such as for the Amerada Hess dataset, estimates of $\sigma_h$ have been made using the relationship given above and reasonable estimates of the rock tensile strength (Figure 6.32). These values have also been compared to the calculated vertical stress.
In the next chapter, based on the conclusions of how best to use leak-off test data outlined in section 6.2.3.3, trends of $\sigma_b$ with depth are defined for the North Sea. Based on these trends, and the results presented in this chapter, the distribution and origin of stress in the North Sea is discussed.
<table>
<thead>
<tr>
<th>Dataset</th>
<th>Gradient of best fit (psi/ft²)</th>
<th>Intercept with depth = 0 (psi/ft)</th>
<th>Mean Pressure Gradient</th>
<th>Stnrd. Dev. of Pressure Gradients</th>
</tr>
</thead>
<tbody>
<tr>
<td>LOP Gradients</td>
<td>$1.43 \times 10^{-3}$</td>
<td>0.701</td>
<td>0.786</td>
<td>0.1055</td>
</tr>
<tr>
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<td>$1.2 \times 10^{-3}$</td>
<td>0.707</td>
<td>0.778</td>
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<td>LCP Gradients</td>
<td>$2.07 \times 10^{-3}$</td>
<td>0.488</td>
<td>0.641</td>
<td>0.1376</td>
</tr>
<tr>
<td>Figure 6.20</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>LOLTP Gradients (Northern)</td>
<td>$1.51 \times 10^{-3}$</td>
<td>0.657</td>
<td>0.753</td>
<td>0.0920</td>
</tr>
<tr>
<td>LOLTP Gradients (Central)</td>
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<td>0.678</td>
<td>0.781</td>
<td>0.1035</td>
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<tr>
<td>LOLTP Gradients (Southern)</td>
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<td>0.827</td>
<td>0.829</td>
<td>0.1175</td>
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<tr>
<td>LOP Gradients (Northern)</td>
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<td>0.764</td>
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<tr>
<td>LOP Gradients (Central)</td>
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<tr>
<td>LOLTP Gradients (Shale)</td>
<td>$1.58 \times 10^{-3}$</td>
<td>0.678</td>
<td>0.767</td>
<td>0.0940</td>
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<td>LOLTP Gradients (Silt)</td>
<td>$2.39 \times 10^{-3}$</td>
<td>0.616</td>
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<td>0.640</td>
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<td>0.744</td>
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<tr>
<td>LOLTP Gradients (Shale)</td>
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<td>0.683</td>
<td>0.768</td>
<td>0.0928</td>
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<tr>
<td>LOLTP Gradients (Sand)</td>
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<td>LOLTP Gradients (norm p)</td>
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<td>0.753</td>
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<td>LOLTP Gradients (over p)</td>
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<td>0.0774</td>
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<tr>
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<td>0.1076</td>
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<tr>
<td>Figure 6.31</td>
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<tr>
<td>LOLTP Gradients (Northern)</td>
<td>$1.47 \times 10^{-3}$</td>
<td>0.700</td>
<td>0.763</td>
<td>0.0850</td>
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<tr>
<td>LOLTP Gradients (Central)</td>
<td>$1.67 \times 10^{-3}$</td>
<td>0.706</td>
<td>0.800</td>
<td>0.0994</td>
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<tr>
<td>LOLTP Gradients (Southern)</td>
<td>$3 \times 10^{-7}$</td>
<td>0.916</td>
<td>0.917</td>
<td>0.1414</td>
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<tr>
<td>σ/depth (Northern)</td>
<td>$2.6 \times 10^{-3}$</td>
<td>0.558</td>
<td>0.713</td>
<td>0.1123</td>
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<tr>
<td>σ/depth (Central)</td>
<td>$3 \times 10^{-3}$</td>
<td>0.560</td>
<td>0.681</td>
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<td>σ/depth (Southern)</td>
<td>$8 \times 10^{-3}$</td>
<td>0.925</td>
<td>0.866</td>
<td>0.0624</td>
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<tr>
<td>LOLTP Gradients (Northern)</td>
<td>$2.11 \times 10^{-2}$</td>
<td>0.739</td>
<td>0.830</td>
<td>0.0604</td>
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<td>LOLTP Gradients (Central)</td>
<td>$1.8 \times 10^{-3}$</td>
<td>0.754</td>
<td>0.856</td>
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<td>LOLTP Gradients (Southern)</td>
<td>$7.7 \times 10^{-8}$</td>
<td>0.976</td>
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<td>0.0444</td>
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<td>Figure 6.34</td>
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<td></td>
</tr>
<tr>
<td>LOP/LP (Northern)</td>
<td>$-5.4 \times 10^{-4}$</td>
<td>0.942</td>
<td>0.919</td>
<td>0.0761</td>
</tr>
<tr>
<td>LOP/LP (Northern)</td>
<td>$6 \times 10^{-7}$</td>
<td>0.939</td>
<td>0.935</td>
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</tr>
<tr>
<td>LOLTP/L (Northern)</td>
<td>$-6.4 \times 10^{-4}$</td>
<td>0.908</td>
<td>0.878</td>
<td>0.1353</td>
</tr>
<tr>
<td>Figure 6.35</td>
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</tr>
<tr>
<td>σ/μ (Northern)</td>
<td>$2.2 \times 10^{-3}$</td>
<td>0.732</td>
<td>0.827</td>
<td>0.0954</td>
</tr>
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<td>σ/μ (Central)</td>
<td>$1.8 \times 10^{-3}$</td>
<td>0.761</td>
<td>0.851</td>
<td>0.0890</td>
</tr>
<tr>
<td>σ/μ (Southern)</td>
<td>$-1.0 \times 10^{-3}$</td>
<td>0.905</td>
<td>0.832</td>
<td>0.0665</td>
</tr>
<tr>
<td>Figure 6.36</td>
<td></td>
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<td></td>
<td></td>
</tr>
<tr>
<td>LOLTP Gradients (Shale)</td>
<td>$1.23 \times 10^{-3}$</td>
<td>0.731</td>
<td>0.793</td>
<td>0.0762</td>
</tr>
</tbody>
</table>

Table 6.5. Gradients and intercepts of the trend lines with depth shown in various figures in this chapter. Also, means and standard deviations of these datasets.
Chapter 7 - Interpretation of Leak-off Test Results and Discussion of Stress Magnitudes in the North Sea

7.1 Introduction

Trends of $\sigma_h$ with depth are established from the leak-off test data in this section. The distribution and origin of stress in the North Sea is investigated from general observations and in the context of models of stress which have been proposed for sedimentary basins. This chapter also includes an investigation into a possible method of constraining the magnitude of $\sigma_H$ by using a combination of borehole breakouts and leak-off test data. Finally, the trend of $\sigma_h$ with depth in the North Sea is compared with trends of $\sigma_h$ with depth established for other areas around the world. The similarities and differences between these trends are discussed. The units used in this chapter are SI units (MPa and m) to allow direct comparison with other data from around the world, most of which does not come directly from the oil industry.

7.2 Trends of $\sigma_h$ with Depth in the North Sea.

In view of the conclusions outlined in section 6.2.3.3, it is considered that there are two possible ways of defining the trend with depth of $\sigma_h$ using the leak-off test data available. These methods are discussed in Sections 7.2.1 and 7.2.2. In section 7.2.3, the value of $\sigma_h$ at the sea floor, and the effect of the sea water on the trends of $\sigma_h$ with depth are discussed. The various possible trends with depth of $\sigma_h$ in the North Sea are compared and discussed in section 7.2.4 and the most suitable trend for comparison with other regions of the world is selected.

7.2.1 Trend of $\sigma_h$ with Depth Derived from Amerada Hess Pressure/Volume Plots.

Values of minimum horizontal stress ($\sigma_h$) have been estimated using the leak-off test pressure/volume plots from the Amerada Hess drilling records. This dataset is rather small, with 38 $\sigma_h$ values from the central North Sea, 7 from the northern North Sea (including west of Shetland data), and only 3 data points from the southern North Sea. It is therefore considered that only in the central North Sea is there enough data to define a reliable trend with depth.
Figure 7.1 presents the $\sigma_h$ values from the central North Sea. To avoid ambiguity and to establish a trend which can readily be compared to other regions, five estimates of $\sigma_h$ made in zones known to be overpressured have been omitted from this plot.

On the basis of lithological descriptions in leak-off test reports, final well reports and gamma ray logs, it has been concluded (section 5.4) that 30 of the Amerada Hess LOPs are from tests conducted in shales, 5 are from tests conducted in lithologies described in the leak-off test records as claystones/siltstones and 3 are from tests where no lithological information is available. The latter test results have been included to prevent a further reduction in the size of the dataset (a similar plot excluding this data shows no significant difference). From the data presented in Figures 6.22 - 6.24, it is expected that there will be different trends of $\sigma_h$ with depth for different lithologies. Figure 7.1 therefore, is perhaps best described as representing an average trend of $\sigma_h$ with depth in the central North Sea for normally pressured shales and claystones/siltstones.

The plotting program (Microsoft Excel 5.0) can fit a variety of types of line to a dataset such as that shown in Figure 7.1 using curve fitting routines. The types of line which can be fitted to such a dataset include: polynomial curves of order 2 to 6 (e.g. $y = a.x^2 + b.x + c$), exponential curves (e.g. $y = a.e^{bx}$), power law curves (e.g. $y = a.x^b$) and straight lines. The value of the correlation coefficient ($R^2$), which is a measure of how well the chosen line fits the data is also calculated by the plotting program. An $R^2$ value of 1 denotes a perfect correlation of the line to the dataset, whereas an $R^2$ value of 0 denotes no correlation. Both power law and polynomial curves give very good fits to the dataset shown in Figure 7.1, with $R^2$ values only varying between 0.986 and 0.987. The choice (between power law and polynomial) of line to describe the trend of $\sigma_h$ with depth is therefore somewhat arbitrary, as the type of relationship used to describe the trend is not, at this stage, intended to imply anything about the origin of $\sigma_h$, but merely used to represent the empirical relationship of $\sigma_h$ with depth.
Figure 7.1 The trend of $\sigma_h$ with depth defined by values determined from Amerada Hess leak-off test pressure/volume plots. Results from tests in overpressured formations have been excluded. 30 of the remaining results are from tests in shales, 5 from tests in claystone/siltstone and 3 from tests with no lithological information.
The line of best fit shown in Figure 7.1 is the 2nd order polynomial fit listed in Table 7.1. The $R^2$ correlation factor for this line is 0.9867, which is marginally higher than that of the power law fit. Higher order polynomials do not yield significantly higher $R^2$ values. The possible causes of this type of trend with depth of $\sigma_h$ are discussed later in section 7.3.

<table>
<thead>
<tr>
<th>Dataset</th>
<th>Type of fit</th>
<th>Trend of $\sigma_h$ with depth($z$) (equation of best fit)</th>
</tr>
</thead>
<tbody>
<tr>
<td>A.Hess, $\sigma_h$ from leak-off test records, central North Sea. Fig. 7.1</td>
<td>Best fit to whole dataset</td>
<td>$\sigma_h = 2.02 \times 10^{-6}z^2 + 0.01346z$</td>
</tr>
<tr>
<td>Erico, central North Sea Fig. 7.2</td>
<td>Best fit to hand picked lower bound</td>
<td>$\sigma_h = 2.28 \times 10^{-6}z^2 + 0.01064z$</td>
</tr>
<tr>
<td>Erico, northern North Sea Fig. 7.3</td>
<td>Best fit to hand picked lower bound</td>
<td>$\sigma_h = 1.35 \times 10^{-6}z^2 + 0.01252z$</td>
</tr>
<tr>
<td>Erico, Central North Sea (shales only) Fig. 7.4</td>
<td>Best fit to hand picked lower bound</td>
<td>$\sigma_h = 2.11 \times 10^{-6}z^2 + 0.01233z$</td>
</tr>
<tr>
<td>Erico, Northern North Sea (shales only) Fig. 7.5</td>
<td>Best fit to hand picked lower bound</td>
<td>$\sigma_h = 1.9 \times 10^{-6}z^2 + 0.01198z$</td>
</tr>
</tbody>
</table>

Table 7.1 Equations of the lines of best fit added to the datasets to define the trend of $\sigma_h$ with depth ($z$).

7.2.2 Trends of $\sigma_h$ with Depth Derived from Erico Leak-off Pressure Data

As discussed in section 6.2.3.2, where pressure/volume plots from leak-off tests are not available (but only single LOP values have been obtained) such as for the data from the Erico Pressure Atlas, these single LOPs cannot confidently be used as estimates of $\sigma_h$. On the other hand however, the trend of these LOPs with depth does appear to reflect the trend of $\sigma_h$ with depth. Further, it is clear in view of the discussion in sections 6.2.3.2, 4.6.5 and 4.6.6, that as LOPs can theoretically have a value anywhere between $\sigma_h$ and $2\sigma_h - p + T$ it will be the lower end of the range of values at any depth which will generally be closest to $\sigma_h$. The approach used here to obtain the trend of $\sigma_h$ with depth from the Erico leak-off test data is similar to that used by Breckels and van Eekelen (1981) who fitted a lower bound curve to leak-off test data from the US Gulf Coast (Figure 4.14).
The trends of $\sigma_b$ with depth for the central and northern North Sea have been estimated from plots of LOP only (i.e. excluding LTP and LCP) against depth. Clearly LTP which, despite their general similarity to LOPs, can be significantly lower than LOPs, cannot be included for this exercise. Because of the complexities introduced by the stress drop seen in the data of Figure 6.15, and the uncertainty in the lateral extent of this stress drop, a trend of $\sigma_b$ with depth for the southern North Sea has not been attempted.

As with the Amerada Hess dataset (Figure 7.1) any LOPs from tests conducted in formations which are suspected to be overpressured have been omitted. Although these LOPs are generally higher than average (Figure 6.29) and so should not affect the lower bound, they have been excluded for the sake of consistency and because of the possibility that a few LOPs from overpressured zones may be close to the lower bound (Figure 6.29) and would thus affect the trend that is fitted. The data defined in section 6.3.2.3 as being from overpressured formations (shown in Figure 6.28 and 6.29) has been excluded from the analyses in this section. The lower bounds from these datasets then, represent the trend with depth of $\sigma_b$ for normally pressured formations in the central and northern North Sea.

Lower bounds to the datasets have been estimated by eye, and drawn in by hand. Equations which describes these lower bounds to the data can be obtained by re-plotting the points which define the line using the plotting program and then adding a line of best fit (section 7.2.1). The lower bounds shown in Figures 7.2 to 7.5 are thus generated by the plotting program using the equation of the line which was drawn by hand.

When fitting the lower bound curve by eye, some subjectivity is involved. The data has a basin wide distribution and so the trend with depth in each borehole will be slightly different. This exercise of producing a trend with depth curve of $\sigma_b$ for the whole of the central or northern North Sea can therefore only give an approximation. Some of the datasets presented in Figures 7.2 to 7.5 are reasonably easy to fit a lower bound to, although there are always a few points which lie below the lower bound defined by the bulk of the data. LOPs which lie conspicuously below the trend defined by the lower bound to the bulk of the dataset can be explained in more than one way. Firstly, as
discussed in section 4.6, LOPs from tests conducted in permeable formations can be ambiguous to define and are often lower than LOPs in less permeable formations. Secondly, the pressure/volume plot may have been misinterpreted by the drilling engineers. Often pressure/volume plots do not show a perfectly linear pressure build-up so that a small pressure fluctuation, caused perhaps by the pump or equipment compliance, may be misinterpreted as leak-off.

In view of the above discussion, it is considered that the lower bounds fitted should be lower bounds to at least 90% of the data. In Figures 7.2, 7.3, 7.4, and 7.5, the lower bounds shown are lower bounds to 98%, 93%, 93% and 94% respectively.

Lower bounds have been fitted in the manner described above to the Erico LOP data from normally pressured formations for the central and northern North Sea. Figures 7.2 and 7.3 show the lower bounds picked for the central and northern North Sea data respectively. The lower bound to the northern North Sea data is less clearly defined than that to the central North Sea data. The lower bound picked for the northern North Sea is thus somewhat more subjective. It should be noted that the data used in these Figures come from a variety of lithologies, and thus, as shown in Figures 6.22 to 6.24, some scatter would be expected.

The functions which represent the best fit to the lower bounds of the datasets in Figures 7.2 and 7.3 are the 2nd order polynomials listed in Table 7.1. As for Figure 7.1, the trends have been set to go through the origin (see discussion below) and at this stage are only intended to describe the empirical relationship between $\sigma_\text{h}$ and depth.

The lower bounds shown in Figures 7.2 and 7.3 represent the trends with depth of $\sigma_\text{h}$ in the central and northern North Sea based on tests performed in a variety of normally pressured formations. As shown in Figure 6.24, LOPs in sands are generally lower than in shales for the same depth. The trends shown in Figures 7.2 and 7.3 are therefore probably close to the trend of $\sigma_\text{h}$ with depth for the low stress bearing lithologies.
Figure 7.2 The trend of $\sigma_b$ with depth defined by the lower bound to the bulk of Erico LOPs from the central North Sea. LOPs from overpressured formations have been excluded.
Figure 7.3 The trend of $\sigma_h$ with depth defined by the lower bound to the bulk of Erico LOPs from the northern North Sea. LOPs from overpressured formations have been excluded.
In order to produce a less ambiguous trend of $\sigma_h$ with depth, the lower bound to single lithologies can be defined. Because of the diminishing size of the dataset as more restrictions are introduced to better define the trends of $\sigma_h$ with depth, only the dataset from test performed in shales is large enough to confidently define a trend.

Figures 7.4 and 7.5 present the LOPs from normally pressured shales in the central and northern North Sea respectively. The lower bounds to these datasets have been added in the way outlined above. The lines of best fit which define the lower bounds presented in these figures are 2nd order polynomials (which have been set to go through the origin) the equations of which are presented in Table 7.1.

A visual comparisons between Figure 7.2 and 7.4 and also between 7.3 and 7.5, clearly show that the trends of $\sigma_h$ defined by the data from shales alone are higher than trends based on the datasets from all lithologies (this will be more clearly illustrated later in section 7.2.4). The difference is particularly obvious in the comparison between Figures 7.3 and 7.5, where it is clear that most of the lower outlying data points in Figure 7.3, which caused some ambiguity in picking a lower limit for that figure, are not present in Figure 7.5 and were thus not from tests performed in shales.

7.2.3 Stress at the Sea Floor and the Effect of Sea Water on the Trend of $\sigma_h$ with Depth.

In figures 7.1 to 7.3, the trends with depth of $\sigma_h$ have been set to go through the origin, i.e. it is assumed that at zero depth the value of $\sigma_h$ is zero. The depths used in Figures 7.1 - 7.3 are vertical depths measured from the sea surface where of course $\sigma_h$ is zero. The shallow rocks of the central and northern North Sea are predominantly recent unconsolidated sediments and it is assumed that they can support no significant effective stress at the sea floor. The total stress at the sea floor is therefore due only to the weight of the overlying sea water and so it seems reasonable to extrapolate the data from the leak-off pressures directly to zero at the sea surface. Due to the presence of the sea water, the first 300m or so of these trends is not strictly accurate. However, the presence of the sea water does not significantly affect the overall trend of $\sigma_h$ with depth.
The depth of water above these wells is between 50 and 170 m for all but approximately 12 wells. The trend of LOP with depth for the few wells where large water depths are encountered (up to 330m) has been investigated and shows only minor variations from the average trend at shallow depths, and no distinguishable difference from the average trend at greater depth. As the trends presented here can only be averages for an area (see above discussion) these small variations in water depth have been ignored.
Figure 7.4 The trend of $\sigma_\text{b}$ with depth defined by the lower bound to the bulk of Erico LOPs from tests in shales in the central North Sea. LOPs from overpressured formations have been excluded.
Figure 7.5 The trend of $\sigma_h$ with depth defined by the lower bound to the bulk of Erico LOPs from tests performed in shales in the northern North Sea. LOPs from overpressured formations have been excluded.
7.2.4 The Trend of $\sigma_h$ with Depth in the North Sea Using Different Methods: Comparison.

It has been seen in section 7.2.2 that it is possible to produce different trends of $\sigma_h$ with depth for the same area depending on which method is used (pressure/volume plot method or lower bound to the whole dataset method) and which lithologies are included in the dataset. Figure 7.6 shows all the trends presented in Figures 7.1 to 7.5 plotted together for comparison.

It can be seen from Figure 7.6 that, as would be expected (section 7.2.2), the trends established from the datasets which include all lithologies are somewhat lower than those established using just shales from the same area.

The fact that the trend established for the central North Sea using data from all lithologies is so close to the trend for the northern North Sea using shales only, is assumed to be simply coincidence.

The trend established for the shales and claystones/siltstones of the central North Sea using the pressure/volume plot method is the highest of the trends presented in Figure 7.6. However, it is actually very close to the trend established for the shales of the same region using the lower bound technique with the Erico data. The maximum difference in the value of $\sigma_h$ predicted by these two trends is just 3 MPa (4.3%) at a depth of 3500m. The fact that these two different methods have been performed on different datasets and yield such similar results is very encouraging.
Figure 7.6 A comparison of the trends presented in Figures 7.1 to 7.5.

- Central (Erico)
- Northern (Erico)
- Central (Amerada)
- Northern (Erico)
- Central (Erico - shales only)
- Northern (Erico - shales only)
In section 7.5 the trend of $G_h$ with depth in the North Sea is compared to other trends of $G_h$ with depth from around the world. For the sake of clarity only one of the trends displayed in Figure 7.6 (that from Figure 7.1), has been picked for this comparison. The trend of $G_h$ in the central North Sea derived from the Amerada Hess data is defined by actual estimates of $G_h$ as opposed to the lower bound to LOPs. Approximately 80% of these data points are from tests performed in shales. The trend defined by the lower bound to the LOPs from shales in the central North Sea is in good agreement with the trend defined by the Amerada Hess data. The trend defined by the actual estimates of $G_h$ (from Amerada Hess pressure/volume plots) is therefore considered the most suitable trend (for the central North Sea) for comparison with other trends from around the world (section 7.5).

7.3 The Distribution and Origin of $G_h$ in the North Sea

Now that both trends of $G_h$ with depth in the North Sea (from both Amerada Hess data and Erico data) and actual values of $G_h$ in the North Sea (from the Amerada Hess pressure/volume plots) have been established, the origins of $G_h$ can be considered.

In section 7.3.1 some general observations about the trends of $G_h$ with depth in the North Sea are discussed. Some of these observations have been touched upon in sections 6.3 and 6.4 where trends in $G_h$ were inferred from the LOLTP data and from the pressure/volume plots of the Amerada Hess data.

In section 7.3.2 the actual values of $G_h$ determined from the Amerada Hess pressure/volume plots and inferred from the trends of $G_h$ with depth derived from the Erico data are investigated in a semi-quantitative way in terms of the models of stress with depth outlined in section 3.4.

The possible causes for the stress drop seen in the southern North Sea are discussed in section 7.3.3. In the final part of this section some conclusions are drawn regarding the general state of stress in the North Sea based on the observations and comparisons with stress models discussed in section 7.3.1 to 7.3.3.
7.3.1 General Observations of the Distribution of $\sigma_h$ in the North Sea

Ideally, if enough data existed, the distribution of $\sigma_h$ would be investigated from leak-off test data using either of the two methods presented in section 7.2 and a discussion of the origins of $\sigma_h$ could be based on the results of these methods. However, pressure/volume plots are not available for the bulk of the data (i.e. the Erico data) and only in the case of the larger datasets such as those presented in Figures 7.2 to 7.5, can the lower bound method be confidently used. However, in the light of the conclusions in section 6.2.3.3, it seems reasonable for the purposes of this general discussion, to infer that, although individual LOLTP values are not necessarily equal to $\sigma_h$, the trends of LOLTPs with depth presented in sections 6.3 and 6.4 do reflect the trends of $\sigma_h$ with depth. Thus it is possible, without rigorously relating LOLTP to $\sigma_h$, to make some general observations about the variation of $\sigma_h$ with depth in the North Sea, and thus to discuss the possible origins of $\sigma_h$, from examination of the LOLTP trends presented in sections 6.3 and 6.4.

In view of the above discussion, the general observations which can be made from examination of the leak-off test data from the North Sea presented in sections 6.3 to 7.2 are based on examination of both “raw” LOLTP data (e.g. Figure 6.18 to 6.31) and interpreted LOP data (e.g. Figures 7.1 to 7.6), and are as follows:

(a) $\sigma_h$ generally (with the exception of the stress drop in the southern North Sea) increases with depth.

(b) The rate of increase of $\sigma_h$ with depth is not constant but increases with depth, i.e. if the gradient of $\sigma_h$ is defined in the same way as the LOLTP gradients (the secant gradient - section 6.3.2), then the $\sigma_h$ gradient increases with depth.

(c) The rate of increase in $\sigma_h$ gradient with depth is similar in the central and northern North Sea, although perhaps slightly higher in the central North Sea (Figures 6.20, 6.21, 6.31, 6.32, 7.6) and is very low in the southern North Sea (Figures 6.20, 6.21, 6.31, 6.32).

(d) The $\sigma_h$ gradients are higher in shales than sands in the central and northern North Sea (Figures 6.23 and 6.24). In the southern North Sea, this lithological stress contrast is not clearly seen.

(e) The $\sigma_h$ gradients are higher in overpressured formations (Figures 6.27 to 6.29)
A general and qualitative discussion of these observations in terms of the origins of $\sigma_h$ now follows.

**Observation (a)**

When a rock is buried, it is subjected to the load of the overburden (that is gravity acting on the mass of overlying rock) thus it is clear that the rock will tend to contract vertically and expand horizontally in response to this vertically applied load. By attempting to expand horizontally, horizontal stresses are generated. With increasing depth, the weight of the overburden increases i.e. the vertical stress increases, and thus the horizontal stress due to the vertical stress also increases. Observation (a) is therefore very easily explained, and similar observations have been made in studies around the world (section 7.5). As discussed below, other factors also exist which could cause horizontal stress to increase with depth.

**Observation (b)**

As sediments are buried they tend to compact due to the increasing confining pressure. With compaction these rocks become more dense and thus give rise to higher vertical stress gradients. It would be expected therefore to see an increase in $\sigma_v$ gradient with depth as long as the rocks are compacting with depth. This is seen in Figure 6.33. It is clear from the above discussion that since $\sigma_h$ is due at least in part to $\sigma_v$, it would be expected that as the $\sigma_v$ gradient increases the gradient of $\sigma_h$ due to the vertical stress would also increase. Observation (b) can therefore be at least partly explained by the effect of compaction. Again, there are other factors, discussed below, which might also be responsible for observation (b).

**Observation (c)**

The grabens of the northern and central North Sea contain thick sequences of Cenozoic sediments whereas the shallow rocks of the southern North Sea consist predominantly of older (generally Mesozoic) rocks. These older rocks have been buried in the past but are now near the surface due to processes such as basin inversion which occurred in the southern North Sea in the Cretaceous and Tertiary (section 2.3.7.3). This tectonic
history is revealed by the density logs which have been run in the shallow sediments of
the southern North Sea. The bulk densities of shallow rocks in the southern North Sea
are typically around 2.3 g/cm$^3$ and increase to around 2.5 g/cm$^3$ at a depth of around
3000m. This contrasts sharply with the density profiles of the central and northern North
Sea grabens, where shallow Cenozoic sediments have densities of around 1.9 - 2.0
g/cm$^3$, which only increase to densities similar to those seen in the southern North Sea at
depths of around 3000m and below. These differences in density profile are reflected in
the values of $\sigma_v$ calculated from density logs (section 4.7) and shown in Figure 6.33. As
the effect of compaction is small in the southern North Sea, the increase with depth in $\sigma_v$
gradient is small and thus the increase in $\sigma_h$ gradient with depth is also small. The very
noticeable difference in $\sigma_h$ gradients between the central/northern and the southern
North Sea that is listed above as part of observation (c) can thus be at least partly
explained by the presence or lack of compaction.

A similar effect could explain the other part of observation (c), that the gradients of $\sigma_h$
are slightly higher in the central North Sea than the northern North Sea. In all density
logs examined, the lithology with the highest density is chalk. The thicknesses of chalk
are greatest in the central North Sea with clastic sediments becoming the dominant
Cretaceous/Palaeocene deposit at more northerly latitudes. It would therefore be
expected that $\sigma_v$ would be higher at depths within and below the chalk in the central
North Sea. Although this expectation is not supported by the calculated vertical stresses
from these two regions shown in Figure 6.33, it should be noted that there are relatively
few (around 20) vertical stress calculations in the northern North Sea, compared to
around 150 for the central North Sea included in this study. It is concluded therefore
that more density logs should be obtained for the northern North Sea to properly test
this hypothesis.

Observations (b) and (c) can be explained by the effects of compaction and variable
densities. In these explanations it has been assumed that the horizontal stress is simply
generated by the vertical stress, and that the ratio of the vertical to horizontal stress
$\sigma_v/\sigma_h$ is constant. The question must be asked, can this simple explanation account for all
the observations? We continue now to consider some other factors and to look in more detail at the trends observed and summarised in observations (a) - (c).

As sediments are buried, they initially compact rapidly (as all the large pore spaces are closed). However, the rate of compaction decreases with further burial, i.e. the more compacted they are, the more difficult it is to compact them further. Models of compaction have been proposed, such as that of Christie and Sclater (1980), which has been used in section 4.7.5 to estimate the densities of shallow sediments. This model predicts what is intuitively obvious from a consideration of the rate at which compaction occurs with burial, i.e. that the rate of increase of \( \sigma_v \) gradients will fall with depth. Although straight lines have been fitted to the data in Figure 6.33 (for the purposes of a first approximation to the trends) it can be seen from careful examination of the central North Sea data (where there are enough data points to define the trend well) that the rate at which the \( \sigma_v \) gradients increases with depth does indeed fall with depth. In other words, the gradient of the trend of \( \sigma_v \) gradients decreases with depth.

In order to express these subtle gradient changes conveniently, let us call the gradient (either tangent gradient of the trend, or secant gradients of the individual points) of the \( \sigma_v \) gradients (where these are secant gradients as previously defined) the 2nd derivatives of \( \sigma_v \) against depth. Thus, it can be seen in Figure 6.33 that the 2nd derivatives of \( \sigma_v \) against depth fall with depth.

If the values of \( \sigma_h \) in the North Sea were due entirely to the weight of the overburden, and were simply related to the overburden by a constant ratio, it would be expected that the trend of \( \sigma_h \) with depth would be similar to that of \( \sigma_v \) with depth. Perhaps the simplest way to investigate this would be to look at the ratio of \( \sigma_h/\sigma_v \) with depth. However, values of \( \sigma_v \) are only available for the Amerada Hess data (see below). Another way of looking at the same problem is to compare the trends of \( \sigma_h \) or LOLTP gradient with the trends of \( \sigma_v \) gradient. As discussed above, it is expected from a consideration of compaction with depth, and can be seen from a plot of \( \sigma_v \) gradient with depth, that the 2nd derivatives of \( \sigma_v \) against depth decrease with depth. Plots of \( \sigma_h \) or LOLTP gradient against depth have been examined to determine if the 2nd derivatives of \( \sigma_h \) and LOLTP
against depth decrease or increase with depth and thus to establish the relationship between $\sigma_h$ and $\sigma_v$. If the 2nd derivatives of $\sigma_h$ or LOLTP increase with depth, remain constant with depth, or decrease at a lower rate than the 2nd derivatives of $\sigma_v$ with depth, the ratio of $\sigma_h/\sigma_v$ must increase with depth.

The trends of LOLTP gradients presented in plots such as 6.20 have a linear best fit line added simply to indicate a first order approximation to the trend. However, on close examination of Figure 6.20 and perhaps more clearly in Figure 6.21, it appears that the second derivative of the LOLTPs is increasing.

However, it must be remembered that the data presented in Figures 6.20 and 6.21 are from tests conducted in both normally pressured and over pressured formations and it can be seen in Figure 6.29 that many of the higher values of LOLTP, particularly deeper in the central and northern North Sea, are from over pressured formations. Clearly then, $\sigma_h$ can be due to high pore pressures as well as the weight of the overburden and thus the effect of high pore pressure can cause the ratio of $\sigma_h/\sigma_v$ to increase with depth. To investigate if other factors can also cause the ratio of $\sigma_h/\sigma_v$ to increase with depth, the data from overpressured formations must be removed.

The dataset from Figure 6.29 which is from normally pressured formations can be examined. This trend has been examined, and it appears that the best fit to the data is actually that shown in Figure 6.29, i.e. a straight line, so that the second derivative of LOLTPs from normally pressured formations is constant. Thus, even without the effect of overpressure, it can be inferred that the ratio of $\sigma_h/\sigma_v$ increases with depth.

As there is a lot of scatter in the dataset from normally pressured formations in Figure 6.29, a reduced dataset of just LOPs from tests performed in shales has also been examined. This dataset is displayed in Figure 7.7.
Figure 7.7 Erico LOP gradients from tests performed in normally pressured shales in the central and northern North Sea. The 2nd order polynomial which is shown is a considerably better fit to the data than a straight line. The gradient of this line of best fit increases with depth illustrating that the 2nd derivatives of LOP against depth increase with depth.
The line of best fit in Figure 7.7 which, judging by the $R^2$ correlation coefficient values provided by the plotting program, best describes the trend seen in this dataset, is the 2nd order polynomial shown (it can also be seen from inspection without the fitted curve that the gradient of this plot increases with depth). Thus, the second derivatives of LOP against depth increase with depth. This indicates rather more convincingly that the ratio of $\sigma_h/\sigma_v$ in the northern and central North Sea increases with depth.

The 2nd derivatives of the trends defined by the lower bounds of the Erico datasets (Figures 7.2 to 7.5) can also be studied to determine whether the ratio of $\sigma_h/\sigma_v$ changes with depth. These trends are lower bounds to the datasets and have been defined by only a few points which were hand picked as explained in section 7.2.2. The 2nd derivatives of $\sigma_h$ against depth (as defined above) for these datasets have been studied in the same way as the other datasets so far, by using just the few trend defining points, i.e. the secant gradients of the trend defining points are plotted against depth and from this plot it can be seen from inspection whether the 2nd derivatives increase or decrease with depth. Using this method, the 2nd derivatives of Figures 7.2 (all lithologies - central North Sea) and 7.5 (shales only - northern North Sea) are seen to increase with depth, whereas the 2nd derivatives of Figures 7.3 (all lithologies - northern North Sea) and 7.4 (shales only - central North Sea) are seen to decrease with depth. The results from Figures 7.2 and 7.5 again indicate that and the ratio of $\sigma_h/\sigma_v$ increases with depth. The decrease in 2nd derivatives from Figures 7.3 and 7.4 do not indicate that the ratio of $\sigma_h/\sigma_v$ increases with depth, but, as mentioned above, this does not necessarily indicate that the ratio of $\sigma_h/\sigma_v$ decrease with depth.

Finally, the actual ratios of $\sigma_h/\sigma_v$ from the Amerada Hess dataset have been studied. These have actually already been presented in Figure 6.35, where it is clear that the ratios of $\sigma_h/\sigma_v$ do increase with depth in the central and northern North Sea. As there are only three data points from the southern North Sea, and because of the complexities introduced by the presence of a stress drop in at least some parts of the southern North Sea, the apparent decrease in the ratio of $\sigma_h/\sigma_v$ with depth shown for the southern North Sea in Figure 6.35 is not considered reliable. The dataset presented in Figure 6.35 for the central North Sea contains results from five tests performed in overpressured
formations. However, when the data points from overpressured formations are removed, the values of $\sigma_h/\sigma_v$ ratio still increase with depth.

In conclusion, it can be seen from the trends of the 2nd derivatives of LOLTPs and $\sigma_h$ with depth (in Figures 7.7, 7.2 and 7.5) and from actual values of $\sigma_h$ and $\sigma_v$ (Figure 6.35) that the ratio of $\sigma_h/\sigma_v$ increases with depth in the central and northern North Sea.

An increase in the ratio of $\sigma_h/\sigma_v$ with depth can be caused by at least two factors. (i) The rock properties might change with depth such that more of the vertical stress is “transferred” to the horizontal direction (ii) Another factor, other than the weight of the overburden may become increasingly important with depth. A component of tectonic stress that increases with depth, as the rocks become more competent and perhaps more strongly coupled with the basement, is considered the most plausible explanation. These ideas are discussed further in the context of stress models in the next section.

Observation (d)

This observation, that trends of $\sigma_h$ with depth are higher in shales than in sands is based on Figures 6.23 and 6.24. It is also clear from the comparisons between Figures 7.2 and 7.4 and between 7.3 and 7.5 that the trend of $\sigma_h$ with depth is higher when defined using just shales than when using all lithologies. Variations of $\sigma_h$ with lithology have been documented in several other sedimentary basins (Warpinski et al., 1983; Thiercelin and Plumb, 1991; Evans et al., 1989). These variations have been reviewed by Plumb (1994a) who classifies basins as tectonically compressed or tectonically relaxed.

In tectonically relaxed basins (e.g. Piceance basin - Warpinski et al., 1983) $\sigma_1$ is the vertical stress and $\sigma_h$ is lower in sands than in shales. This observation can be explained in terms of the elastic rock properties (essentially due to the fact that Poisson’s ratio tends to be higher in shales than sands) applied to the uniaxial strain model (equations 3.11 and 3.12) or the mechanical properties (essentially due to the fact that the angle of internal friction is higher in sands than shales) applied to the Mohr-Coulomb failure model for normal faulting regimes (equation 3.23).
In tectonically compressed basins (where \( \sigma_1 \) is horizontal) \( \sigma_h \) can be higher in sands than shales. This observation can be explained by the elastic models which incorporate horizontal strain (equations 3.1.3 and 3.15) because the sands tend to be stiffer (have a higher Young's modulus) than the shales and thus transmit more of the horizontally applied stress. The Mohr-Coulomb failure model applied to a strike-slip regime (equation 3.23 with \( \sigma_H \) substituted for \( \sigma_v \)) can also predict higher stress in sands than shales. This prediction is possible because the level of \( \sigma_H \) can vary from bed to bed (as opposed to \( \sigma_v \) in equation 3.23 which must be the same in all beds with the same overburden). In a strike slip faulting regime, \( \sigma_H \) can be higher in stronger beds, i.e. those beds able to transmit higher levels of stress. Sandstones are generally stronger in this sense (i.e. they have higher friction angles) than shales, and thus are likely to transmit higher levels of \( \sigma_H \). The ratio of \( \sigma_v/\sigma_H \) (from equation 3.23 with \( \sigma_H \) substituted for \( \sigma_v \)) however, is still higher in the lithologies with lower friction angles, i.e. shales. Thus the Mohr-Coulomb failure model applied to a strike-slip regime, like the uniaxial elastic strain model incorporating horizontal strain, can predict either sense of lithological stress contrast (\( \sigma_h \) can be higher in sands or shales) depending on the degree of horizontal compression.

The sense of the stress contrast observed in the central and northern North Sea, when viewed in terms of either of the possible stress models, implies that the stress state in the North Sea is either relaxed, or “mildly” compressed.

In the southern North Sea, the amount of LOLTP data from sands is small and the stress state is clearly complicated by the stress drop seen in Figure 6.15. However, the LOLTPs do not appear to be significantly lower in the sands. This implies that the degree of horizontal compression is higher in the southern North Sea. The southern North Sea clearly has a more involved tectonic history than the younger rocks of the northern and central North Sea. This history could be responsible for the stress drop seen in the southern North Sea in some way similar to that suggested by Evans et al. (1989a) for the stress drop beneath the Appalachian Plateau (section 7.5.1). If this is so, there is the possibility of remanent stresses in the rocks of the southern North Sea,
which could mask the effect of stresses induced by loads applied to these rock in the present day. On this basis, with the data available in this study, any statement about the tectonic stress state in the southern North Sea cannot be more than tentative.

In section 7.3.2, this discussion is continued as the actual values of $\sigma_h$ determined for the North Sea are examined in a semi-quantitative way in terms of possible stress models (section 3.4) and suggested values of rock elastic and mechanical properties.

Observation (e)
Figures 6.27 to 6.29 clearly demonstrate the influence of high pore pressure on $\sigma_h$. The effect of high pore pressure on $\sigma_h$ has been observed in sedimentary basins in other studies (Breckels and van Eekelen, 1981; Plumb, 1994a). The gradient of $\Delta$LOLTP (norm.) against $\Delta$P(norm.) (section 6.3.2.3) from Figure 6.27 gives the ratio of increase in LOLTP over increase in pore pressure ($\Delta$LOLTP/$\Delta$P), i.e. how much the LOLTP is raised by a certain rise in pore pressure. Figure 6.27 indicates that this ratio is around 0.3. This means that for a pore pressure rise above hydrostatic of say 10 MPa the LOLTP will rise by around 3 MPa above the average trend. The trend in Figure 6.28 shows the same gradient as in Figure 6.27, thus although changes in pore pressure in the dataset of Figure 6.28 can only be inferred from changes in mud weight, the results are in very good agreement.

This ratio of 0.3 is rather low compared to ratios calculated in the same way using hydro-frac measured values of $\sigma_h$ ($\Delta\sigma_h/\Delta P$) in other sedimentary basins. These ratios are reviewed by Plumb (1994a), who lists values of 0.46 to 0.56 from the US Gulf Coast data of Breckels and van Eekelen (1981), around 0.5 for the Vicksburg formation of South Texas (after Salz, 1977), and ratios of up to 0.8 from the Ekofisk chalk field of the central North Sea during reservoir depletion after Teufel et al. (1991).

Both the elastic models and the Mohr-Coulomb failure models described in section 3.4 predict that an increase in pore pressure will cause an increase in $\sigma_h$. The Mohr-Coulomb failure model however, because of the form of the effective stress law which is applied in this model, predicts a set amount of change in $\sigma_h$ for a given change in pore pressure.
pressure and a given value of internal friction angle. On the other hand the elastic models predict that for a certain increase in pore pressure and a certain value of Poisson's ratio, $\sigma_h$ increases by some amount which depends on the poroelastic properties of the rock.

The difference between the ratios $\Delta$LOLTP/$\Delta P$ from the central and northern North Sea derived from Figures 6.27 and 6.26, and the ratios $\Delta \sigma_h / \Delta P$ reviewed by Plumb (1994a) can be explained in more than one way. It could be that in over pressured sections, the LOLTPs underestimate the value of $\sigma_h$ although there is no apparent reason why this should be the case. On the other hand, the poroelastic response of the rocks sampled by the leak-off tests in this study could be different to those of other studies. The wide range of values for the ratio $\Delta \sigma_h / \Delta P$ reviewed by Plumb (1994a), indicates that the ratio reported here is low but not unreasonable. To further pursue this investigation, detailed stress and pore pressure measurements would need to be made and combined with careful laboratory measurements of the poroelastic properties of the rocks in question.

### 7.3.2 Interpretation of $\sigma_h$ in the North Sea in Terms of Models of Stress with Depth in Sedimentary Basins.

The models of stress distribution with depth given in section 3.4 can be used to predict the value of $\sigma_h$ at some point in the crust. Such predictions have been compared to the observed values of $\sigma_h$ in the North Sea in order to investigate further the causes of the observations discussed in the previous section.

For the purposes of this discussion it is assumed that one principal stress is oriented vertically and thus the other two are oriented horizontally. Rock elastic and mechanical properties have been determined for the rocks in the crust in other parts of the world by two methods: (a) by making laboratory measurements on rock cored from the part of the crust where the stress prediction is to be made (e.g. Thiercelin and Plumb, 1991), or (b) by estimating rock properties from measurements made by wellogs run in the section of the crust where the prediction is to be made. Elastic properties can be estimated from interval transit times of P and S waves, and mechanical properties have been estimated
through empirical relationships between the relevant mechanical properties and log measured porosity and shale content (Plumb, 1994; Katahara, 1995; Katahara, 1996).

The lithologies in which the leak-off tests in the North Sea have been conducted are generally known. However, the actual rock elastic and mechanical properties have not been obtained. Although the logs might exist from which these parameters can be estimated, it has been beyond the scope of this study to obtain and analyse this data. Thus a comprehensive investigation into the origin of the $\sigma_h$ values obtained for the North Sea such as that of Thiercelin and Plumb (1991), is not possible.

However, using the models presented in section 3.4 and some reasonable values for the elastic and mechanical properties of the rock types in which the leak-off tests have been performed, the trends of $\sigma_h$ with depth determined in this study, and some of the observations discussed in section 7.3.1 for the North Sea have been investigated in a semi-quantitative way. In particular, the observation that $\sigma_h$ is not a constant proportion of $\sigma_v$ but could also be influenced by rock property changes and tectonic stress, is investigated.

7.3.2.1 The Uniaxial Elastic Strain Model (UESM)

This model assumes that rocks behave elastically (section 3.4.1). Measured values of Poisson's ratio ($\nu$) for over 100 sandstones and over 30 shales are listed in the Handbook of Mechanical Properties of Rocks (Lama and Vutukuri, 1978). The values of $\nu$ listed range from almost zero (negative values can sometimes be measured for small loads) up to 0.5 (values greater than 0.5, the theoretical limit, are also sometimes measured). The value of $\nu$ measured can be very strongly affected by the presence of cracks which either open or close on loading. Often, the measured value of $\nu$ are seen to increase with increasing applied uniaxial load, as cracks oriented parallel to the direction of application of the stress may begin to open. The "intrinsic" values of $\nu$ should therefore be measured under some confining pressure high enough to prevent the opening and closing of cracks from severely affecting the measurements. The values of $\nu$ measured can still show a dependency on the level of confining pressure. It is reported by Lama and Vutukuri (1978) that weaker rocks can show lower values of $\nu$ at higher
confining pressures, however, they do not provide the experimental results to support this. Values of the dynamic $v$ (section 3.4.1.3) from P and S wave velocities both in oilwell logs and in the laboratory which have been discussed, for example by Xu and White (1994), show that the dynamic $v$ decreases significantly with depth due mainly to compaction and the associated increase in S wave velocities. From this work, it can be inferred that the drained $v$ (which is generally applicable to the UESM) generally does not change significantly with depth.

On the other hand, from an intuitive point of view, it seems possible that as the rock becomes more compact, and voids (cracks and pores) within the rock are closed, the $v$ would increase. This would be because, as the vertical load is applied, the rock can initially respond in the horizontal direction by taking up space occupied by voids. When there is no more internal space, the rock must expand more, laterally, in response to the vertical load. It is thought that the change in $v$ as a result of the processes described here, is again small, over the range of confining pressures encountered in the North Sea. In conclusion, it is not thought that changes in $v$ with depth are likely to be significant in the North Sea, however there is a lack of laboratory measured values of drained $v$ over a range of confining pressures to support this.

If $v$ does increase with depth, it could explain (in the context of the UESM) the observation in section 7.3.1, that the ratio of $\sigma_y/\sigma_v$ increases with depth. If on the other hand, $v$ does not increase with depth, the presence of a tectonic stress component which increases with depth in the central and northern North Sea would be the best explanation of observation (c). [It will be shown below that even if $v$ does increase with depth, for reasonable values of $v$, a component of tectonic stress is still required by the elastic models to explain the level of $\sigma_h$ observed in the North Sea].

Although we cannot be certain about how $v$ varies with depth, we can at least take some typical values of $v$ and see what level of horizontal stress is predicted by the UESM. There is a wide range of possible values of $v$, however, typical values obtained when tests are conducted under confining pressure, and thus representative of in-situ
conditions in the North Sea, are between 0.1 and 0.3 for sandstones and shales, with values in shales generally being slightly higher.

These values can be used in the isotropic UESM (equation 3.11) to predict the value of \( \sigma_b \) for a given value of \( \sigma_v \). The pore pressure can be assumed to be hydrostatic when predictions are to be compared with results from normally pressured formations, and is thus easily calculated. The measured values of the poroelastic constant \( \alpha \) are generally less than 1, however, a value of 1 is generally used in conjunction with equation 3.11 (e.g. Thiercelin and Plumb, 1991) and it will be seen below that for the purposes of this discussion a value of 1 is reasonable.

Assuming that the values of \( v \) are higher in shales than in sands, it is clear from examination of equation 3.11 that the sense of lithological stress contrast seen in the North Sea (i.e. higher values of \( \sigma_b \) in shales than in sands) is predicted by this model. However, it will be seen below that the actual values predicted by equation 3.11 are not in such good agreement with those measured.

The isotropic uniaxial elastic strain model (isotropic UESM):

Using values for \( v = 0.2 \) and \( 0.3 \), and \( \alpha = 1 \), and taking the values of \( \sigma_v \) that have been determined for the Amerada Hess data in the North Sea, we can compare the values of \( \sigma_b \) predicted by equation 3.11 with the actual values determined from the leak-off test pressure volume plots. Figure 7.8 plots the values of \( \sigma_v \) from Figure 6.33, the values of \( \sigma_b \) from leak-off pressure/volume plots (from Figure 6.32) and the values of \( \sigma_b \) predicted by equation 3.11 using the above parameters. For the sake of clarity, only the data from the central North Sea has been used, but it is clear that similar relationships between measured and predicted values would be shown for the northern and southern North Sea.
Figure 7.8 Values of \( \sigma_h \) gradient predicted by the isotropic UESM for Poisson's ratios of 0.2 and 0.3, plotted with the actual values of \( \sigma_h \) gradient determined from the Amerada Hess leak-off test pressure/volume plots.
Using these values of $v$ clearly underestimates the levels of stress in the central North Sea for depths greater than around 1500m, and this underestimation increase with depth. In the southern North Sea, the isotropic UESM using the typical values of $v$ suggested above, would clearly underestimate the level of horizontal stress at all depths. Using a value of $\alpha$ less than 1, worsens the predictions of equation 3.11. Thus, the values of the ratio of $\sigma_h/\sigma_v$ at depths greater than around 1500m in the central and northern North Sea and at all depths in the southern North Sea, implies that equation 3.11 is not capable of describing the stress state in these parts of the North Sea. Thus if it is assumed that the rocks behave elastically, some other factor (not accounted for in equation 3.11) must be at work to cause the levels of stress observed.

The transverse isotropic uniaxial elastic strain model:

The transverse isotropic UESM (equation 3.12) predicts that $\sigma_h$ will be higher than that predicted by equation 3.11 if the value of $v$ measured in the plane of isotropy (the bedding plane) is less than that measured perpendicular to the bedding plane ($v'$). This type of measurement is rarely made, however the measurements made by Thiercelin and Plumb (1991) indicate that the value of $v$ is greater than $v'$ in two cases and less than $v'$ in one case. The ratio of $v$ to $v'$ in any case is only between 0.94 and 1.15 and thus the effect of this difference in equation 3.12 is very small.

Equation 3.12 also predicts that $\sigma_h$ will be higher than that predicted by equation 3.11 if the rock is stiffer (higher Young's modulus - $E$) in the plane of isotropy (the bedding plane) than it is when measured perpendicular to this plane ($E'$). Again such measurements are not often made however, Thiercelin and Plumb (1991) list measured values where the ratio $E/E'$ is between 0.98 and 1.67 in 8 out of 9 cases and is 2.89 in one case. Ratios of $E/E'$ between 0.98 and 1.67 have a small effect on the values of $\sigma_h$ predicted by equation 3.12 and cannot account for the measured values from leak-off tests shown for example in Figure 7.8. A ratio of $E/E'$ of 2.89 however has a significant effect on the values of $\sigma_h$ predicted by equation 3.12 and can indeed predict the sort of level of stress determined from the leak-off tests. As a ratio $E/E'$ has only been measured in one out of nine rock samples, it is considered unlikely that equation 3.12 can explain
the high levels of stress seen throughout the deeper levels of the northern/central North Sea and the southern North Sea. Thus again, if it is assumed that the rocks behave elastically, some other factor (not accounted for in equation 3.12) must effect the levels of stress observed.

In this model, the poroelastic parameters $\alpha$ and $\xi$ can assumed to be 1 and 0 respectively, as again, choosing other values for these parameters worsens the predictions of equation 3.12.

**Elastic models incorporating horizontal strain:**

These models (equations 3.13 and 3.15 which incorporate horizontal strain into the UESM) cannot be tested quantitatively with the data from the North Sea as values of horizontal strain are not available. However, it is clear that they are able to predict much higher values of horizontal stress than the elastic models without horizontal strain, and thus would be capable of predicting the observed levels of $\sigma_h$ in the North Sea. Moreover, by assuming that the amount of horizontal strain increases with depth in the central and northern North Sea (discussed below) the UESM incorporating horizontal strain can easily explain the observed increase in the ratio $\sigma_h/\sigma_v$, without requiring large changes in the values of elastic parameters (such as $v$) with depth.

**Elastic Models: A Discussion:**

The isotropic UESM with no horizontal strain, using reasonable values for the rock properties, is not capable of predicting the level of stress observed in the northern and central North Sea below 1500m, or in any of the southern North Sea. It appears that the transverse isotropic UESM is also not capable of predicting the observed levels of stress in these parts of the North Sea.

If it is assumed that a tectonic stress component is present at deeper levels in the central/northern North Sea and the southern North Sea, and thus there is also a component of horizontal strain, the elastic models (incorporating horizontal strain) are able to account for the observed levels of horizontal stress.
The increase in ratio of \( \sigma_v/\sigma_h \) with depth in the northern and central North Sea, discussed in section 7.3.1, can thus be explained by this component of tectonic stress which increases with depth. It is considered likely that tectonic stress in a sedimentary basin environment such as the central and northern North Sea, would increase with depth. The shallow rocks, close to the sea floor, are relatively recently deposited sediments which are poorly compacted and unlikely to be strongly coupled to the deeper parts of the basin. At greater depth, the rocks become more competent and more strongly coupled to the basement rocks below. From a consideration of plate driving forces (section 3.3.5) it seems likely that it will be the basement rocks that will transmit much of the tectonic stress and thus rocks which are part of the basement, or are strongly coupled to it, are likely to be subject to higher levels of tectonic stress.

In the southern North Sea on the other hand, a component of tectonic stress is required at all levels in order for the elastic models discussed above to predict the observed values of \( \sigma_h \) for reasonable values of \( v \). As discussed in section 7.3.1 and in section 2.3.7.3, the rocks of the southern North Sea are mostly Mesozoic and Palaeozoic rocks which have been buried, compacted and uplifted. It therefore seems likely that these rocks will be more strongly coupled to the basement than the rocks at similar depths in the northern and central North Sea and thus will be more strongly influenced by tectonic stress.

The proposition that a component of tectonic stress becomes increasingly strong with depth in the northern and central North Sea, and is present at all depths in the southern North Sea (although still perhaps increasing with depth) is in agreement with observations of borehole breakouts in the North Sea reported for example by Cowgill et al. (1993) and Cowgill (1994). The orientations of breakouts in the southern North Sea are generally consistent with depth and are parallel to the direction seen for most of NW Europe. In the northern and central North Sea on the other hand breakout directions are seen to be very variable in the younger rocks, but become more consistent, and parallel with the rest of NW Europe in the deeper Mesozoic and Palaeozoic rocks.
The UESM incorporating a horizontal strain that increases with depth in the central and northern North Sea, and is present at all depths in the southern North Sea seems to provide a good explanation of the observed stress in the North Sea. From examination of equations 3.13 and 3.15, it is clear that more stress will be taken up by rocks with a higher Young's modulus for a given amount of horizontal strain. Thus, the presence of a tectonic component of strain included in the UESM acts to increase the level of $\sigma_h$ in stiffer rocks (rocks with higher Young's Moduli) such as sandstones. This counteracts the lithological stress contrast effect caused by the vertical stress inducing higher $\sigma_h$ levels in the "weaker" rocks (those with higher $v$ i.e. shales). Thus, as it appears that a component of tectonic stress is present at all depths in the southern North Sea, equation 3.13 or 3.15 could explain why the lithological stress contrast is not seen in the rocks of the southern North Sea. If the tectonic stress in the central/northern North Sea is increasing with depth, the stress level of $\sigma_h$ in the sandstones would be expected to increase relative to those in shales with depth. This effect can be seen rather weakly in Figure 6.24.

7.3.2.2 The Mohr-Coulomb Failure Model (MCFM)

This model assumes that frictional slip on favorably oriented, pre-existing planes of weakness controls the distribution of stress in the rock mass (section 3.4.2). The internal friction angle ($\phi$) of rocks is generally in the range 30° to 45° (Byerlee, 1978). Plumb (1994) measured values of $\phi$ broadly in the range 20° to 50° and established the following relationships between $\phi$ and sedimentary rock properties: $\phi$ increases with decreasing porosity, $\phi$ increases with decreasing clay content. Therefore, $\phi$ is generally higher in sands than in shales. Within either of these lithologies, as $\phi$ increases with decreasing porosity, it will increase with increasing compaction. Thus, on this basis, $\phi$ would be expected to increase with depth.

On the other hand, from a consideration of the typical convex upward shape of experimentally determined failure envelopes plotted on a Mohr diagram, it can be seen that $\phi$ tends to decrease with increasing confining pressure. It is therefore not clear whether $\phi$ will increase or decrease with depth. On this basis, the observed increase in the ratio of $\sigma_h/\sigma_v$ with depth in the central and northern North Sea may or may not be
explained within the framework of the MCFM by a decreasing value of $\phi$ with depth. [It will be seen below that for reasonable values of $\phi$, the MCFM for a normal faulting regime cannot account for the level of stress observed in the North Sea and thus if frictional slip is assumed to control the distribution of stress, the stress regime must be strike slip or thrust, which implies that there is a significant component of tectonic stress, and it is therefore likely to be this tectonic stress component which causes the increase in the ratio of $\sigma_h/\sigma_v$ with depth in the central and northern North Sea].

As $\phi$ is higher in sands than shales, the lithological stress contrast in the central and northern North Sea (discussed in section 7.3.1) is predicted by the MCFM. However, it is shown below that, as with the UESM (without horizontal strain) in the previous section, although the sense of stress contrast is correct, the actual values of $\sigma_h$ using reasonable values for $\phi$ are too low.

The values of $\sigma_h$ predicted by the MCFM for a normal faulting regime (equation 3.23) for vertical stresses which correspond to the tests from the Amerada Hess dataset (Figure 6.32) have been calculated for $\phi = 30^\circ$ and $45^\circ$ where the pore pressures are hydrostatic and the values of $\sigma_v$ are those presented in Figure 6.33. Again for the sake of clarity, only the data from the central North Sea have been presented here (Figure 7.9).

Clearly, equation 3.23 significantly underestimates the level of $\sigma_h$ in the North Sea using reasonable values of internal friction angle. Again, it is apparent that this model would also be unable to predict the levels of stress in the northern and southern North Sea.

It can be concluded that if the stress in the rocks of the North Sea is controlled by Mohr-Coulomb failure on pre-existing faults, then the stress regime cannot be a normal faulting regime (i.e., that which equation 3.23 pertains to). Clearly, the ratios of $\sigma_h/\sigma_v$ indicate that the thrust faulting regime is not common, at least in the central and northern North Sea. Thus if the MCFM controls the stress levels, the stress regime must be of the strike-slip type.
Figure 7.9 Values of $G_0$, gradient predicted by the MCFM applied to a normal faulting regime for friction angles of 30° and 45°, plotted with the actual values of $G$, gradient determined from the Amerada Hess leak-off test pressure/volume plots.
The MCFM outlined in section 3.4 can also be applied to strike-slip faulting regimes by simply substituting $\sigma_H$ for $\sigma_v$ in equation 3.23. However, in order to predict $\sigma_H$ using this model, the value of $\sigma_H$ must then be known and thus this model cannot be taken any further in a quantitative way without making some big assumptions.

Qualitatively however, it is clear that this model would be capable of predicting the level of $\sigma_H$ in the North Sea. If the MCFM controls the stress state, and assuming the fact that $\sigma_H$ is greater than $\sigma_v$ is due to the presence of tectonic stress (as it is pervasive throughout the basin, tectonic stress seems the only reasonable explanation) the increase in $\sigma_H$ gradient with depth again implies that the effect of tectonic stress becomes more important with depth in the central and northern North Sea. Thus although (as discussed above) the value of $\phi$ might decrease with depth, it is unlikely that this is the main reason for the increase in the $\sigma_H/\sigma_v$ ratio with depth.

In southern North Sea, a strike slip (as the ratio of $\sigma_H/\sigma_v$ is close to 1 in many cases) possibly thrust regime must exist if the stress distribution is assumed to be controlled by the MCFM. In the southern North Sea however, the ratio of $\sigma_H/\sigma_v$ is inferred not to increase greatly with depth, and thus the component of tectonic stress (although it may still increase to some extent with depth) would appear to be significant at all depths.

As discussed in section 7.3.1 (observation d) the observed lithological stress contrast in the central/northern north sea and the lack of stress contrast in the southern North Sea can be explained by the MCFM applied to a strike-slip regime where the degree of horizontal compression is greater in the southern North Sea. The slight decrease in stress contrast with depth seen in the central/northern North Sea can also be explained by the MCFM applied to a strike-slip regime in which the degree of horizontal compression increases with depth.

As discussed in section 2.5, there are very few earthquake focal mechanisms available for the North Sea. The exception to this is the region offshore Norway, where a variety of focal mechanisms are seen including normal faulting, strike slip faulting and thrust faulting. Some of these mechanisms are thought to be influenced by post glacial

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rebound. The stress regime in the North Sea generally is rather poorly constrained by focal mechanisms and thus the hypothesis that the stress regime in the North Sea is strike-slip cannot be well supported.

7.3.2.3 Stress Models and the Origins of $\sigma_h$ in the North Sea: Summary

The isotropic UESM predicts the sense of lithological stress contrast seen in the central and northern North Sea. However, the isotropic UESM cannot account for observed levels of $\sigma_h$ below 1500m in the central/northern North Sea, or in the southern North Sea above the stress drop. Also, the isotropic UESM cannot account for the increase in the ratio of $\sigma_y/\sigma_v$ with depth, without assuming a significant increase in $\nu$ with depth.

The effect of transverse isotropy on the UESM is thought to only be small in the majority of cases and is therefore considered unlikely to account for the levels of $\sigma_h$ seen in the North Sea.

The UESM modified to incorporate a component of horizontal strain is able to account for the stress levels seen below 1500m in the central and northern North Sea, and in the southern North Sea. Moreover, if it is assumed that this component of strain increases with depth in the central/northern North Sea, this model can also account for the observed increase in ratio of $\sigma_y/\sigma_v$. The presence of a component of tectonic stress which increases with depth (perhaps as the rocks become stronger and more strongly coupled to the basement) in the central and northern North Sea is the best explanation for these observations in terms of elastic models. The data from the southern North Sea indicate the presence of a tectonic stress component which is significant at all depths (i.e. does not increase greatly with depth) which is perhaps because the rocks of the southern North Sea have been buried and inverted and are thus generally compact, strong and more coupled to the basement.

The presence of a tectonic component of stress (incorporated in the UESM) which does not increase greatly with depth in the southern North Sea, but does increase with depth in the central and northern North Sea explains the lack of clear lithological stress
contrast in the southern North Sea, and the decrease in lithological stress contrast with depth in the central/northern North Sea.

The MCFM for a normal faulting regime also predicts the sense of lithological stress contrast seen in the central/northern North Sea. However, this model assuming a normal faulting regime, is unable to predict the level of $\sigma_b$ observed in the any part of the North Sea, except perhaps the very shallow parts of the central/northern North Sea.

The MCFM applied to a strike-slip regime is able to predict the level of $\sigma_b$ in the North Sea. It can also explain both the observed lithological stress contrast in the central/northern North Sea, the lack of stress contrast in the southern North Sea. The strike-slip regime would most probably be due to the presence of a tectonic stress component. If this tectonic stress component increases with depth in the central/northern North Sea it could explain the observed increase in the ratio of $\sigma_b/\sigma_v$ with depth. Although it is not clear how much $\phi$ might increase or decrease with depth, in view of the existence of a tectonic stress component, it is considered unlikely that changes in $\phi$ are the main cause of the observed increase of $\sigma_b/\sigma_v$ with depth in the northern and central North Sea. Again, the data from the southern North Sea indicate the presence of a tectonic stress component which is significant at all depths (i.e. does not increase greatly with depth).

7.3.3 The Origin of the Southern North Sea Stress Drop

The drop in inferred values of $\sigma_b$ with depth (the stress drop) seen in the data from the southern North Sea Ravenspurn field presents some challenging questions. A similar stress drop has been observed beneath the Appalachian Plateau (Evans et al., 1989), were extensive and very detailed, bed by bed hydraulic fracturing stress measurements were made, as well as laboratory tests for elastic and mechanical properties on rock core samples, in an attempt to explain the observed stress drop (Evans et al., 1989; Evans et al., 1989a; Evans 1989; Engelder, 1993). Evans et al. (1989a) discuss several possible explanations for the stress drop which could also apply to the southern North Sea. However, they concluded that the origins of the stress drop remained somewhat enigmatic. In the light of the discussions above concerning stress models in sedimentary
basins, and the suggestions of Evans *et al.* (1989a), the following explanations for the southern North Sea stress drop are possible.

(a) As the stress drop is observed to lie directly below the thick Zechstein Salt, it could be that the rocks below the salt are mechanically detached from the rocks above. For the stress to be lower in the rocks below the detachment would require the upper layer to be transmitting a relatively large amount of tectonic stress which is not present in the lower layer. From a consideration of the possible plate driving forces outlined in section 3.3 it is difficult to imagine how such a situation could arise.

(b) From a consideration of the stress models outlined in section 7.3.2, a possible explanation for the stress drop could be that either the elastic properties (if the UESMs are assumed to control the stress) or the mechanical properties (if the MCFMs are assumed to control the stress) are very different between the rocks below the salt and the rock above the salt. Generally such large changes in rock properties would not be expected. However, there is one other factor which changes across the salt which it is known can affect at least some rock elastic properties. This factor is the presence of gas in the rocks below the salt. The leak-off tests and hydro-fracs below the salt were performed in the rocks just above the Ravenspurn gas reservoir and within the Ravenspurn gas reservoir respectively. The rocks just above the reservoir are mainly siltstones and may well also be gas bearing, although not permeable enough to be part of the productive reservoir. As outlined in section 3.4.1.3, the undrained $v$ of gas bearing rocks can be much less than that of water saturated rocks. Therefore if undrained conditions prevail, and the rocks do behave in the way prescribed by the boundary conditions of the UESMs, levels of $\sigma_3$ could be considerably lower in the gas bearing lithologies.

It is not known how (if at all) the presence of gas might affect the internal friction angle of a porous rock, thus the effect of gas on the values predicted by the MCFM are not discussed here.
(c) The presence of remanent stresses in the rocks below the salt of the southern North Sea could cause the present day stress magnitude (in the orientation parallel to the tensile component of this remanent stress) to be lower in these rocks. The term remanent stresses used here in the same sense as it is used by Evans et al. (1989a) to mean a localised palaeo-stress component arising from elastic strains locked into the rock by antagonistic force balance. If such a palaeo-stress were locked into the rock in such a way that the tensile component (that which in the antagonistic force balance is counteracted by the compressive component) acted perpendicular to the present minimum horizontal stress direction, a reduction in the total in-situ \( \sigma_h \) value would occur.

Remanent stresses in the North Sea were discussed by Cowgill (1994) as a possible explanation for variable borehole breakout directions in rocks from different chronostratigraphic unit. This is considered a reasonable explanation when the anisotropy of the stress field arising from present day tectonic stresses is relatively small such that relatively small remanent stresses will affect the in-situ stress orientation. However, the magnitude of the stress drop in the southern North Sea is at least several MPa. Whether such a magnitude of remanent stress can persist over considerable periods of Geological time in these rocks is unclear.

(d) The most likely explanation of the stress drop in the Appalachian Plateau is considered by Evans et al. (1989a) to be the effect of a palaeo-overpressure. It has been shown from detailed analysis of the fabric of the rocks in which the stress drop is seen that they once hosted highly over pressured formation fluids which prevented the rock from compacting as it was buried. The analysis also indicates that the matrix of the rock was uncemented at the time the over-pressure formed, but that it became cemented while the overpressure was still present. Some time after the rock had been cemented, the overpressure was released and the rock began to contracted elastically under the confining pressure which had previously been balanced by the high formation pressure. The overlying and underlying formations however, had not hosted high pore pressures, and thus as the overpressured rock contracted it was constrained laterally due to coupling above and below with the normally pressured rocks. The release of the
overpressure thus allowed the rock to contract vertically, but tended to induce horizontal tensile stresses. The magnitude of these tensile stresses has been calculated by Evans et al. (1989a) (for the particular elastic properties of the rock of interest in the Appalachian Plateau) as being up to 15 MPa where the depth of burial was assumed to be just 1700 m.

Clearly, if the type of mechanism described above for the rocks of the Appalachian Plateau took place in the rocks of the southern North Sea which could have been at depths greater than the 1700 m assumed for above, it would be easily capable of producing a stress drop of a magnitude large enough to account for that observed. Although this should be recognised as a possible explanation for the stress drop in the southern North Sea, much more detailed work would be needed to establish if these rocks once hosted an overpressure that was comparable in terms of its timing of formation and release to that of the Appalachian Plateau.

Conclusions:
The stress drop observed in the data from the Ravenspurn field is an anomaly within the North Sea (the extent of which within the southern North Sea is not known) which can be explained in a variety of ways (above). However, to distinguish between the various possible explanations outlined above requires much detailed analysis of the geological history of the rocks and careful measurement of the relevant elastic and mechanical properties of the rock. This work is clearly beyond the scope of this study, however it would provide a fascinating topic for future research.

7.3.4 The Distribution and Origin of Stress in the North Sea: Conclusions.
Based on the general discussions in sections 7.3.1 and 7.3.3, and the semi-quantitative discussion in the context of stress models in section 7.3.2, the following conclusions concerning the state of stress in the North Sea can be made.

The gradients of $\sigma_h$ increase with depth in the central/northern North Sea but are almost constant with depth (excluding the stress drop) in the southern North Sea. This can be partly explained by the effect of compaction (causing the gradients of $\sigma_h$ to increase with
depth and thus the gradient of $\sigma_h$ to also increase with depth) in the central/northern North Sea, and the relative lack of compaction with depth in the southern North Sea.

However, from the trends of LOLTP and $\sigma_h$ gradients with depth (particularly in over pressured formations but also in normally pressured formations) in the central/northern North Sea that generally the ratios $\sigma_h/\sigma_v$ increases with depth. This is also evidenced by the ratios $\sigma_h/\sigma_v$ from the Amerada Hess data, which are seen to increase with depth in the central/northern North Sea. This implies either that the properties of the rock (Poisson’s ratio or internal friction angle) are changing with depth, or that some other factor, such as a component of tectonic stress becomes increasingly important with depth.

The values of $\sigma_h$ observed in the North Sea have been further investigated in a semi-quantitative way using UESM and MCFM of stress in sedimentary basins. This investigation has shown that, using reasonable values for the rock properties, both the UESM and the MCFM require the presence of a tectonic stress to predict the observed values of $\sigma_h$ in the North Sea. Thus, the properties of the rock (Poisson’s ratio or internal friction angle) may or may not change systematically with depth, but the presence of a tectonic stress component that increases with depth is likely to be the main cause of the observed increase in the ratio $\sigma_h/\sigma_v$. This component of tectonic stress must increase (from approximately zero at the sea floor) with depth in the central and northern North Sea and must be present at significant levels at all depths in the southern North Sea. This distribution of tectonic stress is consistent with the observed patterns of borehole breakouts in the North Sea.

The observed lithological stress contrast in the central/northern North Sea and the lack of this stress contrast in the southern North Sea are predicted by both the UESM incorporating horizontal strain and the MCFM for strike-slip regimes where horizontal compression is assumed to increase with depth in the central and northern North Sea, and to be significant at all depths in the southern North Sea.
The fact that there is a generally low level of seismicity in the North Sea (section 2.5) does not support the hypothesis that shear failure controls the levels of stress in the North Sea. However, on the basis of the data studied here, it is concluded that more detailed stress measurements and laboratory rock property measurements are needed to distinguish which of the two types of model is most suitable for the North Sea.

At shallow depths in the central and northern North Sea, above around 1500m, the levels of stress can be explained by either the UESM or the MCFM with no tectonic stress. These rocks are thus probably not technically compressed and the stress regime at these shallow depths is most likely to be a normal faulting regime.

If the MCFM is assumed to control the stress levels in the North Sea below around 1500m, the stress regime must be strike-slip in the central/northern North Sea, and either strike-slip or thrust in the southern North Sea. However, if the UESM incorporating horizontal strain is assumed to control the stress, the stress regime could be either a strike-slip regime or a normal faulting regime. If it is a normal faulting regime, it is one in which tectonic stress exists, but not to the extent that $\sigma_h$ is greater than $\sigma_v$.

High pore pressures give rise to high levels of $\sigma_h$ in the North Sea. However, an increase in pore pressure would not affect the vertical stress. Thus, particularly at deeper levels in the central/northern North Sea, where $\sigma_h$ is close to (but less than) $\sigma_v$, increases in pore pressure can easily change the stress regime from normal or strike-slip to thrust. The possibility of determining the stress regime from a combination of borehole breakouts and $\sigma_h$ estimates is investigated in section 7.4.

The two models of the distribution of stress (UESM and MCFM) have quite different assumptions about the way the rocks behave, and thus what controls the distribution of stress in the North Sea. However, a component of tectonic stress which increases with depth in the central and northern North Sea but is present at significant levels at all depths in the southern North Sea, is required to explain the distribution of stress within the framework of both models.
It should be remembered that the UESM assumes that the rock behaves in an ideal elastic way, and that the MCFM assumes that the rock behaves as an ideal Mohr-coulomb material. If real rocks under in-situ condition, over geological time scales, do not behave in these ideal ways, then such models are not applicable, and the distribution of stress in the North Sea would not be governed by the elastic and mechanical parameters which have been part of this discussion. In particular, processes such as creep (caused for example by pressure solution) could occur in the deeper parts of the North Sea basin, where confining pressures and temperature are moderately high. It is thought that shales in particular could be susceptible to such processes. If this were true, and creep did become an important factor at depth in the North Sea, it would be expected that levels of $\sigma_h$ would increase with depth at a higher rate in shales than in sands. This is not supported by the data presented in Figure 6.24. The possibility that processes such as creep play an important role in determining the distribution of $\sigma_h$ in the North Sea is discussed further in section 7.5.2 where the trend of $\sigma_h$ with depth in the North Sea is compared to trends of $\sigma_h$ with depth in other areas.

7.4 Estimating $\sigma_h$ from a Combination of Breakout Observations and Leak-off Test Data.

Using a combination of $\sigma_h$ estimates from leak-off tests and observations of borehole breakouts, it is possible to estimate a lower limit to the value of the magnitude of $\sigma_h$.

In a section of borehole where a leak-off test has been performed (and thus an estimate of $\sigma_h$ can be obtained) and a borehole breakout has been detected, it is possible, by assuming reasonable values for rock strength properties and by adopting an appropriate rock failure criterion, to estimate the lower limit of the maximum horizontal stress ($\sigma_{h*}$). Determining the magnitudes of $\sigma_h$ throughout the North Sea using this method has not been possible with the datasets obtained for this study. This is because the occurrence of both a leak-off test and a detected borehole breakout in the same section of borehole is rare. However, a possible method is briefly described in this section, and two examples of preliminary determinations of the lower limit of $\sigma_h$ are given. The lower limits of $\sigma_h$ (for the few cases where it has been possible to apply this method to the North Sea
datasets) together with the magnitudes of $\sigma_h$ and $\sigma_v$ (and thus the stress regime) are presented.

The methodology is outlined below. Further details and the relevant equations are included in Appendix 1. Firstly, it must be assumed that, as is generally accepted, borehole breakouts form due to shear failure of the borehole wall (Bell and Gough, 1979; Zoback et al., 1985). Secondly, a suitable failure criterion must be chosen which can reasonably represent the failure of the rock in the borehole wall. An example of such a criterion is the Mohr-Coulomb criterion which is discussed with reference to borehole breakout formation by Zhou (1994) and Zoback et al. (1985). Then, from a consideration of the pre-breakout stress concentration around the borehole wall (see section 4.3 and Appendix 1) in a part of the borehole where a breakout has occurred, and using the estimate of $\sigma_h$ from the leak-off test in conjunction with the chosen failure criterion, and reasonable values for rock strength properties (cohesion and coefficient of internal friction) a lower limit can be placed on the magnitude of $\sigma_h$. The equations which describe the principal stress concentrations at the borehole wall, the Mohr-Coulomb criterion and the equations relating the rock strength properties, and the procedure for using these equations to estimate the lower value of $\sigma_h$ are all given in Appendix 1.

Some of the criteria which have been used to predict rock failure during borehole breakout formation (including the extended von Mises criterion and the modified strain energy criterion) have been discussed by Zhou (1994). Although it is acknowledged that there may be more suitable failure criteria for application to the formation of borehole breakouts, for the purposes of this preliminary exercise in investigating the magnitude of $\sigma_h$, only the Mohr-coulomb criterion is considered.

An extensive study of borehole breakouts in the North Sea has been performed by Cowgill (1994), Cowgill et al. (1993) and Cowgill et al. (1995). The same methodology applied in these studies has been used here to detect breakouts in some of the wells for which leak-off test records have been obtained from Amerada Hess (section 5.4). Where
breakouts have been detected in the sections that leak-off tests records are available, a lower limit for $\sigma_H$ has been estimated.

Since the calculations can become quite involved, a computer program (Zhou, 1994) has been utilised to predict the occurrence and/or non occurrence of breakouts. This program simulates borehole breakouts when rock properties and in situ conditions of stress are input. The output of the program is a schematic representation of the size, shape and orientation (relative to in-situ stresses), of the predicted breakout. Some examples are given below of how the program can be used to estimate at least a lower bound for $\sigma_H$.

Before this type of exercise can be performed, it is necessary to adopt some standard for breakout size. A reliable breakout, according to the criteria of Plumb and Hickman (1985), is one which shows an elongation in one direction (a borehole eccentricity) of more than 0.6 cm and is consistent over a length of more than 30 cm. For this study, only breakouts which conform to these criteria have been used. Thus, when estimating the lower limit of $\sigma_H$ by way of the breakout simulator, a breakout of size comparable to an eccentricity of at least 0.6 cm must be predicted. The breakout simulator does not include absolute borehole diameter, but shows the ratio of the predicted breakout width to borehole width. This ratio can be scaled to the actual borehole diameter to determine the actual breakout width.

*Examples of estimating lower limits for $\sigma_H$ using the breakout simulator.*

(i) Borehole 20/4b-3.

In borehole 20/4b-3, which is in the south Halibut Basin of the central North Sea, a breakout has been detected from dipmeter logs between 3075 and 3150 m. The borehole is deviated by less than 2° from vertical at this point, and for approximately 85% of the length of this section, the borehole is broken out by between 0.6 and 0.75 cm. The orientation of the breakout (inferred $\sigma_h$ direction) is 086°. The breakout is in the Kimmeridge Clay formation and the gamma ray log clearly indicates that the lithology is shale.
A leak-off test has been conducted at casing shoe depth of 3077 m. The final well report indicates that leak-off occurred, and that the leak-off pressure is 63.2 MPa. Applying the same relationship to this leak-off pressure as was applied to the Amerada Hess dataset in section 6.4.2, yields a value for $\sigma_h$ of 61.4 MPa. Although no pressure/volume plot was available for this leak-off test, the value of $\sigma_h$ given above lies on the trend line for the central North Sea Amerada Hess (shales and claystones/siltstones) data presented in Figure 7.1. The final well report also indicates the pore pressure to have been slightly above hydrostatic at 33.4 MPa and the mud weight to have been equivalent to 36 MPa. Some initial values for the rock strength parameters must be chosen. Reasonable values for the given rock type based on published experimental results (Jaeger, 1969; Byerlee, 1978; Fjaer et al.; 1992) should be the starting point. The rock properties needed for breakout prediction based on the Mohr-Coulomb criterion are coefficient of internal friction ($\mu$) and cohesion ($C_u$).

The parameters in the breakout simulation which are known, are therefore minimum horizontal stress ($\sigma_h = 61.4$ MPa), vertical stress calculated from geophysical logs ($\sigma_v = 65.8$ MPa), pore pressure ($p = 33.4$ MPa) and pressure due to mud weight ($p_m = 36$ MPa). Let these parameters be called the set parameters.

These set parameters have been used as the input parameters to the borehole breakout simulation program (Zhou, 1994 - see Appendix 1). Initial values for $\mu$ of 0.6 and $C_u$ of 10 MPa are at the low end of the range of values considered reasonable based on the published data given by the references above. Choosing relatively low estimates for these rock parameters is equivalent to assuming that the rock is relatively weak. This is an appropriate assumption, as it is the lower bound of $\sigma_h$ that is being estimated. A range of values of $\sigma_h$ have been tried as the final input parameter in order to find the minimum value of $\sigma_h$ that will cause breakout.

A small amount of experience soon gives a reasonable idea as to how suitable the rock strength parameters are. If the rock strength parameters are set too low, the Mohr-Coulomb criterion predicts breakouts at all azimuths, i.e. borehole collapse. For
example, using the set parameters given above for borehole 20/4b-3 where \( \mu = 0.6 \), \( C_u = 10 \text{ MPa} \) and with \( \sigma_H \) initially set equal to \( \sigma_v \) predicts that shear failure of the borehole wall will occur at all azimuths. As the horizontal stress is isotropic in this case, the failure presumably occurs due to the shear stress caused by the difference between the vertical stress and the horizontal stresses. As it is clear from the dipmeter data of borehole 20/4b-3, that the borehole has not collapsed, but has simply broken out in one direction, the actual strength of the rock must be higher than that used to produce the simulation described above.

To increase the strength of the rock in the simulation, either the coefficient of internal friction or the cohesion can be raised. By raising either \( \mu \) to 0.8 or \( C_u \) to 15 MPa, and keeping \( \sigma_H = \sigma_v \) no breakouts are predicted.

Using the set parameters, with either combination of rock strength parameters (\( \mu = 0.8 \) and \( C_u = 10 \text{ MPa} \) or \( \mu = 0.6 \) and \( C_u = 15 \text{ MPa} \)) no breakout is predicted until \( \sigma_H = 64 \) MPa. The breakouts predicted using these parameters do not become large enough to produce an eccentricity equivalent to 0.6 cm until \( \sigma_H \) is close to 70 MPa.

Figure 7.10 shows borehole 20/4b-3 with breakouts around the azimuth of \( \sigma_H \) simulated using the set parameters above (with \( \mu = 0.8 \) and \( C_u = 10 \text{ MPa} \)) and with \( \sigma_H = 70 \text{ MPa} \). The borehole breakout shown, is equivalent to an eccentricity of around 0.76 cm for the 17.6 cm diameter section of 20/4b-3.

Figure 7.11 shows borehole 20/4b-3 with breakouts at the azimuth of \( \sigma_H \) simulated using the set parameters above (with \( \mu = 0.6 \) and \( C_u = 15 \text{ MPa} \)) and with \( \sigma_H = 70 \text{ MPa} \). Again, a breakout is predicted. In this case the eccentricity caused by the breakout is equivalent to approximately 0.63 cm.
Figure 7.10 Breakout simulation for borehole 20/4b-3. Shape and orientation of predicted breakout (shaded) is shown for the set parameters (given in the text) and with $\mu = 0.8$, $C_u = 10$ MPa and $\sigma_H = 70$ MPa.

Figure 7.11 Breakout simulation for borehole 20/4b-3. Shape and orientation of predicted breakout (shaded) is shown for the set parameters (given in the text) and with $\mu = 0.6$, $C_u = 15$ MPa and $\sigma_H = 70$ MPa.
It is clear then, that for either of these combinations of strength parameters, the value of \( \sigma_h \) at a depth of around 3075 and 3150 m in borehole 20/4b-3 must be at least 70 MPa.

Figure 7.12 shows the simulation performed again for the same parameters as in 7.10 but using a value for \( \sigma_h \) of 80 MPa. As would be expected, as the horizontal stress anisotropy increases, the breakout becomes larger. The breakout shown in Figure 7.12 corresponds to a borehole eccentricity of around 1.5 cm for a 17.6 cm diameter hole. This is somewhat larger than the eccentricity measured in most of the broken out section of 20/4b-3. The magnitudes of the three principal stresses at 3076 m in borehole 20/4b-3 are thus estimated to be: \( \sigma_v = 65.8 \) MPa, \( \sigma_h = 61.4 \) MPa, \( \sigma_h > 70 \) MPa. The stress regime is therefore strike-slip, although active strike-slip faulting is not necessarily predicted.

(ii) Borehole 15/22-C3

In borehole 15/22-C3, which is on the flanks of the Witch Ground Graben of the central North Sea, a breakout has been detected from dipmeter logs between 4000 and 4060 m (Measured Depth - MD). The log was stopped at a top depth of 4000 m. The borehole is
deviated by less than 5° from vertical at this point, and for approximately 50% of the length of this section, the borehole is broken out by between 0.6 and 0.78 cm. The orientation of the breakout (inferred direction of $\sigma_b$) is 070°. As is the case for the breakout in 20/4b-3, this breakout is in the Kimmeridge Clay formation and the gamma ray log clearly indicates that the lithology is shale.

A leak-off test has been conducted at casing shoe depth 3998 m MD (3712 m TVD). The final well report and Geoservices pressure/volume plot indicate that a re-opening type leak-off occurred. The value of $\sigma_b$ estimated from the pressure/volume plot is 79.7 MPa. The final well report indicates the pore pressure to have been hydrostatic at 37.9 MPa and the mud weight to have been equivalent to 44.5 MPa.

The calculated vertical stress at this point in the well is 78.9 MPa. The vertical stress is slightly less than the LOP, therefore the LOP can only be considered a lower bound to the value of $\sigma_b$ as the leak-off test may have opened a horizontal fracture and thus be sampling the vertical stress. The value of $\sigma_b$ can therefore be considered to be 78.9 MPa or more. Assuming the stress state is not isostatic (and the presence of a breakout indicates that it is not) a thrust faulting regime must exist at this depth in 15/22-C3. Although the stress regime is already known, it is still instructive to run the breakout simulation program in order to quantify the horizontal stress anisotropy.

The breakout simulation program has been run using the set parameters: minimum horizontal stress ($\sigma_b = 78.9$ MPa), vertical stress ($\sigma_v = 78.9$ MPa), pore pressure ($p = 37.9$ MPa) and pressure due to mud weight ($p_m = 44.5$ MPa) and starting with rock strength parameters with $\mu = 0.6$ and $C_u = 10$ MPa, and with $\sigma_{hi} = \sigma_b (78.9$ MPa). Although the effective radial stress ($\sigma'_r$) at the borehole wall (due to the difference between the mud weight and the pore pressure) is higher in this case than in 20/4b-3, the Mohr-Coulomb criterion still predicts breakouts at all azimuths, i.e. complete borehole wall collapse, for the above input parameters. As complete borehole wall collapse is not observed from the dip-meter log for this section, it can be concluded that the rock strength parameters used above are too low.
As in the case of borehole 20/4b-3, by raising either $\mu$ or $C_u$, and keeping $\sigma_h = \sigma_b$ a situation in which no breakouts are predicted can be obtained. For the above parameters, and with $\sigma_h = \sigma_b$, no breakouts are predicted when either $\mu = 0.6$ and $C_u = 15$ MPa, or $\mu = 0.85$ and $C_u = 10$ MPa.

Breakouts of approximately 0.63 cm are predicted by the Mohr-Coulomb criterion using the set parameters and with $\mu = 0.85$ and $C_u = 10$ MPa when $\sigma_h = 90$ MPa. Breakouts of approximately 0.63 cm are also predicted by the Mohr-Coulomb criterion using the set parameters and with $\mu = 0.6$ and $C_u = 15$ MPa when $\sigma_h = 86$ MPa.

It can therefore be concluded that the stress regime at 3712 m TVD in borehole 15/22-C3 is a thrust faulting regime in which $\sigma_v = 78.9$ MPa, $\sigma_b > 78.9$ MPa and $\sigma_h > 88$ MPa.

This methodology outlined for the two examples above (boreholes 20/4b-3 and 15/22-C3) has been applied where leak-off test data is available for sections of borehole for which breakouts have been detected. As mentioned previously, such combinations of data are rare. The results of the analyses, where they have been possible to perform in the boreholes listed below, all indicate a strike-slip regime. The boreholes and depths (TVD) are as follows: 47/2-1 (2940 m), 211/23b-11 (3096 m), 210/24a-a1 (1905 m), 210/19-4b (1910 m).

Conclusions

Further suitable data, particularly coupled with better constrained estimates of rock strength parameters, would significantly add to the work presented in this section. However, on the basis of the data available, at the depths listed above, the predominant stress regime in the North Sea is strike-slip. At greater depths (and possibly in overpressured zones) thrust faulting regimes may exist. The existence of a strike-slip regime at these depths in the central/northern North Sea (and possibly at all depths in the southern North Sea) is in agreement with the conclusions in section 7.3.4, based on leak-off test data alone.
7.5 Comparison of Trends with Depth of $\sigma_h$ between the North Sea and Other Areas.

In this section, the trend with depth of $\sigma_h$ in the North Sea is compared with trends derived from a variety of stress determination methods from other areas around the world. Firstly, the data which constitutes these other trends, and the location from which the data was obtained, are briefly described. Where trends of $\sigma_h$ with depth are available, they are all plotted together, including the trend derived in this study for the North Sea, for comparison. The differences and similarities of trend with depth of $\sigma_h$ between the North Sea and the other areas is discussed largely in the context of the type of geological setting within the crust that these trends have been established.

7.5.1 Trends of $\sigma_h$ with Depth Outside the North Sea.

In this section, trends of $\sigma_h$ with depth from around the world, which have been presented by a variety of authors using several stress determination techniques, are presented. Not all of the data which have been used to define the trends of $\sigma_h$ with depth (below) come from areas geologically similar to the North Sea. Hence the validity of such a comparison is discussed in section 7.5.2. Indeed, other work which has produced trends of stress with depth for various geological settings is reviewed by Engelder (1993) who divides his discussion into two: “stress as a function of depth in crystalline rock” and “stress as a function of depth in sedimentary basins”. The trends described below have also been grouped on the basis of this distinction.

*Trends of $\sigma_h$ with depth in crystalline rock:*

One of the best known and most wide ranging compilations in terms of geographical location and method of determination is provided by Brown and Hoek (1978). The main aim of their compilation was to provide a guide to the pre-excavation state of stress for mining and tunneling engineers. For this reason, the data is biased towards sampling stress in “hard rock” that is predominantly metamorphic and crystalline rock as opposed to the “soft rock” which is the predominant fill of sedimentary basins. The stresses sampled therefore come largely from cratonic blocks and basement forming rocks which, as discussed below, may form a part of the lithosphere that is structurally and mechanically distinct from that formed by the rocks found in sedimentary basins.
Brown and Hoek (1978) have produced a compilation of 120 stress measurements (including results from overcoring and hydraulic fracturing) which come mainly from mines and tunnels in Australia, North America, Scandinavia and South Africa. The vertical stress in their compilation increases on average, from zero at the surface, linearly with depth. The average horizontal stress \((\sigma_h + \sigma_H / 2)\) divided by vertical stress is plotted against depth (Figure 7.13). At shallow depths, it is clear from this data compilation, that there is a very wide range of horizontal stress magnitudes and that high horizontal stresses are very common. From this dataset, it could be concluded that horizontal stresses several times the magnitude of the vertical stress are the norm. As it is the average horizontal stress against depth that is displayed by Brown and Hoek, it is not possible to compare this data directly with the trend of \(\sigma_h\) with depth in the North Sea. The dataset is summarised in Table 7.2.

![Image](https://via.placeholder.com/150)

**Figure 7.13** Variation of the ratio of average horizontal stress to vertical stress \([\sigma_h + \sigma_H / 2] / \sigma_v\) with depth. (After Amadei and Stephansson, 1997; adapted from Brown and Hoek, 1978).

Rock stress magnitudes have been measured using mainly overcoring (and more recently hydraulic fracturing) techniques in the Baltic Shield and the Caledonides of Fennoscandia since the 1950s. A compilation of this data is reported by Stephansson et al. (1986). The distinction made between the rocks sampled in Brown and Hoek’s study (above) and the rocks of sedimentary basins, can also be made here.
From the Fennoscandian Rock Stress Database the trends of stress with depth have been compared for the different measurement techniques (three types of overcoring technique and hydraulic fracturing). Figure 7.14 shows the magnitudes of the mean principal stress for 142 cases in Fennoscandia determined from the Leeman type overcoring technique (Stephansson et al., 1986), plotted against depth. As might be expected in view of the data presented by Brown and Hoek (Figure 7.13) the data indicates a large scatter in the magnitude of the stresses versus depth. However, whereas the dataset of Brown and Hoek (1978) comes from many locations from around the world, the Fennoscandian database consist of data from a relatively restricted geographic range, i.e. Fennoscandia, and thus less variation in the stress measurements might be expected. Stephansson et al. (1993) discuss the variation in stress magnitudes measured in Fennoscandia. They suggest that ridge push from the Mid-Atlantic ridge is the main stress generating mechanism, and that the scatter in the stress magnitude data in the upper parts of the crust is due to stress variations (stress concentration / stress release and refraction) associated with stick-slip motion on faults.

The line of best fit from linear regression analysis of the magnitudes of $\sigma_h$ and $\sigma_H$, determined from Leeman type overcoring, versus depth in the Fennoscandian database is shown in Figure 7.15. The trend of $\sigma_h$ with depth shown in Figure 7.15 is reproduced for comparison in Figure 7.16 and is summarised in Table 7.2.
Figure 7.14 Values of the mean principal stress determined by the Leman type overcoring methods extracted from the Fennoscandian Rock Stress Database. (Adapted from Stephansson et al., 1986).

Figure 7.15 The trend of $\sigma_\theta$ (dashed line) and $\sigma_H$ (solid line) with depth from linear regression analysis of values obtained by the Leeman type overcoring method, from the Fennoscandian Rock Stress Database. (Adapted from Stephansson et al., 1986).
A stress profile ($\sigma_h$ versus depth) has been constructed for the site of the KTB borehole (see section 3.3.9) in Germany. The magnitudes of $\sigma_h$ determined from hydraulic fracturing tests at depths down to 9 km are reported by Te Kamp et al. (1995). The KTB borehole is drilled into foliated gneisses and amphibolites of the NE side of the Bohemian Massif, and so again it should be noted that the geological setting is considerably different from that of the sediments in the North Sea. However, for the purposes of comparison, the trend of $\sigma_h$ with depth determined at the KTB borehole down to 5 km is also shown in Figure 7.16.

The stress magnitude data from the UK mainland has been compiled by Evans (1987) and is also presented in Brereton and Evans (1987). The compilation consists of $\sigma_h$ magnitudes from approximately 30 hydro-fracs and approximately 6 overcoring measurements. These measurements mostly come from crystalline and metamorphic rocks in and around the mining areas of Cornwall in SW England. Also, several data points from the East Midlands and northern England are included in the dataset. In the presentation of the trend of $\sigma_h$ magnitudes with depth in the UK, Brereton and Evans (1987) draw attention to the fact that there appears to be a stress drop at around 750m below surface which they attribute to either a decoupling horizon at about this depth or to a near surface effect of increasing horizontal to vertical stress ratio. Unfortunately, there is not enough data to determine whether this stress drop is a widespread feature of the UK or just a localised effect. The trend of $\sigma_h$ with depth, including the interpreted stress drop, from the UK data compiled by Brereton and Evans, is shown in Figure 7.16.

More recently, hydro-frac stress measurements have been made by UK Nirex at Sellafield in NW England. This data has been described and presented in Chapters 5 and 6. The linear regression line fitted to the plot of $\sigma_h$ with depth from this dataset (Figure 6.11) has been included for comparison in Figure 7.16.

_Trends of $\sigma_h$ with depth in sedimentary basins:_
Hydro-frac data from the sedimentary basins of the US Gulf Coast region has been compiled together with leak-off test data by Breckels and van Eekelen (1981) (section 4.6.4). By using the lower bound to the majority of the leak-off and hydro-frac data (as for the North Sea LOPs in section 7.2.2) a trend for $\sigma_h$ with depth is estimated. This trend has been defined by Breckels and van Eekelen (1981) as a power law curve down to 3500m depth, and then a straight line below 3500m (Figure 4.14). The trend has been reproduced here for comparison in Figure 7.16.

Breckels and van Eekelen (1981) also present leak-off test data from throughout the North Sea. However, they do not distinguish between data from different parts of the North Sea which causes a very large scatter in their dataset, some of which has probably been caused by the stress drop in the southern North Sea (section 6.2.2). This large amount of scatter and lack of control on the location of the leak-off tests prevents this dataset from being useful for estimating a trend of $\sigma_h$ with depth and it has therefore not been included. Presumably for the same reasons, Breckels and van Eekelen (1981), have included a line of best fit for the whole dataset rather than a lower bound to the majority of the dataset and note that this trend cannot be directly compared to the Gulf Coast trend in Figure 4.14.

With the exception of the work by Breckels and van Eekelen (1981), studies of stress magnitudes with depth in sedimentary basins have mostly been limited to a few boreholes.

Magnitudes of $\sigma_h$ from hydro-fracs in basins in Colorado, New Mexico, Utah Wyoming and Michigan have been compiled by McGarr and Gay (1978). The majority of these measurements are made within 500m of the surface, however, there are approximately 10 deeper measurements which stretch down to 5 km depth and define a trend with depth for $\sigma_h$ which is reproduced in Figure 7.16.

Hydro-frac stress measurements have been made in the Piceance Basin, Colorado (Warpinski et al., 1983). These measurements sample the stress in the Cretaceous sands and shales of the Mesa Verde Group between the depths of 1310 and 2470m. Although
these measurements indicated a lithological stress contrast, with lower stresses in the sandstones and shales, and higher stresses in the shales, the average trend of $\sigma_h$ with depth for the interval tested in these wells is reproduced in Figure 7.16.

Stress measurements using the hydro-frac technique have also been made in sands and shales of the East Texas Formations (Thiercelin and Plumb, 1991) at depths of between approximately 1800 and 3000 m, and have been discussed in section 3.4. Again a lithological stress contrast, with higher stresses in shales than sands was detected, and so the average trend of $\sigma_h$ with depth from these wells is reproduced in Figure 7.16 for the depth interval tested.

The Devonian sediments of the Appalachian Plateau are basinal sediments (distal turbidites and black marine shales) which have clearly undergone inversion to be at their present elevation of between 1000 and 2000m above sea level. Evans et al. (1989), also shows how these sediments lie adjacent to the Allegheny Front (this marks the limit of the fold and thrust belt which formed during the Alleghanian Orogeny in the Permian) which has clearly influenced the geomechanical history of these rocks. An extensive set of Hydro-frac stress measurements at a range of depths down to around 1000m has been made in three wells which are within one or two kilometres of each other (Evans et al., 1989). The trend of $\sigma_h$ with depth for all three wells shows a distinct stress drop at a depth of around 700m which coincides with a particular lithological unit common to all wells. The trend with depth of $\sigma_h$ in these wells displays a more complex pattern than is the case for other sedimentary basin trends shown here, and may reflect a geological history more involved in the tectonic evolution of the area than that of most sedimentary basins. Evans et al. (1989) suggest some possible explanations for the observed stress drop. These have been outlined, with reference to the similar effect observed in the North Sea, in section 7.3.3. Although Evans et al. (1989a) conclude that, the origin of the observed stress drop in the sediments of the Appalachian Plateau remains somewhat enigmatic, the trend of $\sigma_h$ with depth has been reproduced including the observed stress drop, in Figure 7.16.
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Table 7.2. Summary of $\sigma_h$ with depth datasets presented in Figure 7.16.
Figure 7.16 Comparison of the trends with depth of $\sigma_n$ described in the text and summarised in Table 7.2
7.5.2 The Trend of $\sigma_b$ with Depth in the North Sea and Other Regions: Discussion

The trend of $\sigma_b$ with depth established from the leak-off test pressure/volume plots in the central North Sea (Figure 7.1) has also been included in Figure 7.16 and the dataset from which this trend is established is summarised in Table 7.2. Although this trend is only defined by 38 data points (as opposed to the several hundred leak-off pressures obtained for the North Sea), in view of the discussion in section 7.2.4, this is considered a reasonable trend for purposes of this comparison.

Perhaps the first thing to bear in mind when considering this comparison (Figure 7.16) is that some of the trends shown come from very different geological settings. Trends from both sedimentary basins and cratonic and basement rocks are shown.

Crystalline and metamorphic rocks which form the cratons and basement rocks of continents and which are now within a few meters to a few kilometers of the surface must have undergone at least one phase of burial (or emplacement at depth in the case of rocks such as granites) and denudation. The state of stress in these rocks is likely to be at least in part a function of the complex mechanical and thermal processes which are active as these rocks are buried and uplifted. Further, the present day structural setting within the lithosphere of rocks which form the cratonic and basement portions of the lithosphere, can be considered fundamentally different to that of rocks within a sedimentary basin. Whatever plate driving forces are assumed to be responsible for the motion of the tectonic plates (see section 3.3), it is generally the cratonic and basement portions of the lithosphere which will be most directly influenced. Indeed, it has been suggested that in some cases sedimentary basins ride passively, piggyback style, on the underlying lithosphere. The presence of active tectonics in parts of the North Sea basin, and features such as borehole breakouts throughout the North Sea, as well as the conclusions in section 7.3.4 (that a tectonic stress component exists in the North Sea), show that these rocks are clearly not simply riding passively on the underlying lithosphere but are in fact, to some degree, coupled to the basement. However, the degree of coupling is a matter of debate. For the above reasons, any comparison of stress against depth between the cratonic and basement forming rocks of the continents, and the rocks which occur in sedimentary basins (which have not undergone any burial
and uplift or have only experienced a small degree of burial and mild inversion, such as some of the rocks in the southern North Sea - see section 2.3.7.3) must be tentative.

The trend with depth of $\sigma_h$ in normally pressured formations in the central North Sea (which is close, although slightly higher - Figure 7.6 - to the trend for the northern North Sea) can be compared with trends of $\sigma_h$ with depth in other regions of the world by examination of Figure 7.16.

Broadly, there is a large degree of similarity between all these trends. To a first order approximation, it can be seen that $\sigma_h$ increases with depth at a similar rate in most areas. For example, despite the very different geological environments of the North Sea and the KTB site, the trend of $\sigma_h$ with depth is similar. Obviously there are important differences which are discussed below, but the broad similarity between these trends is perhaps surprising in view of the very wide range of values of horizontal stress reported for example by Brown and Hoek (1978).

One obvious feature from this comparison is that at shallow levels, the trend of $\sigma_h$ with depth in the central North Sea is among the lowest of all the trends and that it initially increases slowly with depth. The only other comparable trend at this depth is that from the US Gulf Coast. This can be explained from a consideration of the structural setting of the shallow rocks of the central North Sea and the US Gulf Coast. In both cases the rocks are young, relatively unconsolidated, and unlikely to be coupled strongly to the basement rocks below. Rocks in such a setting are thus unlikely to be influenced greatly, if at all, by tectonic stresses. The rocks from most of the other data compilations shown in Figure 7.16 on the other hand are from basement forming and cratonic rocks, which as discussed above are much more likely be tectonically stressed.

The presence of sea water in the central North Sea will generally act to lower the values of $\sigma_h$ at shallow depths, in comparison to the values of $\sigma_h$ at the same depth on land (where depths are measured from sea surface in the North Sea and from ground level on land). Although the presence of sea water goes some way to explaining why the trends
of $\sigma_h$ with depth are low in the North Sea and US Gulf coast data at shallow levels, this effect is thought to be small (section 7.2.3).

The trend from the US Basins (McGarr and Gay, 1978) might be expected to be similar to the North Sea and the US Gulf Coast at shallow depths. However, it should be remembered that this trend is based on around 30 hydro-frac measurements at depths between 0 and 500m, and then very few data points (approximately 10) between 400m and 5 km. The trend cannot be defined accurately at depths between 400m and 5 km, and so the trend shown in Figure 7.16 is simply a straight line between the very shallow points and the few deeper ones, which possibly over estimates the level of $\sigma_h$ at intermediate depths.

A variety of trends are seen in Figure 7.16 at shallow depth from other areas. Some of these trends, most noticeably that from Fennoscandia (Stephansson et al., 1986) show values of $\sigma_h$ which exceed the vertical stress. This is in agreement with the dataset of Brown and Hoek (1978), (Figure 7.13). The mechanisms which give rise to the high horizontal stresses at shallow depths in the crust, apparent in Figure 7.13 and in some of the data trends in Figure 7.16, are poorly understood. The term “Near Surface Horizontal Stress Paradox” has been coined by Engelder (1993) to describe the apparent discrepancy between the commonly observed high horizontal stress magnitudes near surface and the widely observed occurrence of joints (tensile fractures) in near surface rocks where they do not occur at depth. In other words, if, as the observations indicate, the joints formed near the surface, how did tensile fracturing regimes (needed for their formation) originate, given that most data indicate highly compressive near surface stress regimes?

Only the trends form the US Gulf Coast and the North Sea show an increasing gradient with depth. This may be partly due to the fact that these trends are defined by many data points, which allows a curved trend to be defined. However, the effect of compaction (as described in section 7.3.1) on young sedimentary basin rocks is thought to be the main cause of the increase in $\sigma_h$ gradient with depth. This explains why such an increase is only seen in the trends from relatively young sedimentary basin.
The increasing effect of tectonic stress with depth acts to augment that of compaction. At deeper levels (2500 - 3000m) in the crust, Figure 7.16 indicates that the level of horizontal stress is higher in the North Sea than in other areas where trends extend to these depths. This is somewhat surprising, as it is expected that high horizontal stresses would generally be found in regions of active thrusting or strike-slip (such as around the site of the KTB borehole). The trend for the North Sea shown in Figure 7.16 is the highest of the trends compared in Figure 7.6. However, it is clear that most of the other trends shown in Figure 7.6 are not very much lower than the one shown in Figure 7.16. Moreover, because of the curvature of these trends which is still apparent at depth, it appears that they would all be higher than the trend for the KTB borehole for example, at depths of 4000 - 5000 m.

It has been argued in section 7.3 of this study, on the basis of the data presented, that there is an increasingly strong component of tectonic stress with depth in the central and northern North Sea and that this is largely responsible (in conjunction with the weight of the overburden) for the observed level of $\sigma_h$. However, on the basis of the comparison shown in Figure 7.16, it is perhaps surprising that seismicity in the North Sea is apparently low. There are at least two possible explanations for this:

(a) It could be that there is indeed seismicity in the North Sea but that it is generally not detected. The possibility that it is not detected may be due to the lack of local seismometers. Attenuation of the seismic waves as they propagate through thick sequences of relatively unconsolidated sediments may also hinder the detection of seismic events.

(b) It could be that the high levels of $\sigma_h$ (relative to those measured at the KTB site) seen in the deeper parts of the North Sea are not due entirely to tectonic stress (although the presence and distribution of borehole breakouts seen in the North Sea clearly indicate that tectonic stresses are distributed in the way described in section 7.3.2.3) and weight of the overburden in the context of the models of stress discussed in section 7.3.2. Processes such as creep (due for example to pressure solution), which occur over
long periods of time, and are aided by elevated temperatures, high confining pressures and mobile pore fluids, can act to decrease stress differences. In the North Sea, creep would act to increase the value of $\sigma_h$ and thus with increasing depth $\sigma_h$ would be expected to approach or even become equal to the overburden.

Although rocks in the North Sea at deeper levels (3000 m and below) are clearly still able to support shear stress, as evidenced by the fact that $\sigma_h$ is still significantly less than $\sigma_v$, and the presence of borehole breakouts, it could be that the rock is behaving as a viscoelastic material. Such a material can support shear stress over relatively short periods of time, or when the stress is re-applied continuously, such as a renewable tectonic stress (section 3.3.5) or increasing vertical stress due to increasing overburden thickness in a subsiding basin. Over long periods of time, if the stress is not continuously applied, the material acts as a viscous material and stress differences dissipate such that both horizontal stresses and the vertical stress are equal. Viscoelastic behavior of the rocks in the North Sea (where it is assumed that there are renewable sources of stress from tectonic forces and, at least in the northern and central North Sea, from increasing overburden) could therefore explain how they are able to support shear stress, and yet at the same time be susceptible to processes such as creep which may be responsible for the high levels of $\sigma_h$ at depth.

The difference in trend of $\sigma_h$ with depth between the central North Sea and the KTB site could therefore be due to the difference in rock type encountered at these two sites. The rocks of the central North Sea are of course sedimentary rocks, predominantly sands and shales which are porous and thus contain relatively large amounts of fluid. The types of rock encountered at the KTB site are hard crystalline rocks with low porosity which are generally unlikely to contain large volumes of fluid. It seems likely that creep process related to pressure solution and other diagenetic effects would be more active in the sediments of the North Sea and that this could explain the higher levels of $\sigma_h$ in the North Sea relative to the KTB site.
Chapter Eight - Conclusions

Leak-off test data, when treated carefully, can be used to estimate minimum horizontal stress magnitudes ($\sigma_h$) in the crust. This can be done in two ways:

(i) Values of $\sigma_h$ can be estimated, in many cases, from careful examination of leak-off test pressure/volume records where the leak-off tests are performed in sedimentary rocks.

(ii) The trend of $\sigma_h$ with depth can be determined, for a particular geographic domain, where pressure/volume plots are not available, but many single LOPs are available. The trend is defined as a the lower bound to the dataset (or at least 90% of the dataset) when plotted as LOP against depth.

Where there is not enough data to define the lower bound to the LOP against depth plot, it is considered that, although single LOPs can significantly overestimate the value of $\sigma_h$, the trend of LOPs with depth reflects the trend of $\sigma_h$ with depth.

Using leak-off test data in the ways outlined above, and calculated values of vertical stress where available, the following conclusions can be drawn regarding the distribution and nature of stress in the North Sea basin.

The distribution of stress in the central and northern North Sea is distinct, as a result of different geological evolution, from that in the southern North Sea.

In the central and northern North Sea, $\sigma_h$ increases with depth. The ratio of $\sigma_h/\sigma_v$ (in overpressured and normally pressured formations) also increases with depth. The level of $\sigma_h$ is higher in shales than in sands, and is higher in overpressured formations than normally pressured formations.

In the southern North Sea, $\sigma_h$ increases with depth down to the base of the Zechstein halite (approximately 8000 ft below sea level), where, at least in the area of the
Ravenspurn field, the level of $\sigma_h$ decreases sharply. The ratio of $\sigma_h/\sigma_v$ is higher than in the central and northern North Sea at shallow depths, and does not increase significantly with depth. No lithological stress contrast is apparent in the southern North Sea, and the effect of formation pore pressure on $\sigma_h$ is also less clearly seen. However, it must be remembered that there is less leak-off test data in the southern North Sea, and what data there is has been complicated by the presence of the stress drop.

The origins of the southern North Sea stress drop remain enigmatic. However, the distribution of stress in the northern and central North Sea, and in the southern North Sea above the stress drop can be explained in the context of either of the two models (UESM incorporating horizontal strain or MCFM applied to a strike slip regime) which have been widely proposed in the literature to describe the state of stress in sedimentary basins. Using reasonable values for the rock properties, both of these models require a component of tectonic stress to become increasingly important with depth in the central and northern North Sea, and to be present at significant levels at all depths in the southern North Sea. Such a distribution of tectonic stress is consistent with the observed lithological stress contrasts (or lack of them) and the distribution of borehole breakouts in the North Sea.

If it is assumed that the UESM (incorporating horizontal strain) controls the distribution of stress in the North Sea, the stress regime could generally be either normal or strike-slip. If it is assumed that the MCFM controls the distribution of stress in the North Sea, the stress regime must be generally strike-slip.

A combination of borehole breakout observations and $\sigma_h$ estimates from leak-off tests, within the framework of the Mohr-coulomb failure criterion, indicates that the stress regime in the North Sea is generally strike-slip.

A comparison of the trend of $\sigma_h$ with depth in the North Sea to trends of $\sigma_h$ with depth in other areas, has indicated that creep processes might be responsible for the high levels of $\sigma_h$ in deeper parts (3000m and below) of the North Sea.
**Recommendations for Future Work**

If the whereabouts of suitable datasets is known, further comparisons of LOPs (with pressure/volume plots) and hydro-frac derived $\sigma_h$ should be performed in order to test the relationship between LOP and $\sigma_h$ proposed in section 6.2.3. This should ideally be augmented with detailed knowledge of the type of rock in which the leak-off test was performed, and the nature of the borehole wall before and after the test. With the increasingly common use of borehole imaging logs, such information should be available. I have recently observed what appear to be leak-off test induced axial fractures on borehole imaging logs (Schlumberger GeoQuest). The lithology and nature of the borehole wall could, of course, also be determined from the same logs.

The values of $\sigma_h$ determined from leak-off tests or hydro-fracs, could be better interpreted in the context of stress models, if the actual values of parameters such as Poisson’s ratio and friction angle for the rocks in which the tests are performed were known. Ideally, these parameters would be determined in the laboratory, under in-situ conditions, on rock samples cored from the tested intervals. Previous work of this type (for example Thiercelin and Plumb, 1991), has shown that in order to determine all the necessary parameters, many careful, detailed experiments need to be performed, and it is therefore considered unlikely that this could be done for a dataset as large as that of the North Sea. However, it would be useful simply to quantify how parameters (particularly Poisson’s ratio and friction angle) typically change with increasing compaction and confining pressure. If this data were available for typical sands and shales, the amount by which $\sigma_h$ would be predicted to change with depth could be quantified and the magnitude of the tectonic component of stress could be determined. Some of the rock parameters might be obtained from geophysical logs. However, care must be taken that the relevant properties are being measured. For example, P and S wave velocities from logs can be used to determine the dynamic (undrained) Poisson’s ratio, however, it is generally the static (drained) Poisson’s ratio that is of interest in the UESM.

Information on the strain history of rocks in the North Sea would be useful (a) to quantify a reasonable value of horizontal strain to use in the UESM incorporating horizontal strain, and (b) to determine whether processes such as creep might have
played an important role in distributing the stress. Strain studies, coupled with detailed measurements of rock properties, along the lines of those performed by Evans et al. (1989), could reveal the cause of the southern North Sea stress drop.

There is certainly much room for more work to determine the magnitude of $\sigma_H$ in the North Sea. The type of approach outlined in section 7.4 seems reasonable. However, such an approach would be greatly enhanced if again, the rock properties were better constrained, preferably from laboratory experiments. More confidence could also be placed in both the type of failure, and the characteristics of the lithology in which the breakout is detected, using borehole imaging logs.

There are, perhaps, other useful methods of estimating the magnitude of $\sigma_H$. One such method has very recently arisen through discussion between myself and my supervisors at UCL. For a dataset such as those in Figures 7.2 - 7.5, an upper bound to the data can be added as well as a lower bound. It has been argued that the lower bound represents $\sigma_b$, thus an extension of this argument (section 4.6.5) is that the upper bound should be equal to the breakdown pressure given in equation 4.13: $3\sigma_b - \sigma_H - p + T$. If data from normally pressured tests is used such that $p$ will be hydrostatic, and a reasonable value for $T$ (of say 2 - 5 MPa) is assumed, the value of the bound at any point can be solved for $\sigma_H$. There has not been time to rigorously assess this method as part of this study. However, it takes only a few minutes to perform preliminary calculations on say, what Figure 7.2, where at a depth of 3500 m, $\sigma_b$ is estimated to be 65 MPa, and assuming hydrostatic pore pressure and a tensile strength of 4 MPa, the value of $\sigma_H$ yielded from the upper bound is 80 MPa. The vertical stress at this depth, based on the trends established from the Amerada Hess dataset for the central North Sea, is approximately 75 MPa. Thus the preliminary results of this method predict similar values as the method outlined in section 7.4, i.e. a strike-slip regime (with a relatively low degree of anisotropy) is indicated.

Finally, the determination of earthquake focal mechanisms for the North Sea, if indeed there are any earthquakes, would very much aid the understanding of the nature of stress in the North Sea.
Recommendations for future procedures.

Recording leak-off test data systematically and carefully, as has generally been done for the more recent leak-off tests performed by Amerada Hess, has proved useful in the interpretation of the LOP in terms of $\sigma_h$. If the leak-off test is going to be performed anyway, careful and systematic procedures and recording of data will not take up any more rig time. Further, it has been shown in section 6.2.1.2, that by slightly extending the leak-off test procedure, i.e. performing a double leak-off test, almost all such tests can be interpreted in terms of $\sigma_h$, and a lot more confidence can be placed in the interpretation. It is thought that the amount of added value obtained from such a test, would often be worth the extra 10 minutes or so of rig time.

This study, as well as previous studies, such as that of Cowgill (1994), has shown that a lot of very useful information can often be obtained from data which was collected for other purposes. Although the usefulness of such data is not immediately obvious to operators, it is considered short-sighted and wasteful to either discard or carelessly archive such information, rendering it either impossible or very difficult to access at a later data. Data management is a notorious problem in the oil industry, and the increasing volumes of data being generated by rapidly developing oilfield technologies is forcing the issue of data management to the forefront of exploration and production management’s thinking. It is recommended that the type of data used in this study is integrated into data management projects.


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Appendix.

The procedure for estimating the lower limit of $\sigma_h$ using the Mohr-Coulomb Criterion:

For simplicity, only vertical boreholes are considered and it is assumed that the principal stresses in the crust are oriented vertically and horizontally, and that the rock behaves as an isotropic linear elastic material. The stress field around the wellbore wall (in the plane perpendicular to the borehole axis) which arises where there are unequal horizontal stresses can be derived from equations 4.1 to 4.3 and 4.6 to 4.8 and are given for example by Jaeger and Cook (1979):

$$\sigma_r = \left(\frac{\sigma'_{H} + \sigma'_{h}}{2}\right) \left[1 - \frac{\alpha^2}{r^2}\right] + \left(\frac{\sigma'_{H} - \sigma'_{h}}{2}\right) \left[1 + \frac{3\alpha^4 - 4\alpha^2}{r^2}\right] \cos 2\theta + \frac{\alpha^2}{r^2} (p_w - p) \quad (A.1)$$

$$\sigma_\theta = \left(\frac{\sigma'_{H} + \sigma'_{h}}{2}\right) \left[1 + \frac{\alpha^2}{r^2}\right] - \left(\frac{\sigma'_{H} - \sigma'_{h}}{2}\right) \left[1 + \frac{3\alpha^4}{r^2}\right] \cos 2\theta - \frac{\alpha^2}{r^2} (p_w - p) \quad (A.2)$$

$$\tau_{r\theta} = -\left(\frac{\sigma'_{H} - \sigma'_{h}}{2}\right) \left[1 - \frac{3\alpha^4}{r^2} + 2\alpha^2\right] \sin 2\theta \quad (A.3)$$

The stress around the wellbore wall which acts parallel to wellbore axis ($\sigma_z$) is given by Jaeger and Cook (1979):

$$\sigma_z = \sigma'_v - 2\nu(\sigma'_{H} - \sigma'_{h})\frac{\alpha^2}{r^2} \cos 2\theta \quad (A.4)$$

Where $\sigma'_v$, $\sigma'_H$, and $\sigma'_h$ are the effective farfield vertical, maximum horizontal and minimum horizontal stresses respectively, $\nu$ is Poisson’s ratio and all other symbols are as in equations 4.1 to records 4.3 and equations 4.6 to 4.8.

As shear failure during breakout formation occurs on planes within the wall rock, away from the wellbore wall (Zoback et al., 1985) the principal stresses away from the wellbore wall must be considered. The principal stresses in the plane perpendicular to the wellbore axis (let $\sigma_z$ be the maximum principal stress and $\sigma_h$ be the minimum
principal stress in this plane) are derived from the above equations and are given by Zhou (1994) as:

\[
\sigma_\theta = \frac{(\sigma_\theta + \sigma_r)}{2} + \frac{\sqrt{(\sigma_\theta - \sigma_r)^2 + 4\tau_{\theta r}^2}}{2} \tag{A.5}
\]

\[
\sigma_\phi = \frac{(\sigma_\theta + \sigma_r)}{2} - \frac{\sqrt{(\sigma_\theta - \sigma_r)^2 + 4\tau_{\theta r}^2}}{2} \tag{A.6}
\]

In the Mohr-coulomb failure criterion, the greatest and least effective stresses (\(\sigma'\) and \(\sigma'_{\phi}\) respectively) control failure, that is, the rock will fail if:

\[
C_0 + q\sigma'_{\phi} = \sigma' \tag{A.7}
\]

Where \(C_0\) is the uniaxial compressive strength:

\[
C_0 = 2C_u\left(\sqrt{\mu^2 + 1} + \mu\right) \tag{A.8}
\]

in which \(C_u\) is the cohesion and \(\mu\) is the coefficient of internal friction.

And \(q\) is given by:

\[
q = \left(\sqrt{1 + \mu^2} + \mu\right)^2 \tag{A.9}
\]

The most common type of breakout to occur is that in which the borehole wall fails on vertical planes, such that it can be inferred that the vertical stress is the intermediate stress. The other possibility is that when the vertical stress is exceptionally high relative to the horizontal stresses, shear failure occurs on planes oriented at an angle to the borehole axis. It is not possible to confirm which type of breakout is present just dipmeter data. However, it can easily be shown from the above equations that the vertical stress at the wellbore wall (\(\sigma_z\)) is only the maximum principal stress when \(\sigma_v\) is
more than twice as large as $\sigma_h$. As all the data in this study indicate that $\sigma_h$ is considerably more than half $\sigma_v$ it can be assumed that any breakouts present have been formed under the condition that the maximum and minimum principal stresses in the vicinity of the borehole wall lie in the plane parallel to the borehole axis, i.e. the horizontal plane.

In order to estimate the minimum value $G_h$, values for the parameters in equations A.1 to A.9 must be obtained. Firstly, an interval of borehole is selected in which a breakout has been confidently identified, and a leak-off test has also been performed. The leak-off test pressure/volume record provides an estimate of $G_h$, and reasonable values for the rock properties are selected. By assuming the rock is weak, more confidence can be placed in the estimate of the lower bound of $G_h$. $\mu$ is generally in the range 0.6 to 1.0 (Byerlee, 1978). The lower of these values (i.e. 0.6) is taken as this represents a lower frictional strength. Reasonable values for the cohesion of sedimentary rocks are assumed to be in the range 10 to 50 MPa. For most of the leak-off tests from the Amerada Hess drilling, the pore pressure is known. The mud weight is also usually recorded in the leak-off test record, otherwise it can be assumed that the mud weight is slightly above the pore pressure.

A value for $G_h$ is "guessed at" and substituted into equations A.1 to A.4. The principal horizontal stresses around the wellbore can then be determined from equations A.5 and A.6 and these values in turn are used in the failure criterion that is described by equation A.7. If by using the chosen value of $G_h$ the failure criteria predicts that breakout occurs, another lower value of $G_h$ is chosen and the process is repeated until the minimum value of $G_h$ that predicts breakout will occur is obtained.

It is reasonably practical to calculate the principal stresses at the borehole wall at the azimuth of $G_h$ and apply the failure criterion to these values "by hand". However, as shear failure around the wellbore generally initially occurs at some small distance from the borehole wall, and at an azimuth not exactly parallel to either of the farfield principal stresses, the calculation becomes much more involved. Fortunately, a computer program has been published (Zhou, 1994) which performs the calculations represented by equations A.1 to A.9 and outputs the predicted shape of the breakout. Examples of the
breakout simulation output have been presented and discussed in terms of estimating the minimum magnitude of $\sigma_H$ in section 7.4.