NGU Report 98.043

Lineament architecture and fracture distrubution in metamorphic and sedimentary rocks, with application to Norway

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REPORT

Report no.: 98.043	ISSN 0800-3416	Grading: Open

Title:

Lineament architecture and fracture distribution in metamorphic and sedimentary rocks, with application to Norway

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County: Norway		Commune:	
Map-sheet name (M=1:250.000)		Map-sheet no. and -name (M=1:50.000)	
Deposit name and grid-reference:		Number of pages: 78 Price (NOK): 195,- Map enclosures:	
Fieldwork carried out:	Date of report: November 1998	Project no.: 2685.00	Person responsible: [signature]

Summary: Lineaments can be regarded as zones in the crust with enhanced fracture intensity (fracture zones) which have a distinct architecture. They can be divided into two main types: fault zones or joint zones. These structures are found both in metamorphic and sedimentary rocks, and have some common characteristics.

The fracture distribution can be ascribed to: (i) The central part of the lineament, which consists of a dense network of short fractures, secondary minerals and fault rocks (sub-zones A-B), and (ii) the marginal part (sub-zones C-D), revealing lower fracture intensity, pronounced lineament-parallel fracturing, and absence of clogging minerals. (iii) Outside this, the distal part (sub-zone E; < 250 m) exhibits relatively low fracture frequencies of variably oriented structures, grading outward into areas of (iv) general background fracturing. This lineament-parallel zonation is commonly symmetric around the central part of master joint zones and steep fault zones, whereas inclined fault zones generally have the highest strain in the hangingwall.

There are some general differences between lineaments in metamorphic and sedimentary rocks. Typically, fracture zones in metamorphic rocks show a more symmetric fracture distribution, more distinct zonation, and wider deformation zone than fracture zones in sedimentary rocks.

Fracture systems have significant influence on conductivity in both metamorphic and sedimentary rocks, hence they are important for fluid extraction from wells and leakage in tunnels. The lineament-parallel symmetry of fracture sub-zones suggests that conductivity along the fracture zone in general is uniform, whereas lineament-normal conductivity is heterogeneous and controlled by impermeable sub-zone(s).

Considerable variance in lineament architecture is demonstrated by a variety of fracture frequency profiles, fracture styles and distribution of fault rocks. Lithology, depth of deformation (PT-conditions), strain intensity and strain rate and the nature of the actual local and far-field stress systems are taken into account in a model to explain the different lineament architectures. Fracture distribution and orientation, fracture types, deformation products, and strain intensity are applied as variables, and used to determine fracture environment, stage of development and influence of strain-hardening and strain-softening. These data are applied to generate fracture profile types which are characteristic for each lineament type.

Keywords: Strukturgeologi	Grunnvann	Fagrapport
Sunnfjord		

CONTENTS

ABSTRACT

Lineaments can be regarded as zones in the crust with enhanced fracture intensity (fracture zones) which have a distinct architecture. They can be divided into two main types: fault zones or joint zones. These structures are found both in metamorphic and sedimentary rocks, and have some common characteristics.

In metamorphic rocks the spatial distribution of fractures varies in a systematic fashion. The common fracture network can be ascribed to: (i) The central part of the lineament, 0-50-m wide, which consists of a dense network of short fractures and fault rocks (sub-zones A-B), and where secondary minerals are common, sealing most fractures, and (ii) the marginal part (sub-zones C-D), commonly 20-50-m wide, revealing lower fracture intensity, pronounced lineament-parallel fracturing, and absence of clogging minerals. (iii) Outside this, the distal part (sub-zone E; < 250 m) consists of relatively low fracture frequencies of variably oriented structures, grading outward into areas of (iv) general background fracturing. The main difference between joint zones and fault zones is that the former reveal a more symmetric pattern whereas inclined fault zones (normal and reverse) tend to have asymmetric patterns with increased strain in the hangingwall.

In sedimentary rocks, the same fracture sub-zone pattern as in metamorphic rocks may be observed. There, the central fault zone (sub-zones A-B) is characterised by undeformed or fractured rock lenses separated by high-strain zones filled with fault rocks and open or mineralised fractures. Zones of enhanced fracture frequency are commonly present on both sides of the fault (sub-zones C-E), reaching fracture frequencies that may be up to two orders of magnitude greater than those recorded in the undeformed country rock. The higher fracture frequencies are usually recorded in the hangingwall fault block. The fracture frequency in most cases drops off quickly away from the fault zone, and enhanced fracture frequencies are usually not detectable more than a few tens of metres away from medium size faults, and a few hundred metres in large-scale faults.

There are some general differences between lineaments in metamorphic and sedimentary rocks. Typically, fracture zones in metamorphic rocks show a more symmetric fracture distribution, more distinct zonation, and wider deformation zone than fracture zones in sedimentary rocks.

Fracture systems have significant influence on conductivity in both metamorphic and sedimentary rocks, hence they are important for fluid extraction from wells and leakage in tunnels. In the case of metamorphic rocks, lineament-perpendicular conductivity in fracture zones can be regarded as potential low in the central part, high in the marginal part (especially zone C), and intermediate in distal parts. In sedimentary rocks, small-scale faults (deformation bands in siliciclastic sediments) and fault rocks within fault zones frequently suppress permeability with two orders of magnitude, and up to five orders of magnitude in some cases.

The lineament-parallel symmetry of fracture sub-zones suggests that conductivity along the fracture zone in general is uniform.

Considerable variance in lineament architecture is demonstrated by a variety of fracture frequency profiles, fracture styles and distribution of fault rocks. Several parameters vary in the development of lineaments (master joint and fault zones), including lithology, depth of deformation (PT-conditions), strain intensity and strain rate and the nature of the actual local and far field stress systems. These influences are taken into account in a model to explain the different lineament architectures. Fracture distribution and orientation, fracture types, deformation products, and strain intensity are applied as variables, and used to determine fracture environment, stage of development and influence of strain-hardening and strain-softening. These data are applied to generate fracture profile types which are characteristic for each lineament type.
It is hoped that the resulting model may increase the ability of predicting and evaluate fracture distribution and potential of fluid flow in lineaments.

1. INTRODUCTION

The understanding of fracture systems is of great importance in assessing present and ancient stresses in the Earth's crust, and also represents the basis for general structural geological analysis in the brittle regime. Furthermore, regional fracture systems are frequently crucial in applied geology and engineering geology because of their influence on rock stability and fluid flow.

Regional fracture studies commonly utilise data obtained by remote sensing, applying active or passive sensors carried by aircraft or satellite. In the analysis of such data, the geometry and position of assumed fractures are inferred, but the character of the structure(s) in principle cannot be determined with any certainty without field control.

1.1 Definitions

To keep the necessary level of precision in regional fracture analysis, it is absolutely crucial to apply terms for which precise definitions are given. In the present study, therefore, following the definitions given by O'Leary et al. (1976), Bates & Jackson (1987), Nystuen (1989) and Gabrielsen & Aarland (1995), we apply the following definitions, given in the sequence of increasing precision:

A lineament is a linear or curvilinear feature that is directly visible on the Earth's surface, or stands apart on either topographical or geophysical maps, or on satellite or aerial photographs. Lineaments are assumed to reflect a zone of weakness or geological inhomogeneity in the bedrock, such as a fracture, rock boundary, fold (trace-) linear rock body or ore bodies. Lines of intersection between the surface and foliations are not looked upon as lineaments (Fig. 1).

A fracture is a planar or subplanar inhomogeneity (in a rock) generated by external or internal stresses (Gabrielsen & Aarland 1995). This encompasses all types of mode I (extension) fractures such as joints and open and mineral-filled fissures (veins), mode II (shear) fractures and mode III (hybrid) fractures (Hancock 1985).

A *fracture (or lineament) population* is any arbitrary group of fractures within an (geographically restricted) area. This may include *fracture (lineament) zones* and *fracture complexes*, and can be subdivided into sub-populations which may consist of *fracture (lineament) sets* and *fracture (lineament) systems* (Gabrielsen et al. 1984) (Fig. 2). The term *fracture zone* is applied to a lineament (macro-scale) about which knowledge regarding displacement is missing, so that it cannot be assigned to one of the more precisely defined structures: *fault zone* or *(master) joint zone*. For similar meso-scale structures, we introduce the term *fracture trains*, which is defined as narrow zones with a high-frequency of subparallel fractures, and *fracture swarms*, which is high-frequency networks of fractures.

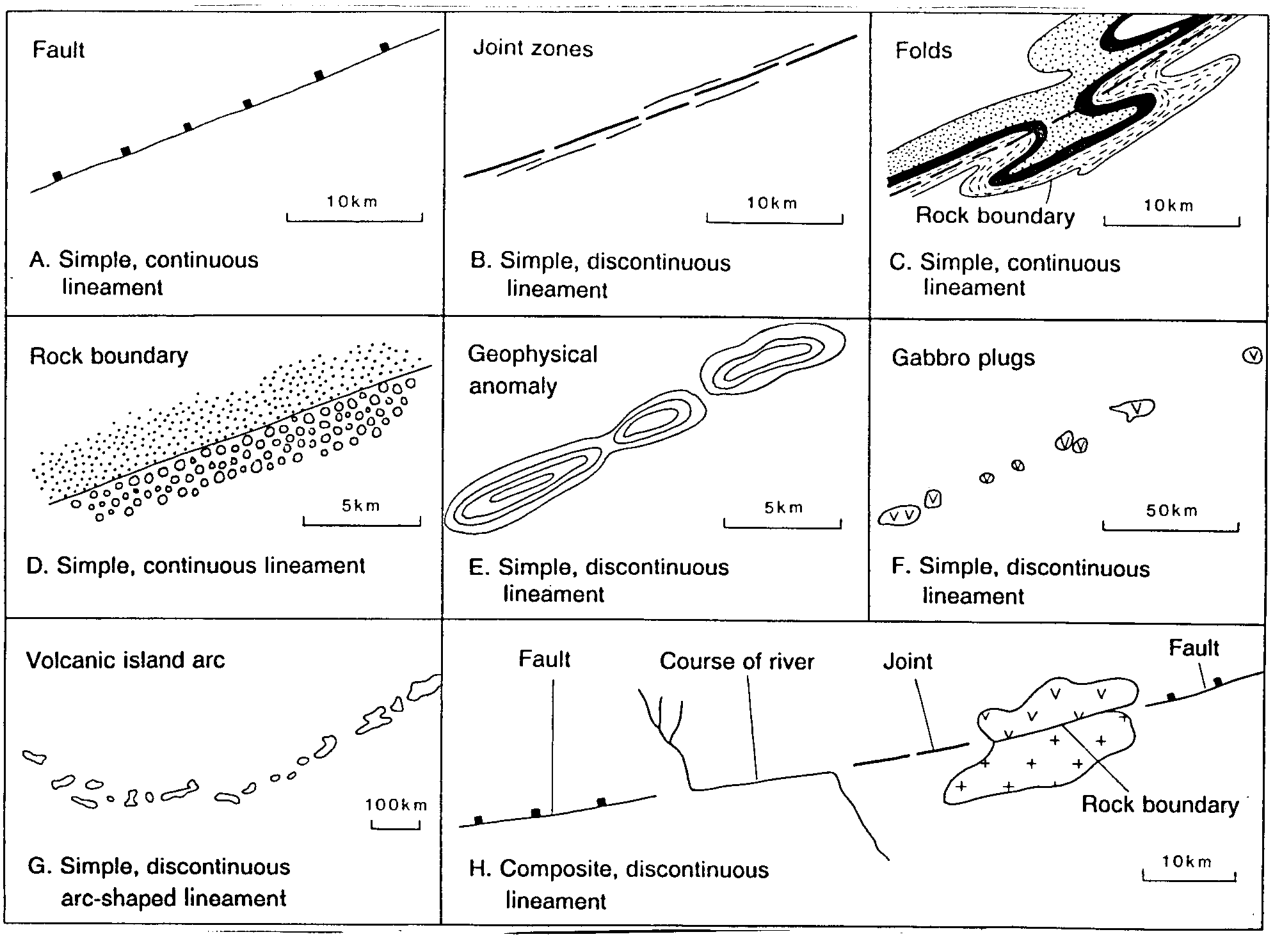

Figure 1. Lineament classification diagram. From Nystuen (1989).

When applying the definitions given above, it becomes obvious that the term "lineament" should denote linear features detected from remote sensing, and which are *believed* to be an inhomogeneity of one of the types mentioned in the definition. As soon as the nature of the lineament has been confirmed through field investigations, the term should be changed accordingly, to give the reader maximum information on the nature of the lineament.

1.2 Methods in mapping of lineaments and regional fracture systems

Although lineaments frequently stand out clearly in satellite imagery, aerial photographs and geophysical maps, the detailed interpretation of such features may be subtle. Recent remote sensing technology, data processing and imaging techniques give options for high resolution and high detectability, whereas recognisability (for definition of these terms, see Sabins 1978, pp 11-12) of structural features depends principally on field control. Also, it is obvious that resolution and detectability depend on the data type used in the remote sensing study, implying that all types of available data sets need to be scrutinised. This implies that, whenever possible, potential field data (gravimetric, magnetic, VLF, EM), passive (e.g. photogrammetric and/or scanning) and active (radar) sensors, carried by high altitude (satellites) or low-altitude (airborne) should be applied in concert.

Generally, lineament data can be considered either from a statistical point of view, where the entire lineament population is analysed, or as a data set that can be applied in the analysis of individual fracture zones or faults. The present study applies the latter approach, and therefore no further treatment of methods in statistical lineament analysis will be given.
Assuming that facilities for interactive image processing are available, analysis of single lineaments includes the following steps:
1) Identification of lineament in (high altitude) satellite imagery (commonly Landsat or SPOT; standard image processing).
2) Analysis of geometric and reflective properties of the lineament, preferentially on specially processed and magnified images.
3) Mapping of lineament on aerial photographs. If available, more than one scale should be applied (e.g. 1:25.000 and 1:10.000). Stereographic methods should be applied whenever possible.
4) Correlation of a map of the actual lineament with topographic (preferentially digital) data.
Selection of field stations, followed by field study, including detailed mapping of the internal architecture of the lineament, fracture frequency mapping, registration of strain-markers, and sampling.
6) Petrolographical/structural analysis of fault rocks and fracture surfaces.
(7) The final step includes updating of the lineament/fault map.

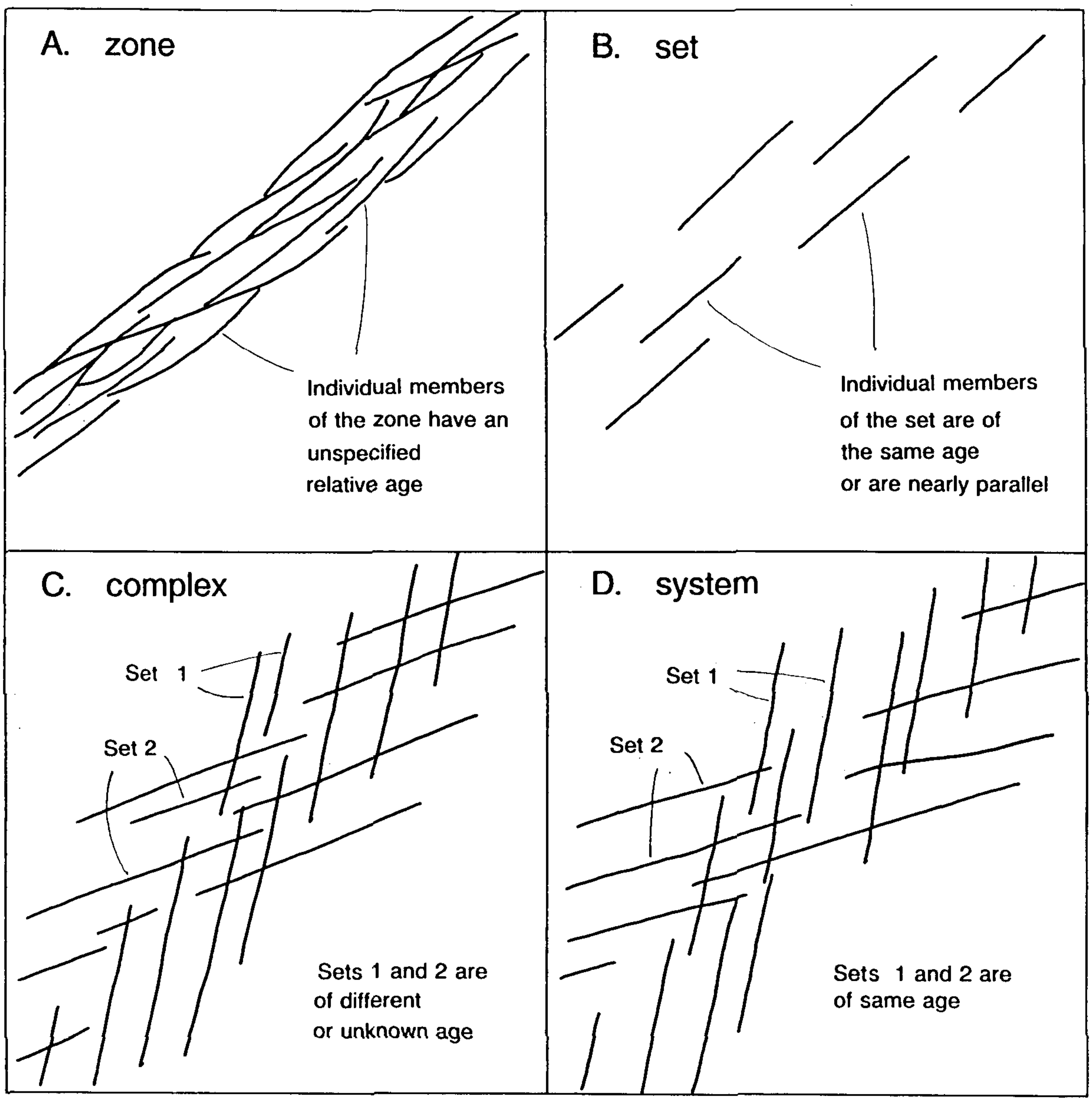

Figure 2. *A fracture population may include fracture (lineament) zones and fracture complexes, and can be subdivided into sub-populations which may consist of fracture (lineament) sets and fracture (lineament) systems (Gabrielsen et al. 1984; Nystuen 1989).*

This procedure has been followed in several recent lineament studies, covering metamorphic terranes as well as unmetamorphosed sedimentary basins. It is the intention of the present contribution to report on general findings of these studies, and to establish working procedures for future lineament field investigations.

1.3 Previous studies

Lineament studies have been performed in a number of geological terranes over a long period, applying the procedures listed above whenever possible. Major observations from some of the study areas will be summarised in the following.

Many of the lineament studies so far published in Norway have been concentrated on the statistical approach, applying satellite imagery (Ramberg et al. 1977; Gabrielsen & Ramberg 1979a,b,c; Gabrielsen 1981a,b; Gabrielsen et al. 1980; Rindstad & Grønlie 1987; Rindstad et al. 1987, Rueslåtten et al.1996) or combinations of satellite imagery and lineament data obtained from analysis of potential field data (Rathore & Hospers 1986, Karpuz et al. 1993, 1995a,b). Altogether, this has contributed to a large data base on lineaments from different lithologies, scales and tectonic settings.

The importance of field control in such studies is obvious, and although such data have been gathered systematically in several recent studies (Gabrielsen 1979-81, unpublished; Pilskog Øvrelid 1994; Odling 1994; Karpuz et al. 1995a,b; Aamodt 1997; Bråstein 1997), such data are still rare in the published literature. Exceptions to this are the recent works by Karpuz et al. (1995a,b).

The application of such analyses range from recent and palaeostress investigations (Fossen et al. 1997; in press; Reemst et al. 1996; 1997; Braathen & Henriksen 1997; Braathen 1998), to several areas of applied geology, such as hydrogeology (Rohr-Torp 1994; Banks et al. 1992; 1995; Henriksen 1996) and engineering geology (Fossen & Gabrielsen 1997a; Fossen et al. 1997a,b; Blikra & Anda 1997).

2. FRACTURE DISTRIBUTION AND ARCHITECTURE OF LINEAMENTS IN METAMORPHIC TERRANES

2.1 Introduction

Studies of lineaments in metamorphic terranes (Ramberg et al. 1976; Gabrielsen & Ramberg 1979; Pilskog Øverlid 1994; Aamodt 1997; Berg et al. 1997; Braathen et al. 1997; Gabrielsen et al. 1997b) frequently reveal an architecture (=spatial fracture pattern and zonation) that has some striking similarities to that seen in larger extensional faults in non-metamorphic sedimentary rocks. Gabrielsen (1981 unpublished) using fracture frequency distribution plots, logged transverse sections normal to the strike of lineaments in Nordland and Troms, defined four 'zones' in lineaments that can be demonstrated in the field to be faults with significant displacement (Fig. 3). A similar pattern is achieved for lineaments in southwestern Norway (Aamodt 1997), as well as from the following case study from the Sunnfjord region of western Norway. We therefore believe that a general, systematic fracture pattern/zonation may be applied to lineaments in all types of metamorphic and crystalline rocks, of Norway, as well as on a global scale.

Lineaments in metamorphic terranes may be divided into steep or inclined fault and joint zones. In most of the cases, and especially for major ('mature') fault zones, the lineaments have a more or less symmetrical architecture, in which distinct parts of decreasing fracture frequency (*zones B-E*) border a central high-strain part (*zone A*), which is the site of different types of fault rocks (Fig. 3). A similar, general pattern is also evident for more 'immature' lineaments (minor fault zones and joint zones), where the *central part* is equivalent to zones A and B, and the *marginal part* zones C and D. In addition, a *distal part,* zone E, is occasionally identified. The enhanced fracture frequency encountered in zones A-E is generally superimposed on a *background fracture system* (Gabrielsen et al. 1998). It is noted, however, that the central part (zones A and B) can be lacking, probably reflecting a less intense mode of deformation (i.e. more mature vs. immature fault zones). This general architecture is further discussed in section 4.

2.2 Methodology

Fracture data have been recorded from approximately a hundred traverses normal to lineaments, in various metamorphic and crystalline rocks, and in different parts of Norway (Fig. 4a). In order to avoid significant influence from other geological factors, such as lithological contacts, which may have a marked mechanical contrast and, which accordingly are likely to focus stresses, most traverses were located within relatively homogeneous rock units.

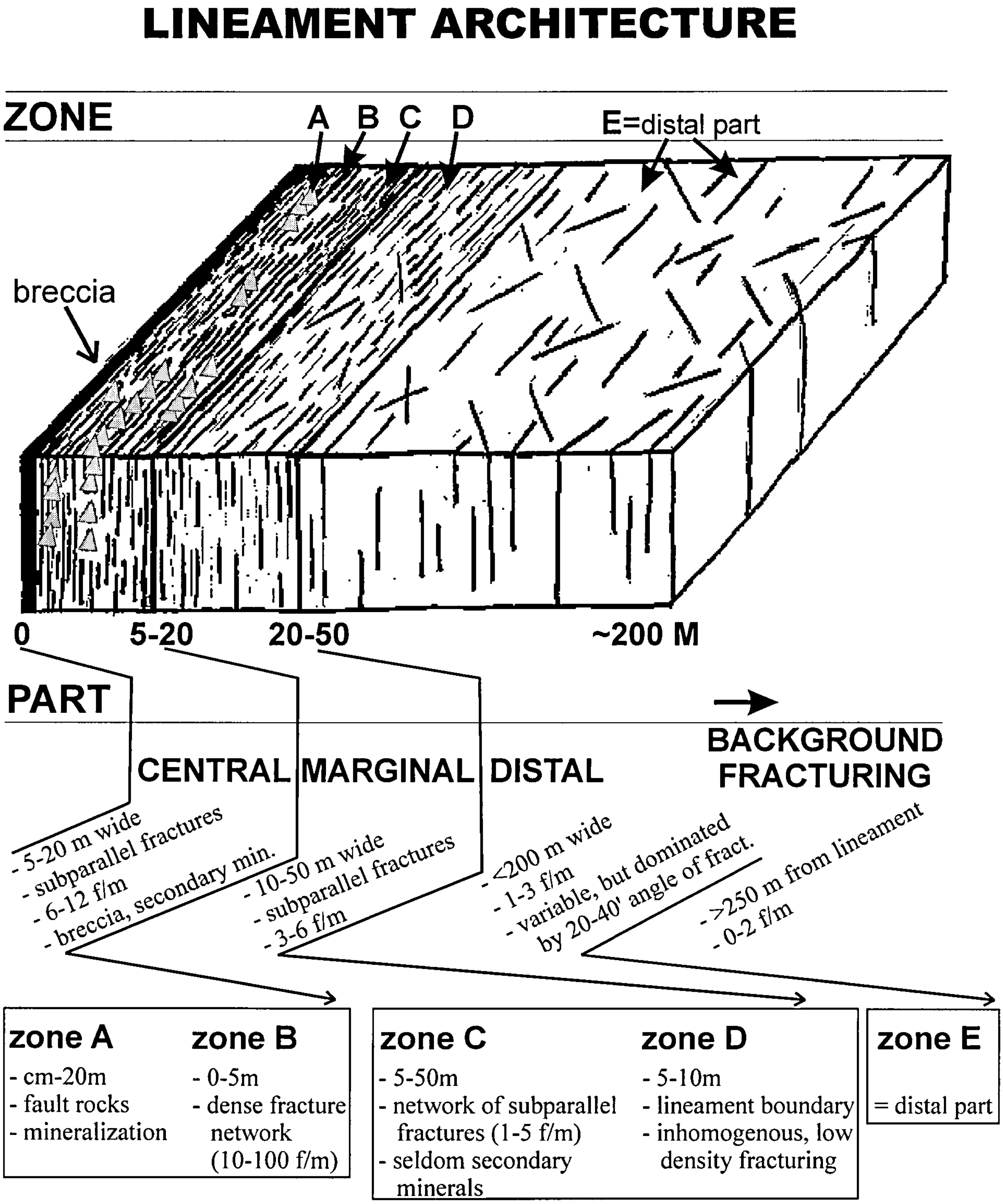

Figure 3. Architecture of lineaments. See text for further explanation.

The length of the traverses varied between 50 and >2000 metres. The longer traverses were selected to ascertain that the background fracture frequency (Gabrielsen et al. 1998) was known in each case.

Two methods (Goldstein & Marshak 1988) were applied during recording along the fracture-profiles: (i) The *fixed circle method* implies that all fractures at given stations, commonly every 5 metres, along a traverse line are recorded. Counting and measuring are limited to fractures bisecting a fixed circle (diameter = 80 cm). Consequently, only fractures longer than c. 20 cm are included, and most registered fractures are longer than 70-80 cm. (ii) The *line-intersection method* implies that all fractures intersecting the traverse line within one meter intervals are counted and measured. Recording intervals in many cases vary along the traverses, depending on the variation in fracture intensity. The general rule is that high fracture-frequency parts are more densely recorded than low-frequency parts. With this method fractures longer than c. 5 cm will be included.

Frequencies in the line-intersection method commonly fall in the range of 0-2 fractures per meter (f/m) for the low-strain parts, which is similar to the fixed circle method. However, in high-strain parts frequencies as high as 20 f/m (locally > 100 f/m) can occur for the line-intersection method, while frequencies of 12 f/m seem to be the upper limit for the fixed circle method. For both methods, fractures subparallel to the traverse line (especially subhorizontal fractures) will be depressed in the data set due to the low rate of intersection along the line. This effect is more severe for the line-intersection than the fixed circle method.

2.3 Case study: The Sunnfjord region, western Norway

Rocks of the Sunnfjord region of western Norway (Fig. 4a,b) may conveniently be divided into two main units; the so-called upper and lower plates, which are separated by a low-angle, west-dipping mylonite zone, the Nordfjord-Sogn Detachment (e.g., Norton 1986; Andersen & Jamtveit 1990; Wilks & Cuthbert 1994). This fault zone consists of greenstone and greenschist, quartzo-feldspathic schist and orthogneisses, mylonitized to various degrees. The lower plate comprises various Precambrian orthogneisses with inclusions of supracrustal rocks (mica-schists, psammites and quartzites; Griffin et al. 1985, Milnes et al. 1997). Rocks of the upper plate, mainly consisting of Caledonian allochthonous units, are more diverse, including high-grade basement units with overlying low-grade sedimentary cover, and dismembered ophiolite units (meta-gabbro, greenstone, greenschist) with overlying metasediments (Andersen et al. 1990, Furnes et al. 1990). The tectono-stratigraphic sequence is topped by Lower-Middle Devonian coarse-clastic deposits (e.g. Steel et al. 1985), eroded from the upper plate rocks, and metamorphosed to a verylow grade. These deposits occur in four E-W oriented structural depressions or synclines; structural features that are partly modified by steep, east-west brittle faults (e.g. Andersen & Jamtveit 1990, Torsvik et al. 1997, Braathen and Henriksen 1997, Braathen 1998). In most cases these faults form the boundary between the upper plate rocks and the underlying Nordfjord-Sogn Detachment.

The E-W faults form one distinct set of lineaments that can be identified on satellite images and aerial photographs (see Fig. 4b for classification). Another pronounced set is defined by N-S lineaments with subordinate NW-SE and NE-SW structures. In contrast to the E-W faults, the N-S lineaments do not displace lithological markers to any extent. Thus, they are master joint zones or immature fault zones, similar to the most common type of lineaments in Norwegian bedrock.

In the Sunnfjord region, fracture data have been collected along forty traverses normal to lineaments and in various rocks (located in Fig. 4b); twenty-two traverses targeted on N-S lineaments and eighteen on E-W lineaments/faults. The length of the traverses varies from 100 metres to 2000 meters. The fixed circle method (see above) was employed in data collection, and has been performed on ca. 3500 sites. A total of ca. 6000 fractures have been measured.

In addition to the above mentioned traverses, seven short line-intersection fracture profiles (< 20 metres) across exposed central parts were obtained. These high-resolution profiles document the complex, delicate structure of the central part of lineaments.

2.3.1 N-S lineaments

The fracture traverses normal to N-S lineaments, illustrated by examples in Fig. 5, show an increase in the fracture frequency towards the lineaments, in nearly all cases. Typically, frequencies of 0-3 f/m predominate at some distance, whereas frequencies exceeding 3-4 f/m predominate close to the lineaments (Figs. 5a, 5b and 5c). Near the core of the lineaments, high frequency zones commonly are encountered. However, local high-frequency zones also occur in distal areas of most traverses as well. Reviewing the traverses one by one, it may be concluded that there is no distinct boundary between marginal and distal parts. In most cases the increase in frequency is gradational. For lineaments which are shorter than 50 km, a significant increase is usually recorded between 100 and 150 metres away from the lineament (Figs. 5a and 5b; classification in Fig. 4b). In association with lineaments that are distinctly longer than 50 km, the boundary between the marginal part (zone E) and the background fracture system appears at a distance between 200 and 300 metres.

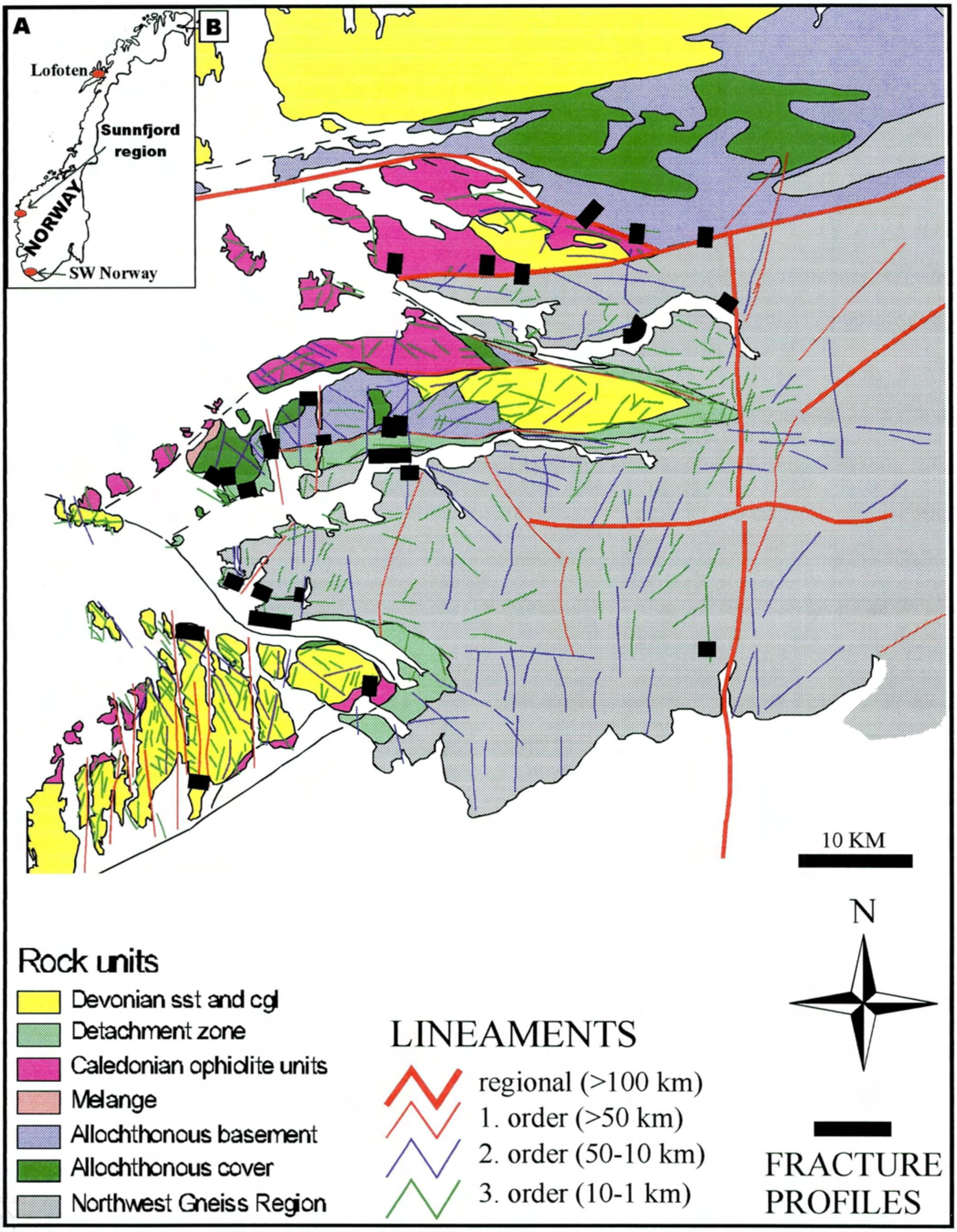

Figure 4. (A) Outline map of Norway, locating fracture-study areas in metamorphic bedrock. (B) Bedrock map of the Sunnfjord region, modified from Osmundsen & Andersen (1994). Lineaments are interpreted from a Landsat TM scene, band 5, black-white photograph in scale 1: 100 000, and classified with basis in their length. The traverses where fracture data have been recorded are located on the map.

There are cases where lithological contacts occur along the traverse, although most profiles were restricted to one rock unit. Where such boundaries occur, and especially where the mechanical properties of the various rocks are significantly different, an enhanced fracture frequency is commonly recorded (Fig. 5d). In the traverse shown in Fig. 5d, the dominant orthogneiss contains pockets/lenses of mica-schist between 700 and 1200 meters away from the core of the lineament. The same stretch reveals the typical low frequencies of 0-3 f/m, and additional peaks of 3-6 f/m, at some sites as high as 8-10 f/m. However, excluding these parts of the traverse, which are characterised by shear along contacts between different lithological units, the frequencies increase systematically towards the lineament.

In order to establish a general pattern for the fracture distribution, the entire data-set from traverses perpendicular on N-S lineaments have been compiled into one plot (Fig. 6a). This plot illustrates the strain intensity with respect to the distance from the lineament. A distinct break occurs in a distance of 250 to 300 metres from lineaments. The more distal parts reveal typical frequencies of 0-2 f/m, whereas proximal and central parts have an extensive spread, including sites with 0-2 f/m, but more typically reaching 3-5 f/m. Frequencies exceeding 7 f/m are obtained in the most proximal part.

Fracture frequencies have been calculated as a running average of ten sites sorted by distance (Fig. 6b; see caption for methodology), which depress the local effects of each traverse. This further emphasise the systematic distribution of fractures perpendicular to lineaments. Again, a boundary can be identified approximately 250 to 300 metres away from the lineaments, separating the distal part with frequencies between 0,5 and 1,5 f/m from the marginal part with values from 1,5 to 2,5 f/m. This type of diagram provides additional information pertaining the central part of the lineament (<50 metres away), which has frequently been mapped in more detail (zones A-B-C). This part exclusively exhibits average frequencies reaching 3,5 to 4 f/m. Another important aspect of the central zone is the different deformation style, with abundance of clay minerals, zeolite and breccias. The latter occur as small pods and discontinuous bands along smaller lineaments, whereas larger lineaments in many cases have enduring breccias exceeding 5 metres in width. Secondary minerals occur as fill along dense, short fractures in complex, lineament-parallel networks, sealing most small structures.

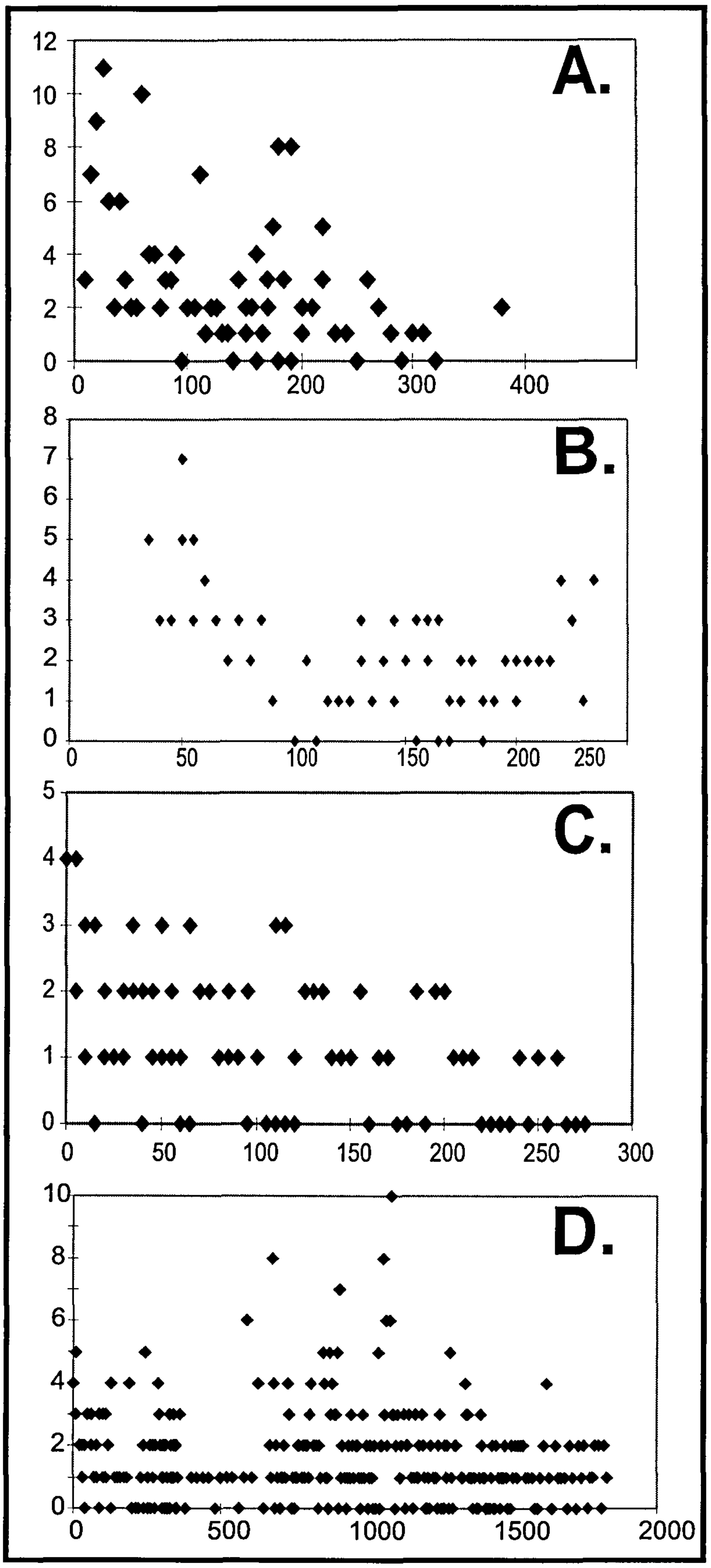

Figure 5. *Examples of distance from lineament versus fracture-frequency plots, based on the fixed circle method. The data are recorded in traverses normal to N-S lineaments. See text for further explanation.*

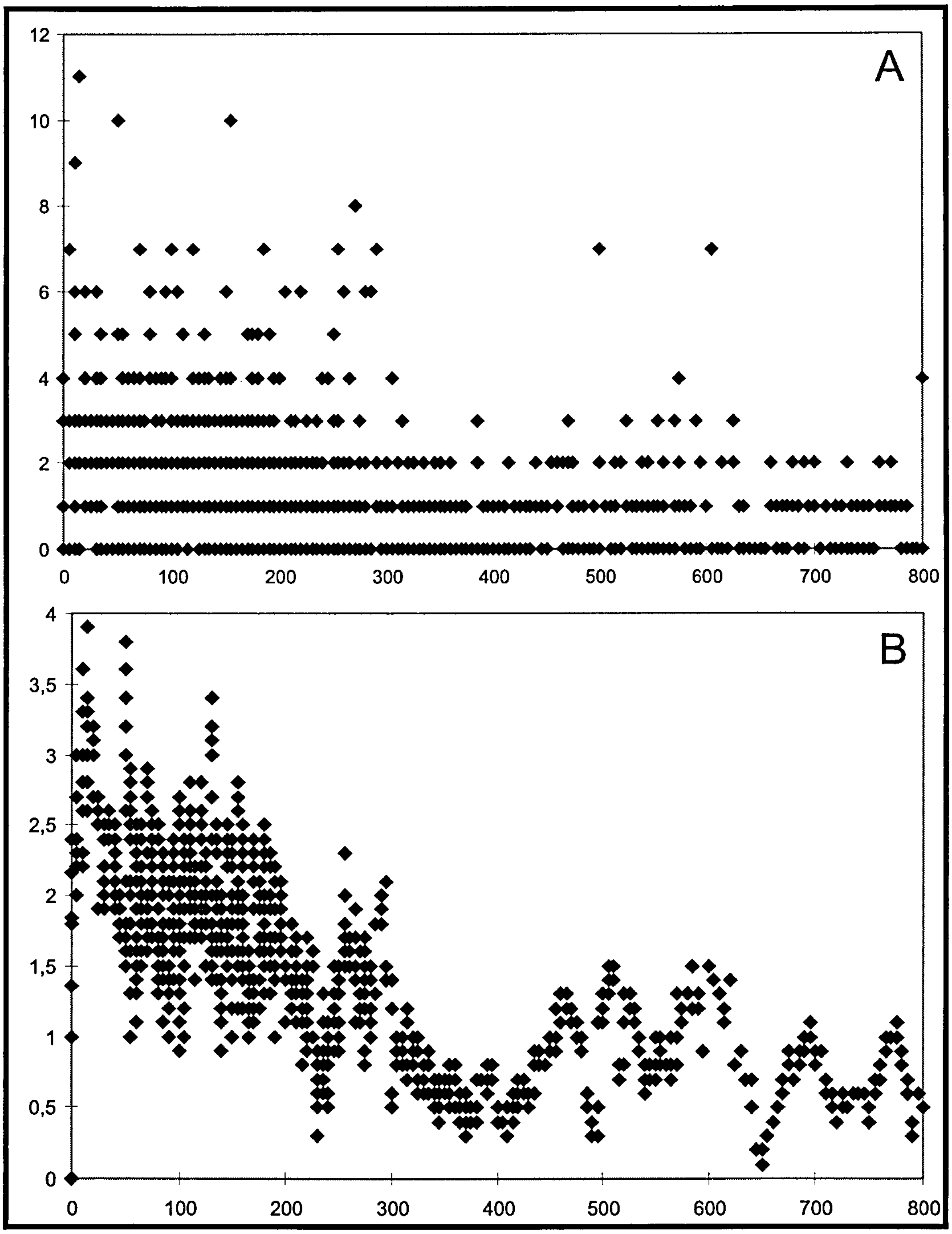

Figure 6. *Distance-frequency plots for the entire data set on N-S lineaments. (A) Distance-frequency for each site, and (B) distance-frequency as a running average of 10 sites sorted by distance. This implies that all sites have been sorted, in the spreadsheet, by distance to the centre of the lineament. Then a calculation of the average frequency of ten neighbouring sites is performed for each row throughout the spreadsheet. These average values are plotted.*

Analyses of the orientation of fractures suggest that strike vary in a systematic fashion around lineaments. In the Sunnfjord region most fractures have steep dips. Generally, the strikes plot within an angle of 60° from the N-S lineaments, i.e., the dominant strike of fractures is between NE and NW (Fig. 7a). Within 200-250 metres of the lineaments there is a tendency towards smaller angles (0-40°) between subordinate fractures and the regional lineament trend. This is clearly expressed in the central part (zones A-B-C)(0-50 metres; Fig. 7b), where most fractures plot within 50° of the lineament, and where the inner 25 metres (zones A-B) are dominated by angles between 0-35°. Applying a logarithmic scale, a distinction can be made for the inner part (Fig. 7c), which commonly displays angles in the range of 0-10°, i.e., fractures that are subparallel to the lineament are dominant.

A highly generalised result is achieved when the angle between the strike of the fractures and the lineament is plotted as the running average of ten fractures sorted by distance (Fig. 7d). In this case, common orientations between 200 and 300 metres are within 10-45°. From 200 to c. 25 metres, fractures with strike angles deviating less than 10° from the trend of the main lineament are dominant. There are also a tendency towards higher angles. The inner 25 metres show both very low angles, and a concentration in the range of 5-30°.

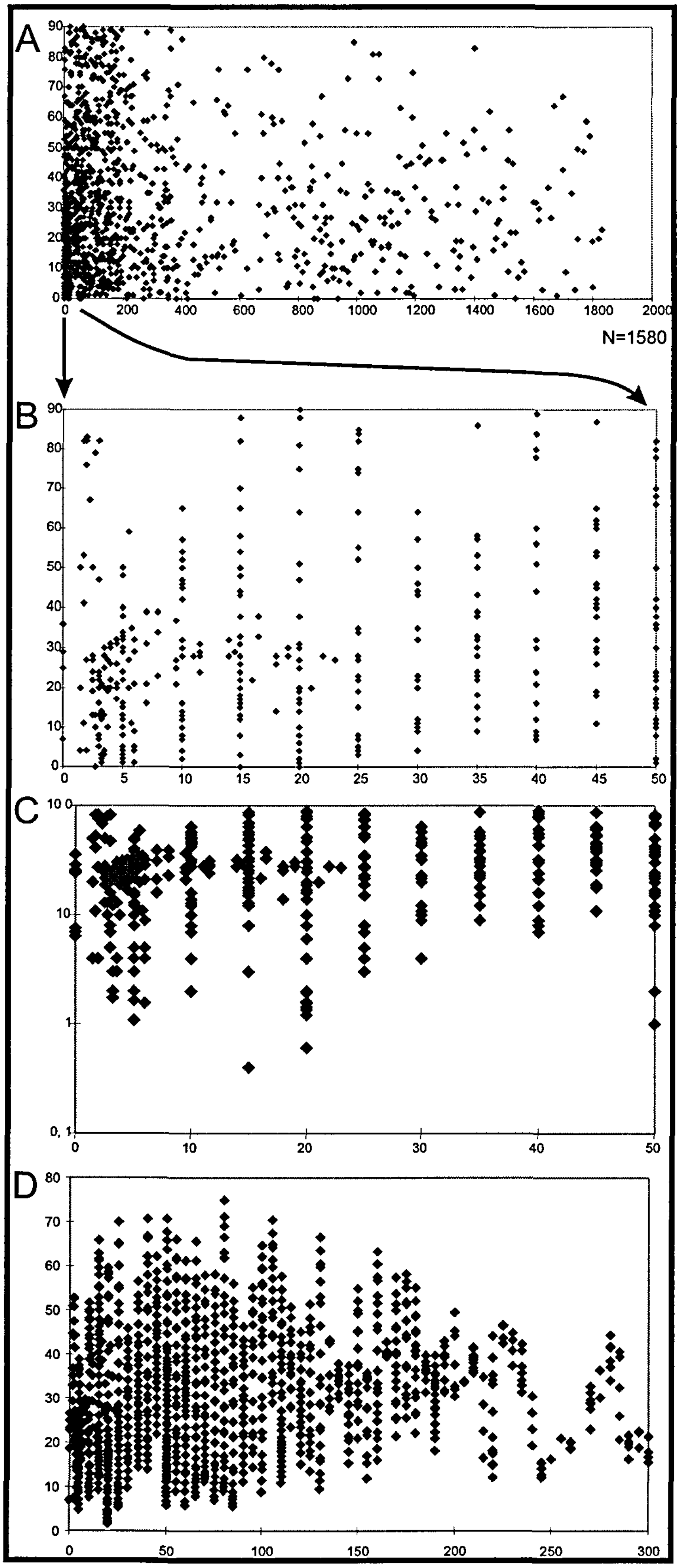

Figure 7. *Plot showing the distance to lineament versus angle between each fracture and the lineament. (A) Entire data set on N-S lineaments, (B) all data within 50 metres of the lineament, (C) all data within 50 metres, with the angle plotted in a logarithmic scale, and (D) all data within 300 metres, where the angle is calculated as the running average of 10 angles sorted by distance.*

2.3.2 E-W lineaments/faults

The eighteen traverses targeted on E-W lineaments relate to major normal faults (see above) dipping moderately to steeply to the north (Fig. 4b; Braathen 1998). Along these major structures thick fault breccias (5-50 metres) are common. Such fault rocks have yielded Late Permian and Late Jurassic-Early Cretaceous ages (Torsvik et al. 1992, Eide et al. 1997). The majority of fractures have been recorded in the hangingwall of these faults, where the most continuous outcrop-sections are evident. The footwall typically consists of less exposed, soft schists of the Nordfjord-Sogn Detachment or Caledonian nappes. The distribution of exposures to some degree limits the possibility to establish the asymmetry of the structures. Other factors of uncertainty are the different rocks in the foot- and hangingwalls, and the variable length of the profiles.

The eighteen traverses differ both in length and in the character of fracturing. In the examples of Fig. 8, no distinct general frequency distribution can be seen for two of the respective traverse plots (Fig. 8a,c). The other two plots, on the other hand, show an increase in frequency towards the lineament (Fig. 8b,d).

Results from the detailed analysis of the central part of lineaments (Fig. 8e,f) show that breccia (zone A) is common. The fault rock is bound by a dense network of short fractures (zone B) with typical frequencies of 10 to 30 f/m, locally exceeding 100 f/m. Such extreme frequencies seem to be controlled by the mechanical properties of the rock, e.g., quartzite has significantly higher frequencies than schist.

When the entire data set based on the fixed circle method is plotted, a clearer pattern is emerging (Fig. 9a). In the hangingwall, the frequency increases from 0-3 f/m in distances exceeding 300 metres, to 1-4 f/m inside 300 metres. In addition, the inner 100 metres have local frequencies ranging between 4-7 f/m. Between this inner part of the hangingwall and the footwall there is a clear break, and frequencies of the footwall drop to the range of 0-2 f/m, i.e. the typical level of the background fracture system in rocks of the Sunnfjord region. The running average of five measurements further emphases this pattern (Fig. 9b). Distal parts reveal frequencies of 0-1 f/m whereas marginal and central parts typically have 1.2-3 f/m, locally reaching 3.5 f/m.

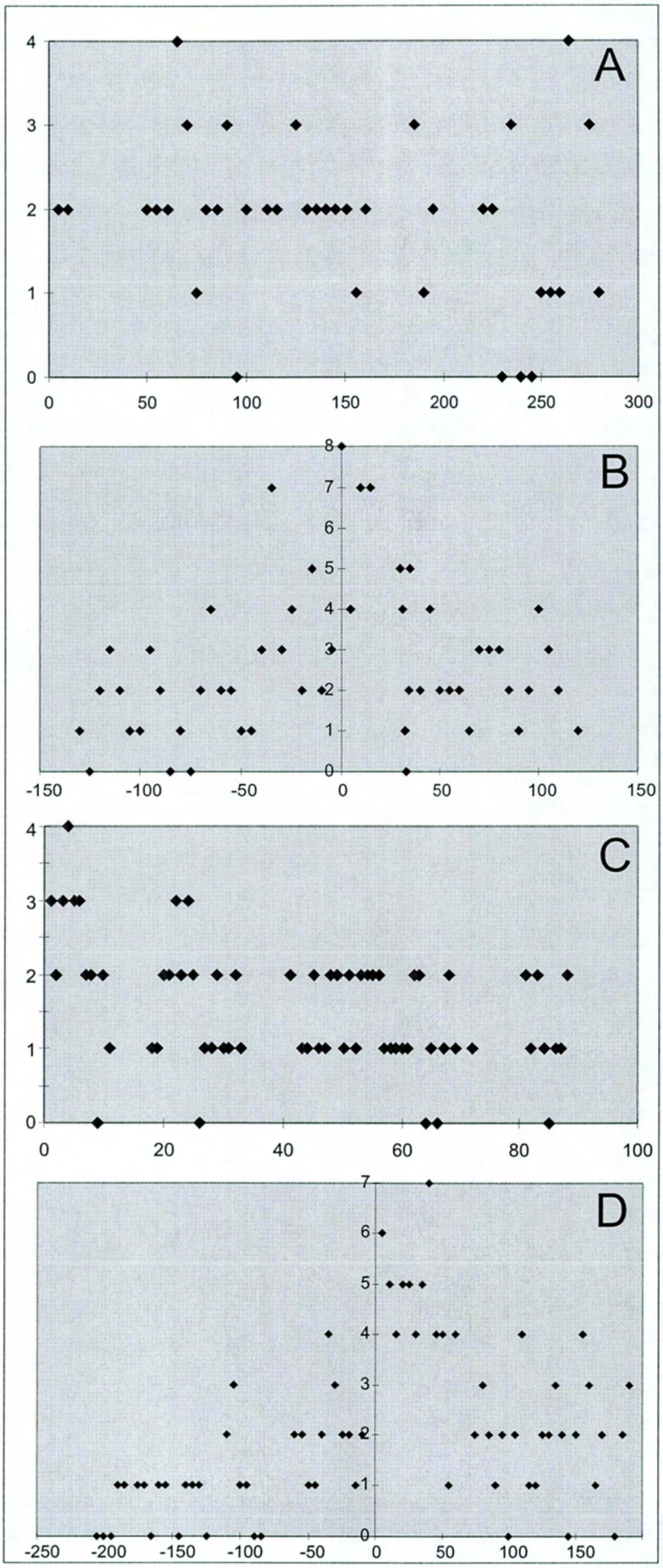
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Figure 8A-D

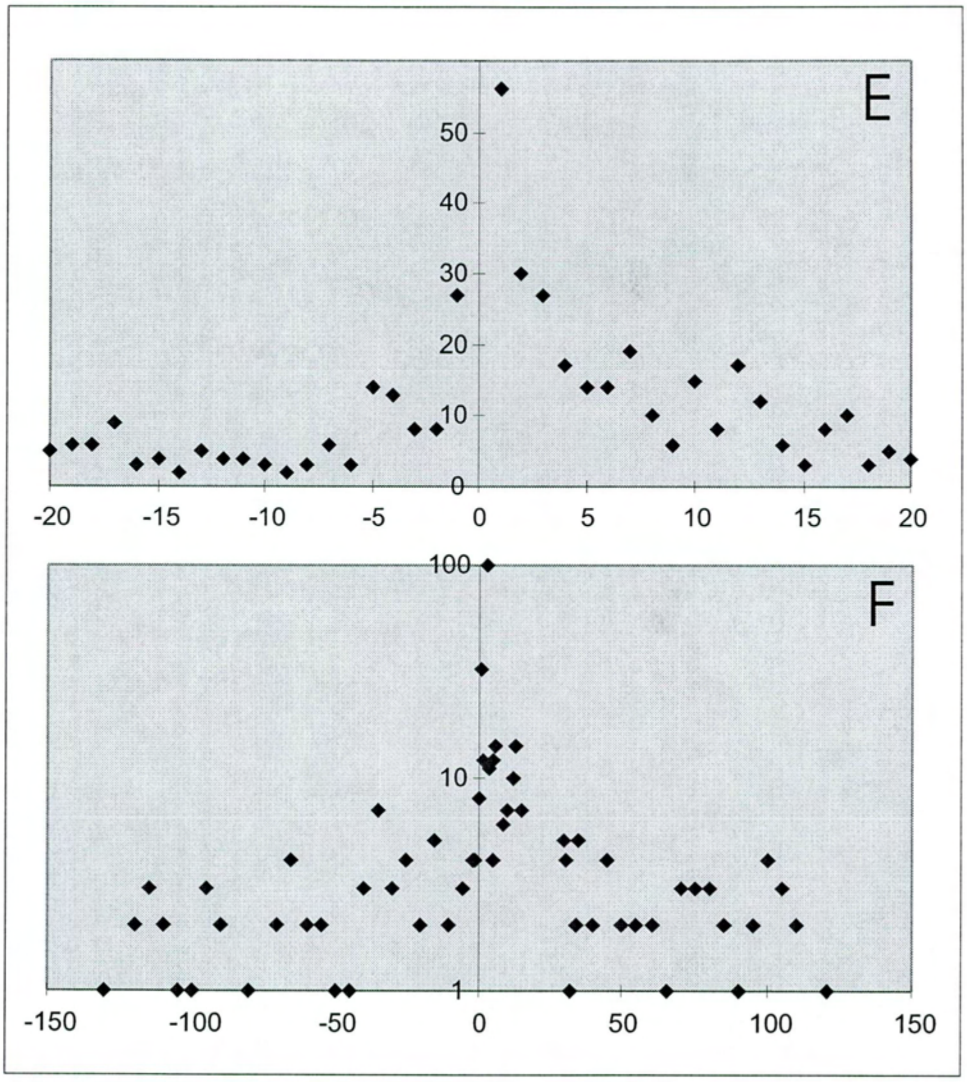


Figure 8. *Examples of distance (in metres) from lineament versus fracture-frequency plots, based on the fixed circle method, and in two plots combined with the line-intersection method (E-F). The data are recorded in traverses normal to E-W faults/lineaments. Negative distance numbers are in the footwall. (A) Standal, (B) Alnakken, (C) Endestad and (D) Hyllset. In the case of (E) Kleppen and (F) Alnakken the line-intersection method has been applied to the inner 10 metres (-5 to 5 m). Note that the Alnakken plot (F) has a logarithmic vertical scale. See text for further explanation.*

A wide distribution occurs when the angle between the lineaments and the strike of the associated fractures is analysed (Fig. 9c).There is, however, a tendency towards lower angles in the inner 150 metres, where most angles smaller than 10-15° occur (Fig. 9d).

In *summary*, the study from the Sunnfjord region embraces two types of lineaments: steep immature fault zones and inclined mature fault zones. Their main difference is that steep fault (and/or joint) zones have a symmetric fracture distribution whereas inclined fault zones show increased fracturing in the hangingwall. Characteristic for both types is the style of deformation, which can be categorise into distinct parts (central, marginal, distal) and sub-zones (A-E). This pattern is also consistent with results from other parts of Norway (Lofoten, Gabrielsen 1981, unpublished; southwestern Norway; Aamodt 1997). It should be noted, however, that reviewing the fracture-profiles one by one, the general pattern is not always achieved.

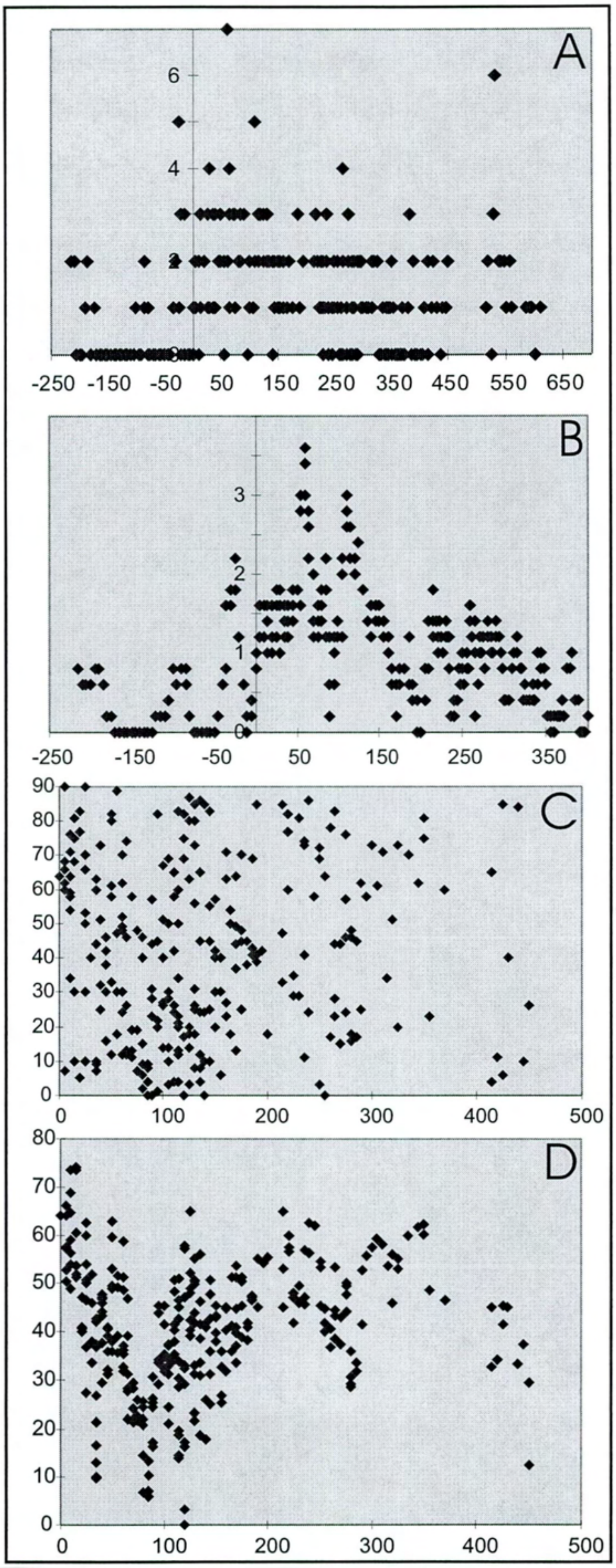


Figure 9.
(A) Distance versus frequency plot for the entire data set on E-W faults/lineaments.
(B) Distance versus frequency plot as a running average of five measurements sorted by distance for the data-set in A.
(C) Distance versus angle (strike) between each fracture and the E-W lineament for the entire data set.
(D) Distance versus angle plot as a running average of five measurements sorted by distance for the data-set in C.

3. FRACTURE DISTRIBUTION AND ARCHITECTURE OF LARGER EXTENSIONAL FAULTS IN NON-METAMORPHIC SEDIMENTARY ROCKS

3.1 Introduction

The central part of larger extensional faults in non-metamorphic sedimentary rocks are frequently dominated by less deformed or fractured rock lenses oriented sub-parallel to the dominant fault surface (Fig. 10). Lenses are usually separated by high-strain zones filled with ultramylonite, microbreccia, breccia or fault gouge and open or mineralised fractures. Pseudotachylites may occasionally be encountered. In some tectonic environments and in faults characterised by relatively low strain, the faults may be marked by swarms of deformation bands, characterised by mm-wide zones of grain-size reduction, which reduce porosity. By continued fault movements, strain hardening may cause spatial expansion of the zone of deformation bands transverse to the fault, ending with the development of a slip surface and formation of fault lenses (Aydin 1978, Aydin & Johnson 1978, 1979, Gabrielsen & Koestler 1987, Fossen & Hesthammer 1997, 1998).

Fracture analysis of cores from exploration wells drilled in the vicinity of major faults in the Norwegian continental shelf (Gabrielsen & Koestler 1987; Pedersen, pers. comm.1994; Gabrielsen et al. 1995, Gabrielsen et al. 1998, Aarland & Skjerven 1998) suggests that an enhanced fracture frequency, which can be directly related to the faults, is detectable only within a distance less than a couple of hundred metres away from the fault zone itself (Fig. 11), even for faults with dip-slip displacements in the order of 2000 - 4000 metres. There is also a tendency that the fracture distribution is asymmetrical, in that the hangingwall is more heavily fractured than the footwall. A similar pattern has been encountered in field analysis of faults with lesser displacement (one hundred to a few metres) in the Ebro Basin, in Utah and in Kilve (Bristol Channel) (Fig. 12). In the two former study areas (Ebro Basin and Utah) the rocks are siliciclastics (mainly conglomerates and sandstones), whereas the latter area (Kilve) is dominated by claystones, siltstones and carbonates.

Figure 10A

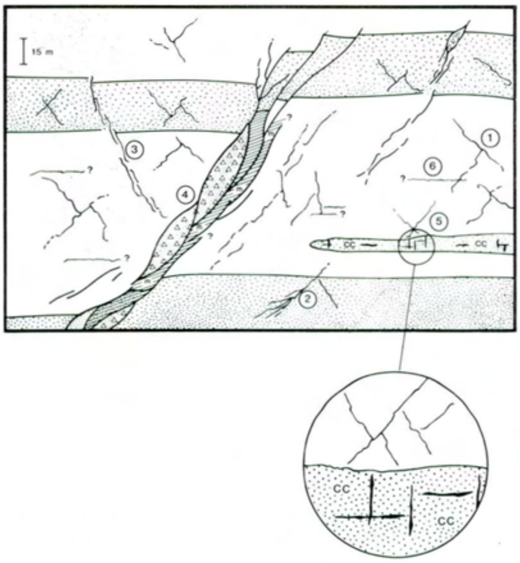


Figure 10B

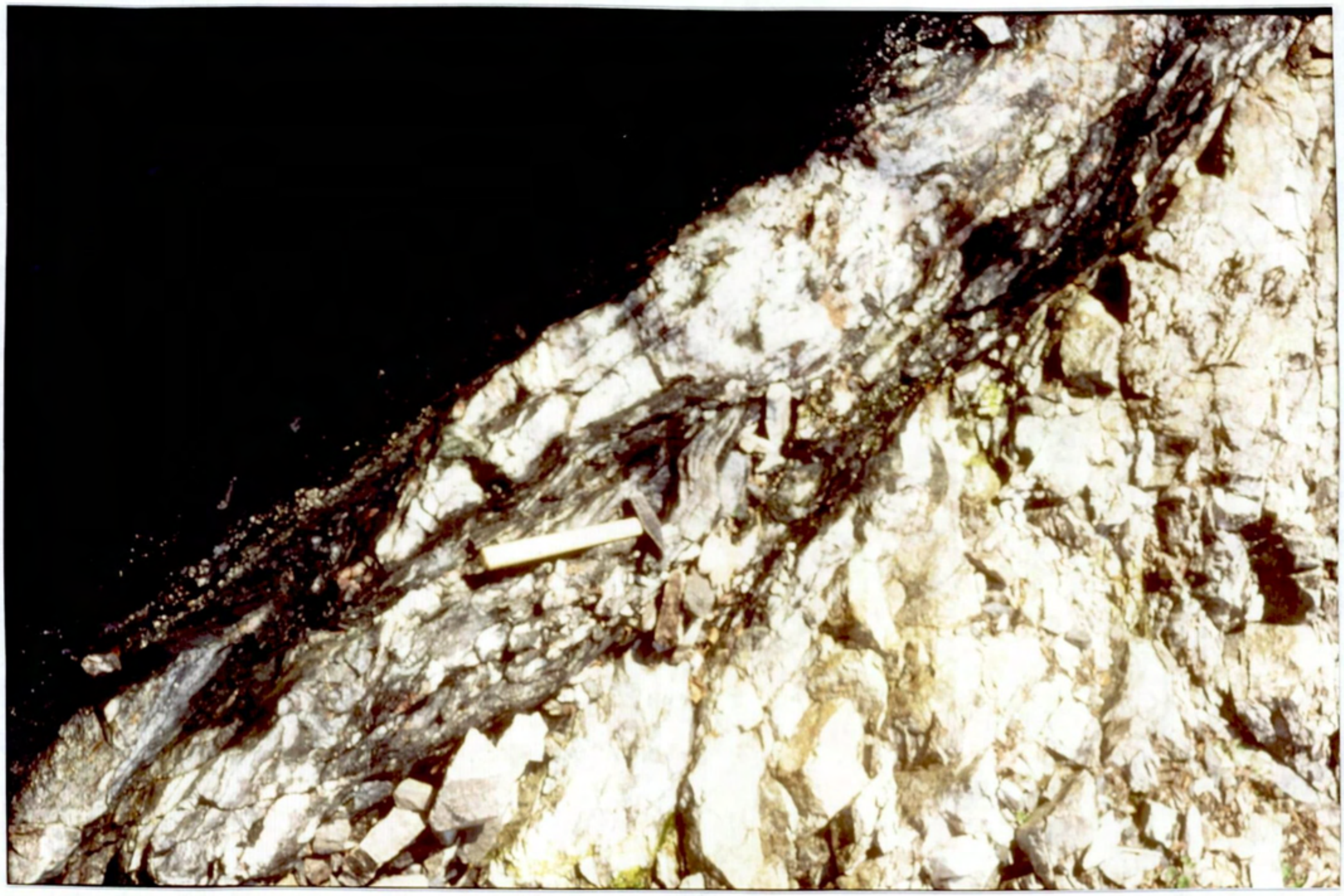


Figure 10. (A) Fracture types commonly found in hydrocarbon reservoirs. #1 Single fractures, #2 anastamosing fracture, #3 fracture swarm, #4 macro-fault with lensoid intrinsic architecture, partly mineralised, #5 contrast in fracture geometry due to different strength of layers, #6 sub-horizontal fractures associated with unloading. Triangles = breccia, m = mineralised zone, cc = calcite mineralisation. (B) Lensoid rock body entrapped in semi-brittle extensional fault zone, Hinnøya, Troms. Note hammer for scale.

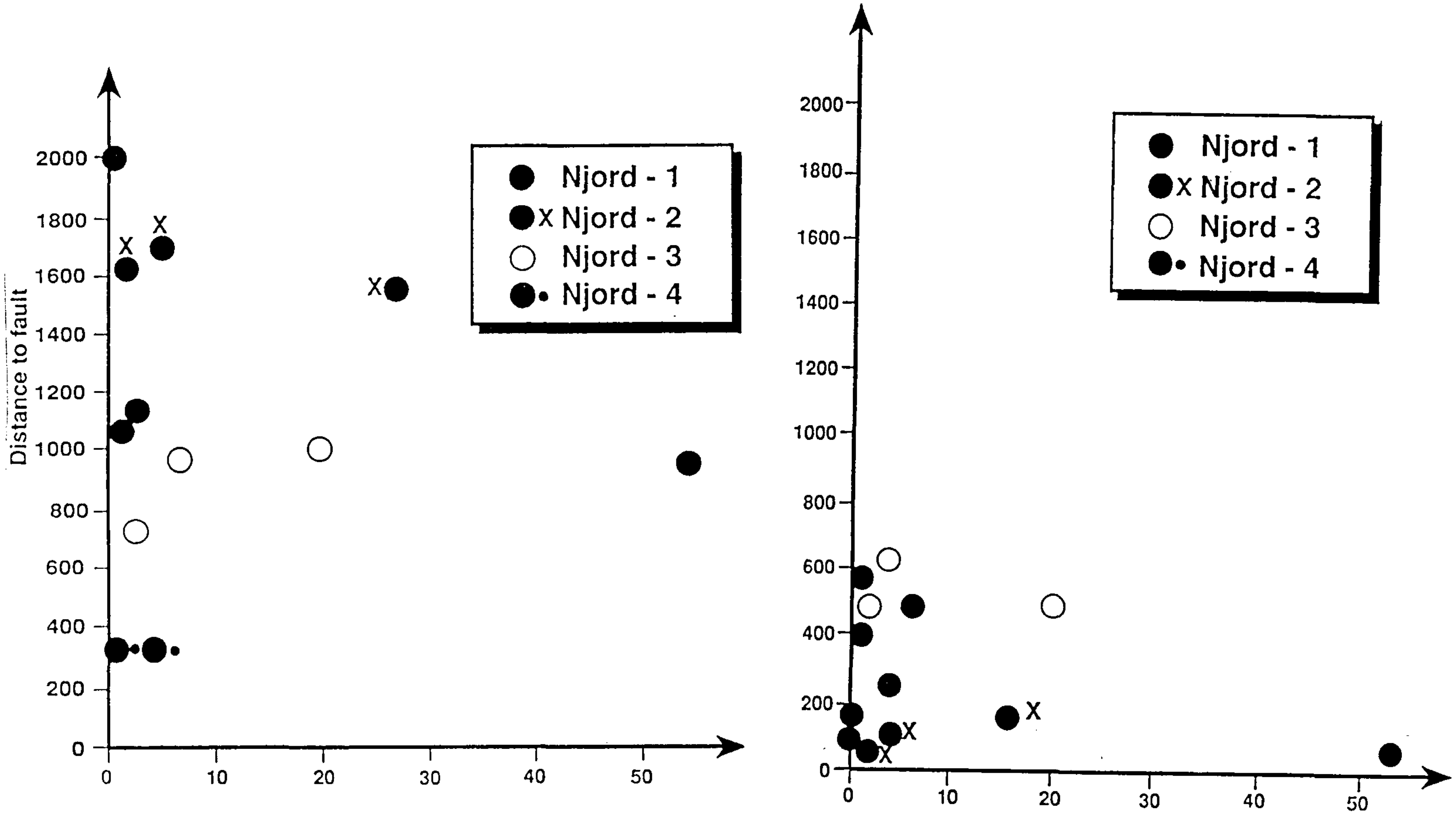


Figure 11. *Fracture frequency (f_f^a) vs. distance to master fault (vertical displacement in the order of several hundred metres), on sediments, Njord Field, mid Norwegian shelf. From Gabrielsen et al. (1993).*

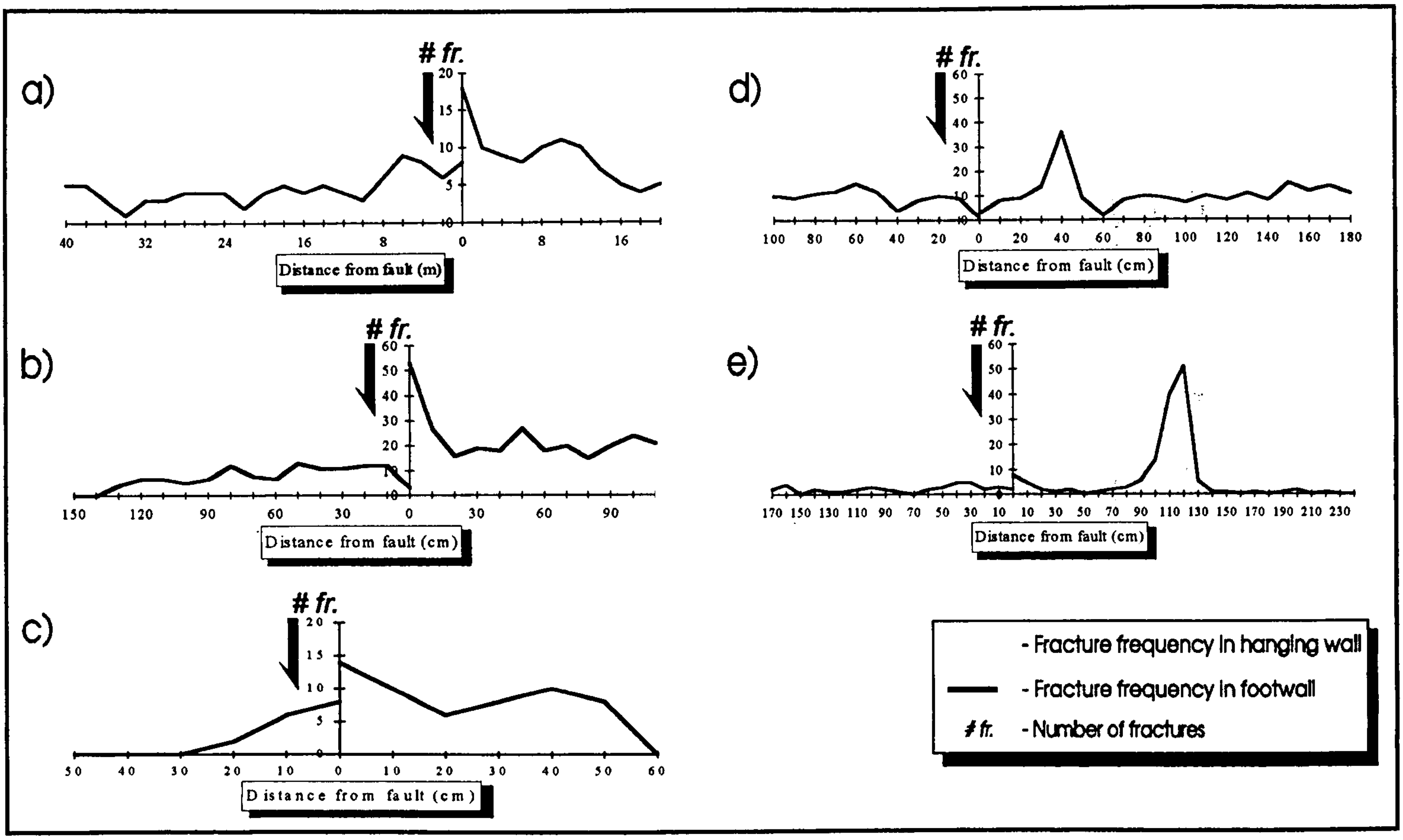


Figure 12. Fracture distribution in footwalls and hangingwalls of extensional faults in sediments, Mesaverde Group, Utah. From Gabrielsen et al. (1997).

An asymmetrical fracture distribution is also reported from large faults in sediments of East Greenland, however, with a reversed distribution as compared to that seen in the areas described above, in that the footwall is more heavily fracture than the hangingwall (Steen 1997).

The analysis of a large number of faults led Knott (1994) to suggest that the term 'damage zone' to be applied to the zone of exceptionally high fracture-frequency near the fault plane. It should be noted, however, that in certain circumstances, stress distribution in the footwall and hangingwall may be such that zones devoid of fractures are developed (Gabrielsen et al. 1998). In addition, strain softening in the fault zone may strongly influence further development of associated fractures.

The damage zones of the hanging- and footwalls may be composed of several fracture populations representing different types of structuring (Gibbs 1984, Gabrielsen et al. 1998)

(Fig. 13), all of which may be associated with fracture swarms near master faults. *Fault-parallel* fracture swarms may be dominant closest to the master fault, perhaps representing increased total slip during conditions of strain hardening. Such structures may develop into *splay faults* both in the footwall and the hangingwall (Fig. 13). Splay faults may define borders of rock lenses or shear (extensional, strike-slip or contractional) duplexes to be entrapped within the greater fault zone by continued strain. Finally, several types of *accommodation faults* may be present in the hangingwall. Such structures are related to synthetic or antithetical collapse of the hangingwall, perhaps reflecting the shear angle at a late stage of hangingwall deformation (Fossen & Gabrielsen 1995), or the forced deformation of the hangingwall as it is transported above an uneven fault surface (Gabrielsen et al. 1998).

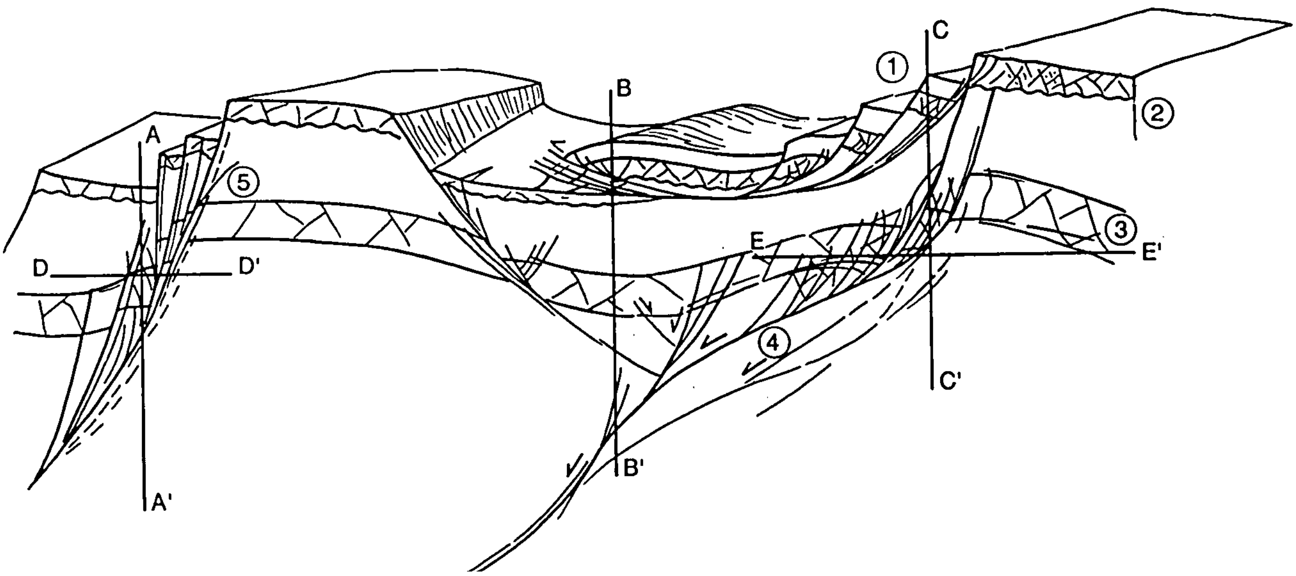


Figure 13. Different outlines of damage zones in footwalls and hangingwalls of extensional faults with different geometries. Damage zones encompass both antithetic and synthetic fractures, and the changing geometries result in different fracture frequencies when measured along different profiles (A-A', B-B', etc.). From Gabrielsen et al. (1997).

In summary, 'mature' faults in unmetamorphosed sedimentary rocks are frequently built around a central fault zone characterised by undeformed or fractured rock lenses separated by high-strain zones filled with ultramylonite, microbreccia, breccia or fault gouge and open or mineralised fractures. Zones of enhanced fracture frequency are commonly present on both sides of the fault, reaching fracture frequencies that may be up to two orders of magnitude greater than those recorded in the undeformed country rock. The higher fracture frequencies are usually recorded in the hangingwall fault block, although the opposite situation may also occur. The fracture frequency in most cases drops off quickly away from the fault zone, and enhanced fracture frequencies are usually not detectable more than a few tens of metres away from medium size faults (dip slip in the order of tens to one hundred metres), and a few hundred metres in large-scale faults (dip-slip 1000 - 4000 metres). This typical regular, exponential distribution in fracture frequency profiles may be overprinted and complicated by fractures belonging to fracture populations which relate to different types of accommodation structures (Fig. 13).

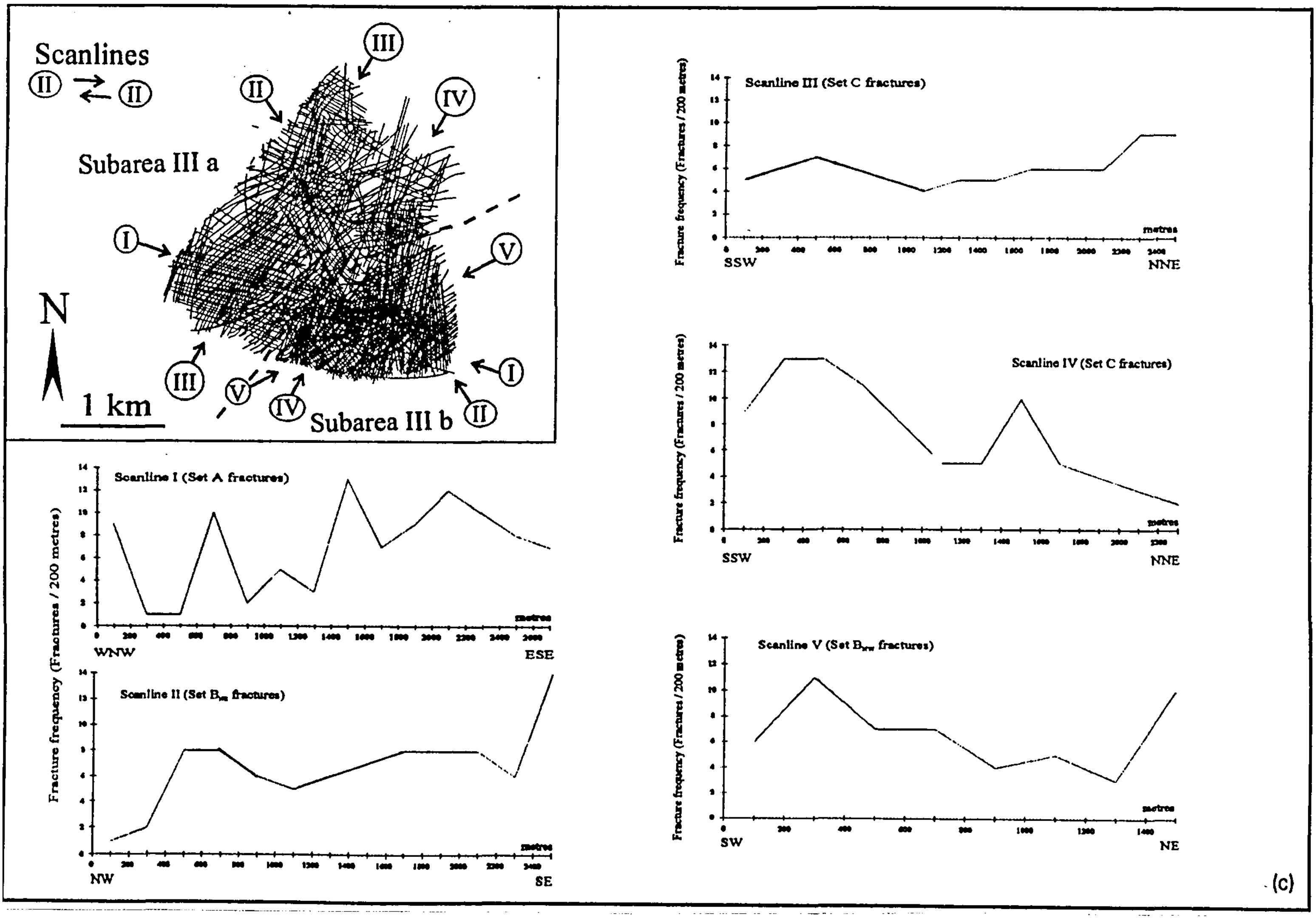


Figure 14. *Fracture distribution in (unmetamorphosed) coarse, clastic sediments of the Montserrat Fan Delta, Ebro Basin, Spain. The master fault (El Vallés-Penedés Fault) is situated immediately to the east of subarea IIIb. Note the steep increase in fracture frequency in the southeastern part of scan-line II. From Alsaker et al. (1996).*

3.2 Case study: The Ebro Basin, north-eastern Spain

The Ebro Basin is located in the north-eastern Iberian Peninsula. It is characterised as a Pyrenean foreland-basin with an E-W trending axis, and formed subsequent to emplacement of the Alpine South Pyrenean thrust system during the Paleogene. The northern margin of the Ebro Basin is defined by a south-verging imbricate structure which evolved during the Late Cretaceous (Maastrichtian) to Oligocene. Towards its south-eastern margin, the Ebro Basin is bounded by the Catalan Coastal Range (Anadon et al. 1985, 1986, Marzo & Anadon 1988).

In a recent fracture study in the Ebro Basin of north-eastern Spain, Alsaker et al. (1996) observed a close relation between fracture frequency and proximity to master fault zones. Thus, the areas closest to the extensional El Vallès-Penedès fault (sub-area IIIb in Fig. 14), which borders the Vallès-Penedès Graben, containing more than 4000 m of Miocene-Pliocene sediments, are characterised by fracture intensities which are 2-5 times as high as those found in the centre of the basin. The enhanced fracture frequency affects an area up to 1 km into the footwall of the master fault. In this study, it was also noted that the fracture frequency varies significantly with fracture orientation.

3.3 Case study: The Norwegian continental shelf

Master faults in off-shore areas are only rarely cored, because of the risk of encountering high-pressure zones that may produce uncontrollable pressure kicks, but also because cores are recovered only with difficulty. However, several larger faults have been cored, although very little information from such studies has been published. Gabrielsen et al. (1990, 1998) reported some data from such studies, but a more comprehensive report has recently been given by Aarland & Skjerven (1998).

In this case, the eastern Brage Horst fault was penetrated by a deviated well, and a sequence totalling 300 metres of deformed siliciclastic sediments became available for investigation. The eastern boundary fault of the Brage Horst, which is situated NE of the Oseberg Field at the eastern margin of the Viking Graben in the northern North sea, has a total dip-slip throw of the order of 200-250 metres, and is characterised by a lazy ramp-flat-ramp geometry (Fig. 15). In this operation both the central (zone A/B according to the nomenclature of the present report) and the more distal parts (zones C/D) of fault was sampled.
distribution with a marked increase in fracture frequency in the hanging wall as well as in the footwall (Figs. 15 and 16). In more detail, the study showed that the fault consists of two major branches, characterised by calcite-mineralisation, and enveloping a heavily fractured segment, together with a minor, separate footwall fault. The fault clearly affects the bedding orientation in a wide zone (Fig. 17), and reveals a variety of fracture orientations (Fig. 18).

The overall picture of the fracture distribution is that of a heavily fractured hanging-wall fault block, with an asymptotic growth in fracture frequency towards the fault zone, and a stable, mainly unfaulted foot-wall. This is in good correspondence with the results reported from the Ebro Basin and from Utah (above).

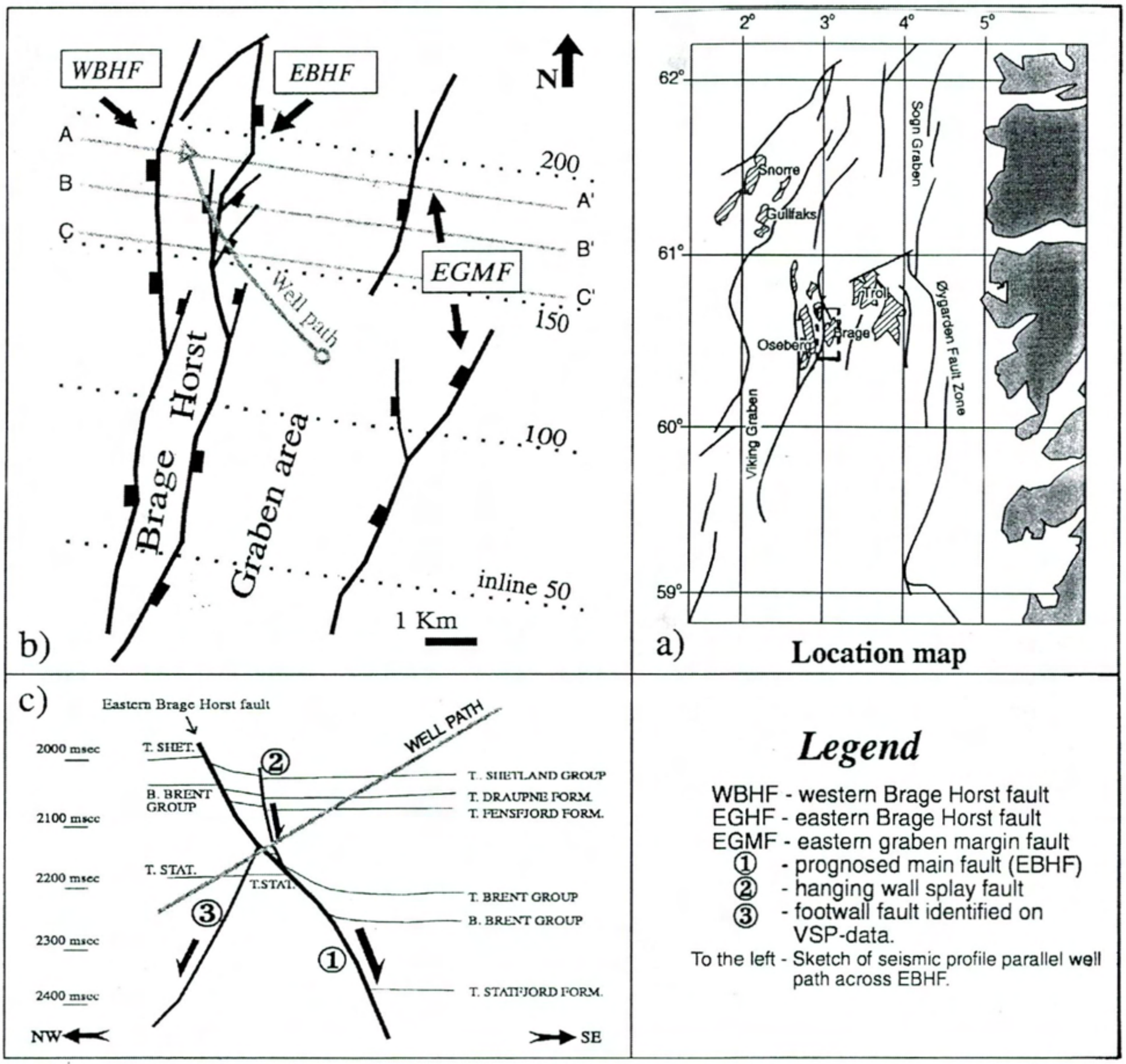


Figure 15. Structural framework of the Brage Field, northern North Sea, with position of well path through the fault characterised in Figs. 16, 17 and 18.

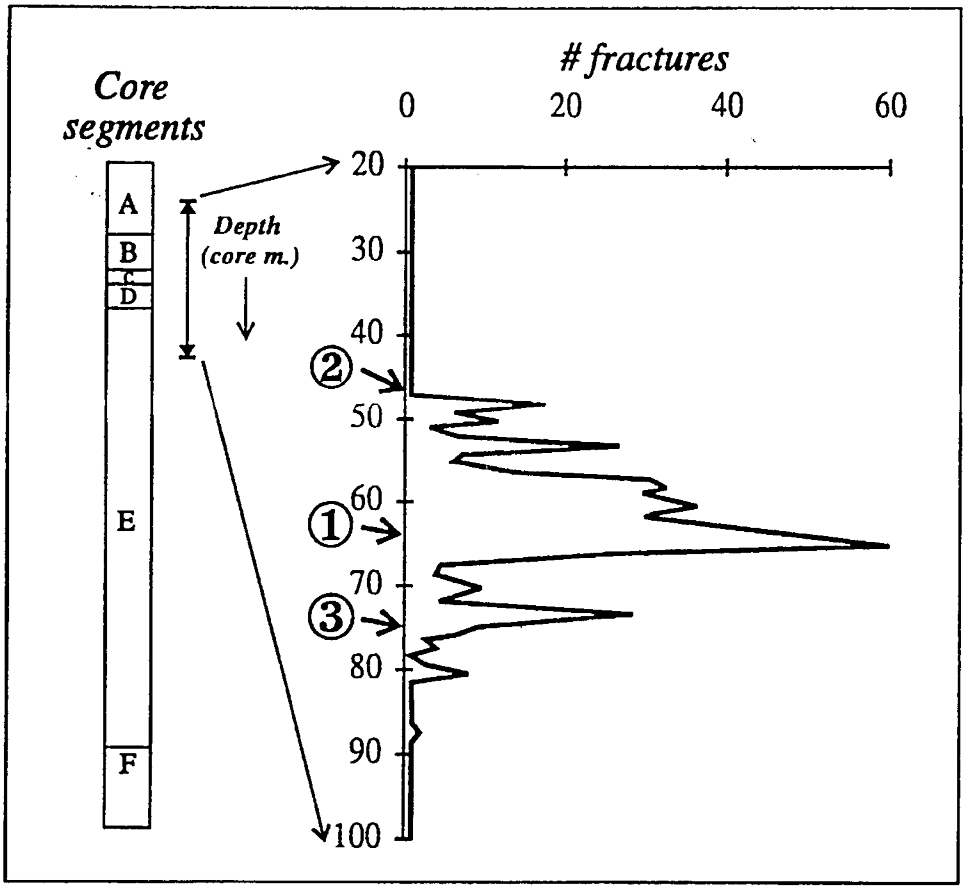


Figure 16. *Fracture distribution across fault zone of the Brage Field in the northern North Sea. 1 and 2 refer to calcite-filled fracture zones, whereas 3 refers to predicted depth of main fault. From Aaland & Skjerven (1997).*

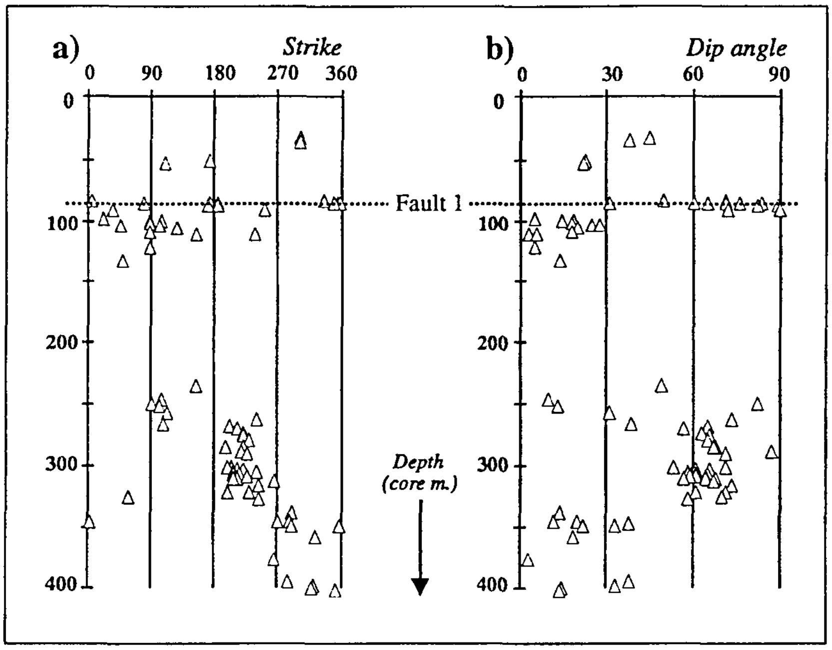


Figure 17. *Bedding orientation associated with fault zone at the western margin of the Brage Horst. From Aaland & Skjerven (1997).*

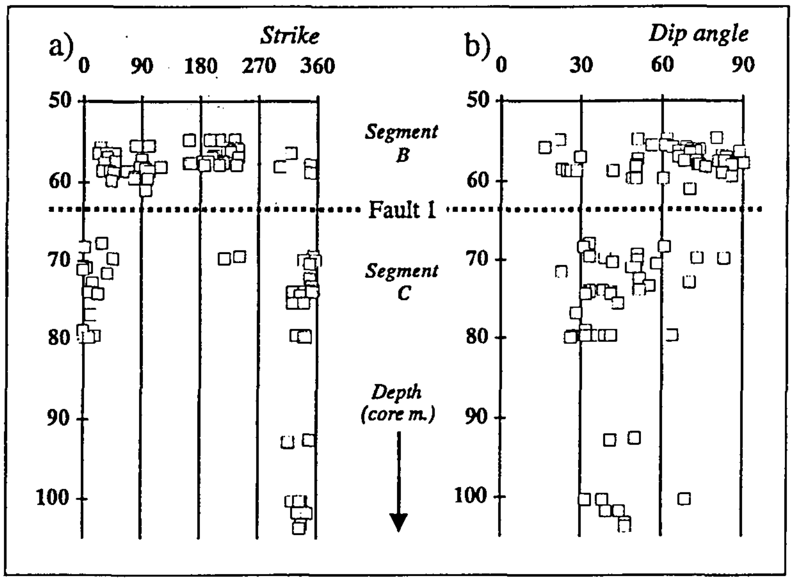


Figure 18. *Fracture orientation in damage zone of fault zone at the western margin of the Brage Horst. From Aaland & Skjerven (1997).*

4. MODEL FOR SPATIAL FRACTURE DISTRIBUTION

The presented studies of lineaments in metamorphic and sedimentary rocks open for a subdivision into two main types: faults and master joint zones, of which the latter apparently is a common structure in the metamorphic bedrock of Norway. Lineaments in the presented sedimentary basins are faults. Each type of fracture zone is characterised by a distinct architecture, which mainly differs in one major aspect - inclined faults appear to have an asymmetric fracture distribution with most of the strain in the hangingwall, whereas master joint zones and steep faults have a more symmetric distribution.

There are some general differences between lineaments in metamorphic and sedimentary rocks. Typically, fracture zones lineaments in metamorphic rocks show a more symmetric fracture distribution, a more distinct zonation, and a wider deformation zone than do fracture zones in sedimentary rocks. However, many striking similarities justify a general model for fracture zones.

Our detailed study on fracture distribution clarifies the detailed architecture of fracture zone lineaments, which can be divided into *central-, marginal- and distal parts*, and an additional, outboard part of general *background fracture* frequency. In more detail, as seen particularly along more mature lineaments (faults), the central and marginal parts may be further subdivided into the *zones A to D* (Gabrielsen 1981, unpublished). These parts, i.e. zones A-D, commonly make up the topographic feature of the lineaments, whereas *zone E* constitutes parts further away from the lineament. Thus, fracture data presented above support a model (Fig. 3) for fractures distribution around lineaments, commonly observed in crystalline bedrock of Norway, but also applicable to sedimentary rocks, which is described in a central to distal order:

4.1 Central part (Zones A-B)

The central part (zones A-B) is commonly 5 to 20 (max. 30-50) metres wide. It consists of a swarm of fractures in a complex network, generally subparallel to the lineament, with common frequencies in the range of 6-12 f/m, locally reaching 20-100 f/m or more. Breccia is common and appears to increase in width and length depending on the size of the first order structure. Longer and wider lineaments tend to have more extensive breccias.

Zone A, which is the innermost zone of the lineament, defines the area in which strain has been concentrated. This zone is commonly filled with fault rocks, usually constituting discontinuous networks or lenses, and is generally oriented subparallel to the master fault (e.g. Swensson

1990). Mineralisation is frequently encountered in zone A (Fig. 19a,b,c). In the Precambrian of Norway, it is not uncommon to find several generations of fault rocks ranging from (high-PT) mylonites to (low-PT) microbreccias and fault gouge (e.g. Fossen et al. 1997), indicating that some of the major lineaments are profound, crustal-scale zones of weakness that have repeatedly been reactivated in response to changing tectonic stresses and general geological conditions. In the examples investigated in the Precambrian of Norway, the zone A has been seen to vary in width from a few decimetres to approximately 20 metres.

Figure 19A

Figure 19B

Figure 19.

Different types of fault rock encountered in 'zone A' of lineaments.

(A) Strongly mineralised and brecciated crystalline rocks in mature zone A, Hinnøya, Troms.

(B) Fault gouge of vertical lineament in crystalline rock, Hinnøya, Lofoten.

(C) Incipient brecciation and mineralisation (red zeolites) in granitic gneiss, Ullabødalen, Sunnfjord.

Figure 19C

Zone B is a zone of very high fracture frequency (ff^{a} 10 - 100 m^{-1}), sometimes grading into protobreccia, and is commonly seen to border zone A on both sides (Fig. 20). The fractures are generally oriented parallel to the strike of the lineament, and may be affected by mineralisation or bleaching, suggesting that the zone acted as a zone of high fluid flow during deformation. By far the most common fracture type of zone B is mode II, although mode I fractures may be present. The width of zone B is variable, but has rarely been seen to be more than a few metres. Zone B is regarded as a high-strain region within the fault zone, albeit with lesser strain concentration than that typical for zone A.

Zone B is frequently seen to constitute the central part of lineaments (zone A lacking), perhaps indicating that the this represents the first stage in the fault activity, the fault slip being transferred to the inner part (zone A) by continued displacement and associated strain softening.

Figure 20.
(A) 'Zone A' (where the hammer is placed) and 'zone B' (left of hammer) of vertical fault in gneiss at Sandvika, near Askvoll in Sunnfjord.
(B) Zone A (two < 25 cm zones of breccia - deeply eroded) and zone B in quartzite along the Eikefjorden fault near Alnakken, Sunnfjord. Note the dense network of short fractures (locally > 100 f/m; zone B). The stick is 1 meter long.
(C) Close up of the photograph in Fig. 20B, showing details of zone B.

Figure 20B

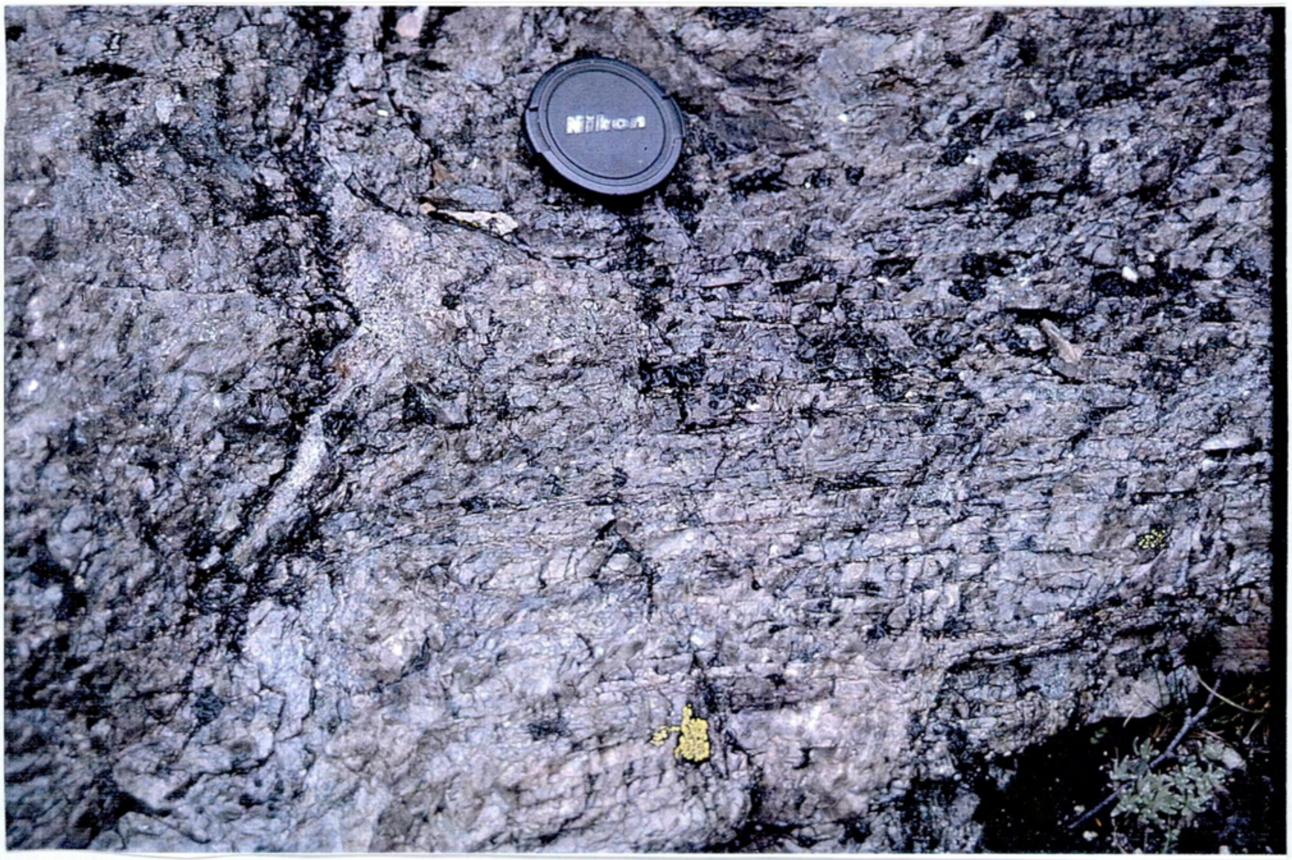


Figure 20C

4.2 Marginal part (Zones C-D)

The marginal part (zones C-D) occurs outside the central part, commencing 1 to 30 metres from the core of the lineament, and is 10 to 50 metres wide. Typical average frequencies are in the range of 3,5-4 f/m, and can locally, in narrow fracture trains, reach 6-12 f/m. This close to the lineament there is an increased tendency for subparallel fractures to the first order structure. In contrast to the central part, the marginal part rarely shows secondary minerals growth; most fractures thus tend to be open (not sealed).

Zone C defines a zone of fractures sub-parallel to the main lineament (Fig. 21a,b). Fracture frequencies are significantly lower (ffª 1-5 m^{-1}) than that of zone B, and mineralisation and leaching are usually absent. The dominant fracture type is mode II, although mode I fractures are far more common than in zone B. The border between zones B and C is commonly abrupt.

In cases where zones A and B are lacking, zone C may constitute the central part of the lineament, indicating that the lineament is characterised by a low total strain. In most cases, zone C is the widest zone of the lineament, and makes up its major part. Typical widths of zone C would be between 5 and 50 metres on both sides of zones A and B.

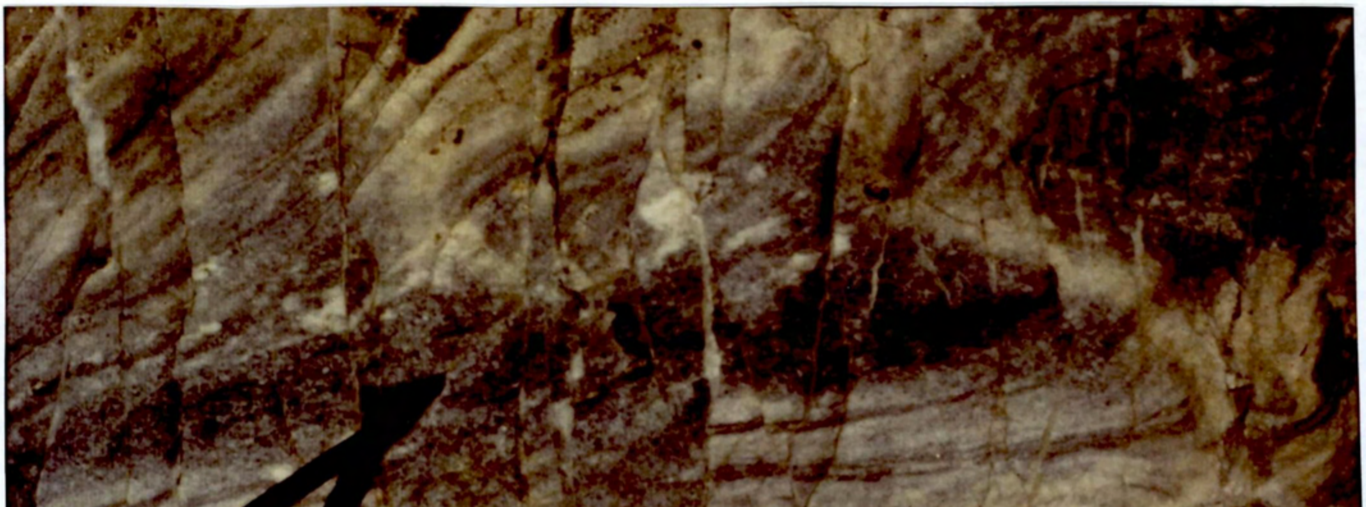


Figure 21. *(A) Transition between 'zone B' and 'zone C' of steep fault in marble, Storvatnet, Hinnøya, Troms. (B) Typical 'zone C' (where Ane and Silje are working) in gneiss, Ullabødalen, Sunnfjord.*

Zone D is less homogeneous than zones B and C when fracture orientation is considered, and usually displays a bipolar strike distribution. Zone D defines the geometric, physical and commonly also the topographical border of the lineament Two fracture sets are typically seen in the steep valley sides of lineaments which have a clear topographic expression (Fig. 22). Zone D is typically only 5-10 metres wide, and the fracture sets invariably have the characteristics of mode I fractures.

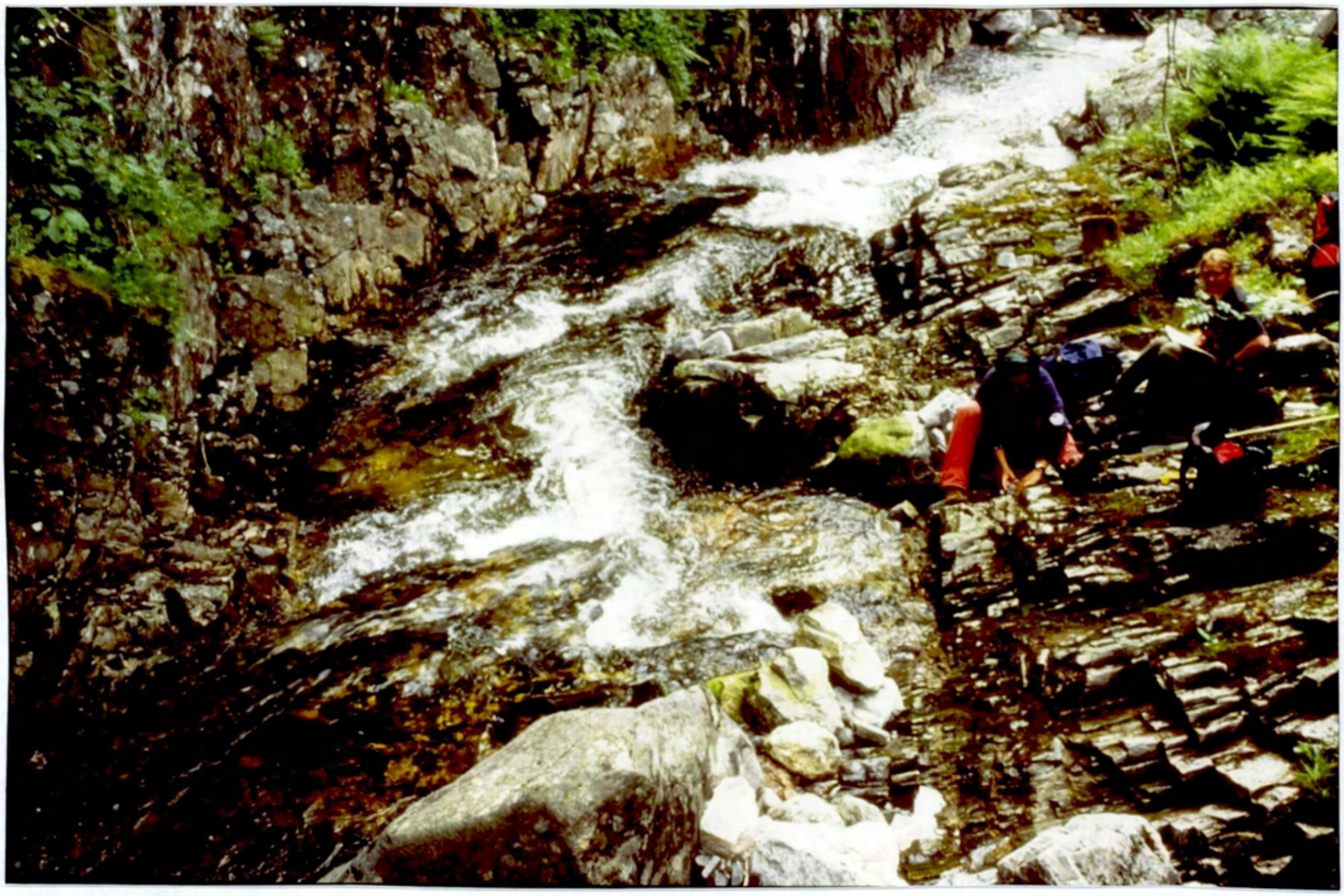


Figure 21B

4.3 Distal part (Zone E)

Zone E occurs 50 to 250 metres outside the lineament. It varies in width from 25 to 200 metres, and is characterised by frequencies of 1,5-2,5 f/m. However, even in this zone narrow (c. 1 meter wide) fracture trains may occur. Fracture-orientations show a tendency for angles between 0-40° to the lineament.

4.4 Background fracturing

Outside the lineaments, in areas not affected by these structures, the general background level of fracturing is present, with typical frequencies in the range of 0,5-1,5 f/m. This zone starts 150 to 300 metres away from the lineaments, and covers the entire stretch to the next lineament. Background fracturing thus represents the common fracturing of rocks over vast areas. However, even in this area narrow fracture trains (small lineaments) or more highly fractured lithological boundaries may occur.

Figure 22. Typical 'zone D' of steep fracture zone, Ullabødalen, Sunnfjord. Note the bimodal orientation of fracture surfaces in the rock-face.

5. DISCUSSION

5.1 Lineament classification and formation

The presented classification and definition of terms related to brittle features, as well as the discussed case studies, suggest that it may be convenient to define three main types of fracture zones, namely:

(i) inclined fault zones, which can be shown to offset lithological or other markers to a significant extent

(ii) steep fault zones, which can be further subdivided into:

(a) master fault zones, with significant offset of markers,

(b) immature fault zones, which show minor or no apparent displacement, but contain breccia,

(iii) joint zones, which are without displacement and breccia, or where criteria for identifying significant displacement are not available.

In the following we address the fundamental questions: How do fault and master joint zones form? Is the formation of fault zones and master joint zones related, and do they differ only in respect of the amount of incremented/total strain?

Steep master joint zones, steep fault zones and inclined fault zones reflect different principal stress configurations at the time of initiation and development (see section 6). This simple picture, however, is complicated since the architecture and fracture distribution within lineaments may be ambiguous. This because one lineament may carry the characteristics of more than one of the types. Such complications may most easily be explained by changing stress configuration with time, perhaps causing transformation of one lineament type into another, and hence, the overprint of contrasting structural features. There are, however, some profound limitations in such transitions.

In our opinion, fault zones develop either as (shear; mode II fractures) faults directly, or by shear superimposed on joint (tensional; mode I fractures) zones. This implies that a fault zone may develop from a joint zone, but that the opposite process is impossible, making a fault zone 'less primitive' in the hierarchy. In addition, the limits between the two types would be transitional, since the number of mode II fractures would tend to be growing with time at the expense of mode I fractures.

The formation of joints or faults requires that the tensile strength (T_0) of the rock is exceeded, and that the differential stress is of sufficient magnitude ($<4T_0$= joints, 4-$8T_0$= extensional shear, $>8T_0$=compressional shear, e.g., Hancock 1985). Experimental studies have shown that a cloud of micro-cracks develop as the failure-stress is approached; cracks which are becoming more localised and then rejuvenated and linked as the shear failure occurs (e.g., Kranz 1983; Pollard & Aydin 1988). This fracture process results in a general strain softening which, in contrast, may be followed by strain hardening related to formation of shear or deformation bands.

The formation of master joint zones and fault zones have been ascribed to several mechanisms, such as: (a) burial or uplift, (b) regional stress field, (c) local stress field, (d) intrusions, (e) meteorite impact (e.g., Engelder & Gaiser 1980; Davis & Reynolds 1996), (f) hydrofracturing. However, mechanisms for generation of regional joint zones are in fact less well understood (e.g., Nur 1982; Wise et al. 1985). The present case studies from crystalline rocks show that meso-scale fractures along these major fracture zones can be related to both mode I (extensional) and mode II (shear) fracturing, with fracture occurrences varying spatially and systematically perpendicular to the main structure. Although the main structure may have no visible displacement, internal strain on fractures along fracture zones favours faulting in addition to jointing, as exemplified by the formation of breccias. This suggests that many of the apparent master joint zones studied are actually immature faults, where mode II fracturing is successively superimposed on the pre-existing mode I joints.
The regional character of the master joint zones strongly supports that they formed in response to regional stresses, where $\sigma_1 \geq \sigma_2 \geq \sigma_2$ and $\sigma_v \leq \sigma_z = \sigma_h$*. In combination, an initiation of master joint zones as major (crustal scale) tension joints (mode I) is probable. According to Wise et al. (1985), the mechanisms are either (i) very minor horizontal stretching, or (ii) failure parallel to a compressional stress. In either case the formation of joints occurs above strata at depth that are characterised by more ductile behaviour, and the deformation may grade into faulting by time or at depth.

In addition to the genetical explanations proposed by Wise et al's (1985), we favour a mechanism for formation of master joint zones which emphasises the fundamental link between immature and mature fracture systems/zones, a process somewhat similar to formation of micro-cracks before faulting (e.g., Kranz 1983). Steep master joint zones may be generated in a principal stress configuration of $\sigma_3 = s_h$ and $\sigma_1 = s_v$, or $\sigma_3 = s_h$ and $\sigma_1 = s_H$. These two situations can, of course, not be distinguished as long as only joints (fractures without displacement parallel to its plane) are present in the zone. However, it is not impossible that the steep master joint zone is the surface expression of a deeper, inclined normal fault (see also Mercier et al. 1983), implying that the steep master joint zone can be transformed into a steep fault zone, and an inclined master fault zone thereafter. This general deformation process may be arrested at any time due to reduced or released stress, or because of mechanical changes, by that stopping the fracture zone development at a given step in its evolution. In both steep and

* σ_1= greatest principal stress, σ_2= intermediate principal stress, σ_3= least principal stress, σ_v= vertical stress, σ_h= minimum horizontal stress, σ_H= maximum horizontal stress, σ_z= stress parallel with the shortening axis.

inclined fault zones, principal stresses are indisputably $\sigma_3 = s_h$ and $\sigma_1 = s_v$. Accepting this model, it is obvious that the opposite transition is impossible.

Fracture zones, with their enhanced fracture frequencies, commonly constitute mechanical weaknesses that are favourable for reactivation. Thus, they may have been repeatedly rejuvenated during changing stress conditions, and during uplift and exhumation. This has important implications for porosity and permeability, and thereby the potential flow of fluids, within or along the zones.

5.2 Fractures and conductivity

Several factors are regarded significant for conductivity along fracture systems. Firstly, the permeability of single fractures is important, and is mainly controlled by filling and/or fracture surface roughness. Filling of a fracture by fault gouge or secondary minerals will hamper fluid flow by reducing the permeability or, when entirely filled, will seal the fracture. Roughness or anisotropy of fracture surfaces contribute to the aperture and thereby permeability. Lee & Bruhn (1996) concluded that mesoscopic faults have two main types of anisotropy; (i) directional aperture-enhanced anisotropy parallel to the fault movement by grooving and fibre growth, and (ii) isotropic aperture, where cross-hatches occur along the fault in combination with (i). Such cross-hatches may favour orthogonal-to-fault-movement fluid transport (Sibson 1996), since there is an tendency for increased aperture parallel to the hatches.
Secondly, conductivity along fracture systems will depend on fracture-connectivity. According to Long & Witherspoon (1985), Odling & Webman (1991) and Odling (1992), fracture networks with shorter fracture length and higher density tend to have lower permeability than networks with longer fractures and lower density. This is so because of the higher degree of interconnection for longer fractures.
Thirdly, stress is important both with respect to fracturing mode, as discussed above, and reactivation. The in-situ stress field acts on pre-existing fractures in two ways (e.g., Banks et al. 1996); (i) by influencing the shape and aperture of fractures, and (ii) by provoking new movement(s) along the fractures. In the former case, elastic strain in the rock will favour wider aperture and increased permeability along fractures parallel to the axis of maximum stress (σ_1), whereas fractures oriented normal to σ_1 will tend to close.
Reactivation of fractures may be a very important factor for fracture permeability, since most fractures tend to become sealed with time due to fluid circulation and fluid-rock interaction. This is well exemplified in Norway, where post-glacial rebound of metamorphic bedrock during the last 10.000 years clearly affects groundwater well yield (Rohr-Torp 1994; Morland 1997), i.e. increase with the uplift rate. With basis in the uplift rates, Gudmundsson (in press) calculated that the related tensile stress is sufficient to reactivate pre-existing fractures, as well as create new fractures in the upper hundred metres of the crust. This points to a link between uplift stress, fracture reactivation and well (fluid) yield.

5.3 Fluid flow in crystalline rocks

Our model on fracture systems, with symmetric patterns around master joint zones and immature fault zones, and asymmetric patterns with increased strain in the hangingwall of inclined mature fault zones, has important bearings on the potential conductivity, and thereby the groundwater potential, in fractured crystalline rocks. The model divides lineaments into parts and zones (see Fig. 3), which have distinct characteristics:

The *central part* (zones A-B) is dominated by an intense fracture network which is interlinked in three dimensions. The innermost part (zone A) is frequently characterised by breccias and secondary minerals (Figs. 19 and 20). Consolidated breccias and fracture minerals tend to clog or seal fractures. This results in reduced permeability, despite the high fracture connectivity and density. Thus, the central zone is less likely to yield much groundwater. However, in cases of reactivation, the renewed movement will be focused on the weaker rocks of the central zone. This may re-open fractures that were sealed, thereby enhancing conductivity.

The *marginal part* (zones C-D) probably constitutes the most promising part with respect to groundwater, and perhaps zone C in particular. This is because of a relatively high frequency of longer fractures with large surfaces, which are both parallel and oblique to the general lineament trend, and still with a relatively high interconnection rate (Figs. 21 and 22). The absence of secondary minerals further strengthens the potential for such fractures to act as good channels for water.

Distal parts (zone E), as well as areas with *background fracturing*, exhibits low frequencies of variably oriented fractures and rarely any secondary mineral growth. The low frequency reduces the interconnection between fractures, thus indicating a low to moderate conductivity. In areas with such low fracture frequencies, the in-situ stress may act upon single fractures rather than on fracture networks (zones A-D). Thus, permeability is likely to be higher along fractures parallel to σ_1, and correspondingly reduced along fractures oriented normal to σ_1.

Evans et al.'s (1997) study of permeability in faulted granite reveal a pattern that is in agreement with this model. Their conclusion was that the fault zone (our zones A-B) has low permeability, whereas the surrounding damage zone revealed high permeability (our zones C-D). The low-strain host rock (our zone E and background fracturing) was characterised by medium permeability.

The spatial distribution of high-conductivity zones within lineaments will vary. The along-strike variation is regarded as secondary to the across-strike variation, since the described parts and zones are dispersed parallel to the main structure. For example, the potential high conductivity in zone C will probably remain high for considerable along-strike stretches. Late reactivation, parallel to the main structure, would also add to strike-parallel conductivity.

Across-strike variations is regarded as more significant, since transverse conductivity would be hampered by tight zones (e.g., mineralised faults) as well as by highly (along-strike) permeable zones, causing strike-parallel leakage. Impermeable zones (zones A-B) are thus crucial for strike-normal conductivity. This suggests that fault zones, commonly showing more extensive brecciation, alteration and deformation bands, may be less prosperous with respect to groundwater/fluids than joint zones.

In summary, the general pattern appears as follows: Joint zones will have relatively high conductivity, fault zones relatively low conductivity, and 'hybrid' zones (immature faults) medium conductivity. In more detail, the marginal part (damage zone) and especially zone C has the highest conductivity. Therefore, groundwater wells aimed on zone C potentially will have an increased well yield.

5.4 Effects of fractures in siliciclastic sedimentary rocks

5.4.1 Meso-scale faults

Meso-scale faults in siliciclastic sedimentary rocks frequently display the same complexities as those of fault zones in crystalline rocks. Development of lensoid country rock bodies, complex juxtaposition patterns, fault rocks, clay smear and the co-existence of within-fault-zone mode I and mode II fractures contribute to this (Gabrielsen & Koestler 1987, Knipe et al. 1996, 1997 Gabrielsen et al. 1997a), as does the development of damage zones in the footwalls and hangingwalls.

In reservoirs where fracture analysis is based primarily on reflection seismic data, wireline logs and cores, a scale problem in fracture analysis exists. This is because little information is available on fractures on the scale between faults with throws above the resolution of reflection seismic data, which at the very best is in the order of ten to twenty metres (Badley 1985), and fractures mapped in cores which typically display offsets of centimetres and occasionally decimetres (Gabrielsen & Koestler 1987). In a number of recent studies, attempts have been made to bridge this problem by the study of scaling relationships in fault populations.

5.4.2 Variables in development of natural fracture populations in siliciclastic sedimentary rocks

To prevent mixing of fracture populations in core logging, it may be convenient to divide fractures into three groups, namely (1) fractures related to gravitational surface instabilities, (2) fractures generated by volume reduction during burial (including those related to water-escape), lithification, and uplift and unroofing, and (3) fractures of tectonic origin. Mode I and mode II fractures, as well as hybrid fractures, may occur under all these conditions, presumably with changing frequencies. The identification of the different types of small-scale fractures is not trivial, since characteristics such as morphology and texture may be similar for fractures that initiated and developed under different geological conditions. Therefore, the total geological environment of the fracture population needs to be taken into consideration in fracture analysis.

Synsedimentary gravitational surface instabilities may trigger faults that reach large dimensions, e.g. in the form of synsedimentary faults or growth faults, slumps and slides (Hardin & Hardin 1961, Rider 1978, Crans et al. 1980, Stow 1986). These structures are related to surface slope, intrinsic instabilities like inverse gravitational contrasts, or tilted contact surfaces between layers with contrasting shear strength or fluid pressure (Cloos 1968, Bruce 1973, Mandl 1988 p. 28-29). It is frequently assumed that tectonic tilting or seismicity may contribute to destabilisation causing synsedimentary faulting (e.g. Mayall 1983). In such circumstance, local stresses may easily exceed the shear strength of sand and other unconsolidated or poorly consolidated sediments, and may result in the development of discrete shear zones (Owen 1987, Maltman 1987). Such shear zones may not be readily distinguishable from tectonic shear zones in the field, inasmuch as morphological and textural characteristics of fractures may not be diagnostically different from those related to tectonic deformation (Petit & Laville 1987, Gabrielsen & Aarland 1990). However, enrichment of clay minerals (Sverdrup & Bjørlykke 1992) and the association with fluidisation processes (Owen 1987) may help to identify synsedimentary faulting. Consequently, fractures associated with synsedimentary gravitative features may sometimes only be identified as such by their overall geometry and the geological environment in which they occur (e.g., Martinsen 1994).

Fracture frequency and spatial distribution for synsedimentary gravitational faults are not frequently reported in the literature, but field studies of slides in both fluviodeltaic (Sverdrup & Bjørlykke 1992), marine sediments (Farrell & Eaton 1988) and accretionary prisms (Behrmann et al. 1988) have indicated that fractures related to soft-sedimentary gravity-driven, near-surface deformation may be abundant, and that they locally may develop frequencies comparable with that of regional (contemporary) or local tectonic deformation (Martinsen & Bakken 1990). It is suggested, therefore, that fracture systems associated with synsedimentary gravitational instabilities may contribute significantly to the bulk fracture strain in a sedimentary rock, and may even dominate such populations on the local scale. However, synsedimentary gravitational instabilities may be strictly local phenomena, where the strain (e.g. extensional faulting in the upper extensional listric fan of a slump) can be compensated for in the outcropping (compressional) toe-zone, so that the bulk contribution to regional deformation may be zero.

It is therefore anticipated that fractures generated by synsedimentary, gravity-induced deformation potentially represent a source of error in scaling relationship studies if one fails to identify such features. Since gravity-induced sliding is not uncommon in the Norwegian shelf at different stratigraphic levels (Gabrielsen & Robinson 1984, Gabrielsen 1985, Alhiali & Damuth 1987), it may be anticipated that fractures associated with gravity-induced near-surface sliding locally contribute significantly to the total fracture frequencies obtained from core logging in this area.

Fractures related to compaction and decompaction. Compaction fractures may be initiated at very shallow levels in the subsurface (a few tens of metres; Maltman 1988), and continue to develop throughout the burial history of the sediment (Fig. 23). Agents other than sediment loading (e.g. ice; Banham 1975, Bell 1981 Visser et al. 1984) can contribute to their initiation.

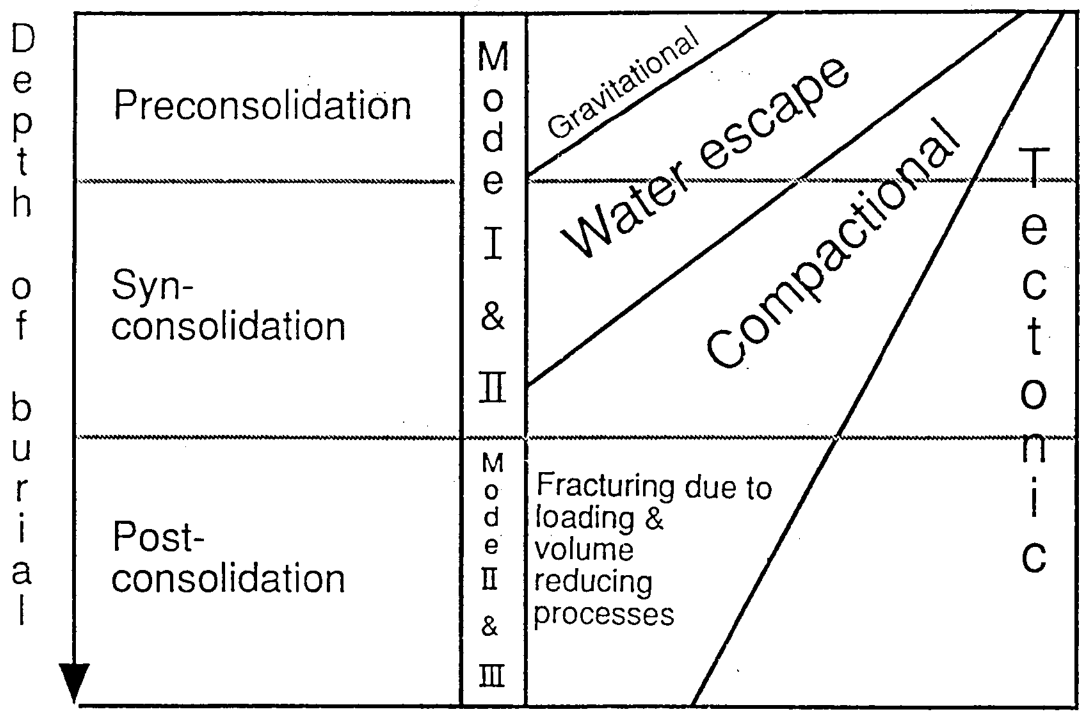


Figure 23. Scheme for possible environment for different fracture types in sediments during burial and unroofing. Note that tectonic influence can occur simultaneously. From Gabrielsen et al. (1993).

According to some authors (Engelder 1985, 1993) fracture populations generated by compaction and decompaction may be dominated by tensile (mode I) fractures. For tensile fractures, Voight & St.Pierre (1974) demonstrated that the horizontal stresses σ_x and σ_y are related to the vertical stress σ_z, and that they vary through burial according to:

$$\sigma_x = \sigma_y = (\upsilon/(1-\upsilon))\sigma_z + [\alpha E \Delta T/(1-\upsilon)] \quad (3)$$

(where υ = Poisson ratio, E = Young's modulus, α = the thermal expansion factor of the rock). Hence, the factor $(\upsilon/(1-\upsilon))\sigma_Z$ describes the Poisson effect and $[\alpha E\Delta T/(1-\upsilon)]$ is the thermal effect. Engelder (1985) used this relationship to substantiate the claim that sediments which have undergone burial and lithification may attain tensile stresses which exceed the tensile strength of the rock during unroofing, so that sets of mode I fractures develop. The expression (3) clearly demonstrates the influence of the rheological properties of the deforming rocks for the vertical stress generated as a consequence of loading. The practical implications of this is shown in Fig. 24, which suggests that a sandstone will undergo tensile fracturing at a level where a shale remains undeformed during erosion and uplift. Diagnetic processes and fluid pressure influence the rheological properties of the sediments, and hence, the fracturing during later eventual erosion and upheaval.

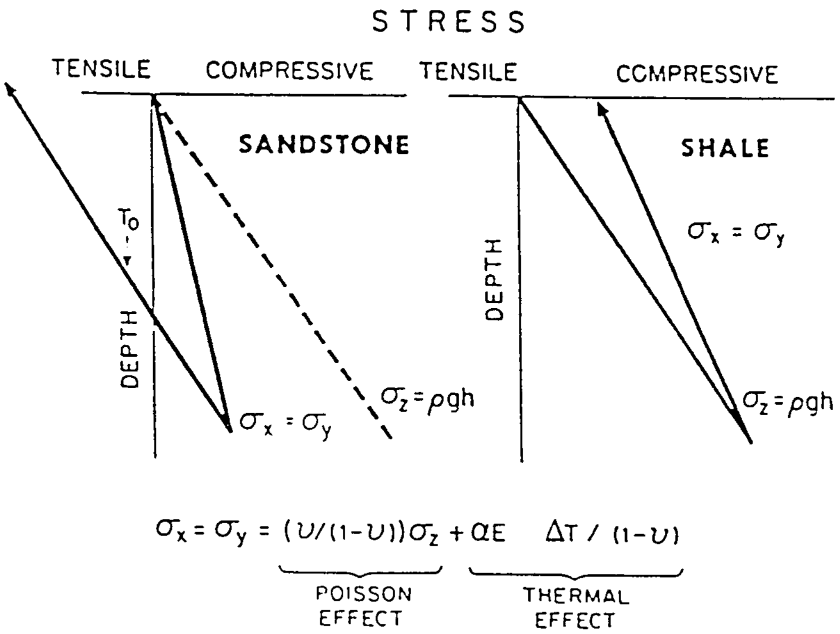


Figure 24. *Stress path for sandstones and shales during burial and unroofing. From Engelder (1985).*

Mode II fractures generated through vertical loading are also frequently reported in unconsolidated and poorly consolidated sediments, and are indeed reproduced in sand box experiments (e.g. Mandl et al.1977). The shear strength may vary from a few kPa to some MPa during lithification (Maltman 1994). Such values may be well within the range of effective stresses near the surface or at depth in zones with high fluid pressure, so that shear fractures may also be expected to develop. In fact, all fracture populations in most of the hydrocarbon wells in the Norwegian continental shelf logged by us, are dominated by mode II fractures, and layers with a substantial number of mode I fractures are only rarely found. In field studies (Montserrat Fan-Delta northeast Spain; Alsaker & Gabrielsen submitted, and Mesaverde Group, Utah; Alsaker unpubl. data 1995), mode I fractures constitute a much larger part of the fracture populations. This is in accordance with previous observations from these areas (Engelder 1993, p.56). At the moment it is unclear whether this reflects the sampling procedures (most investigated wells are vertical), or whether it is a consequence of different stress histories in the different basins, or if it can be explained by different depths of burial of the investigated sediments.

Fracture logging of siliciclastic sediments in cores are most frequently performed in hydrocarbon reservoirs which may be siltstone- and sandstone-dominated in intervals up to a few hundred metres thick, presently buried at depths between some hundred to 3000-4000 metres. In vertical intervals of about 400 metres thickness, buried at depths between 1000-1400 metres, σ_{eff} changes in excess of 10 MPa (calculated from data by Haimson & Doe 1983), and such stress differences may, depending on P_{fluid}, be sufficient to induce fracturing of the sediment. Fracture frequency studies of wells of the Njord and Oseberg Fields in the Norwegian continental shelf, covering up to 350 metres (between 2750 and 3100 metres below present sea-bed) thick sand- and siltstone-dominated intervals, display no simple relationship between fracture frequency and depth (Fig. 25), suggesting that the effect of changes in σ_{eff} in intervals of such magnitudes may be almost completely overruled by other parameters like P_{fluid} and lithology (Gabrielsen et al. 1993). This observation may seem to be supported from similar studies elsewhere (Haimson & Doe 1983). Surprisingly, however, no correlation between fracture frequency and over-pressured intervals was found in wells studied in the Njord Field on the mid Norwegian shelf (Pedersen, pers. comm. 1994).

In most cases, where $\sigma_H = \sigma_h$, fractures related to compaction and decompaction would have predictable orientation and geometry, as they are related to a vertical $\sigma_v = \sigma_1$. It is also to be expected that fractures related to compaction are more evenly distributed throughout the deforming sediment (Gabrielsen & Koestler 1987). The two sets of mode II-fractures would expectedly be planar structures with a dip of 60°, and would be defined by deformation bands with the characteristics of the early stages of deformation, i.e. grain compaction and porosity reduction. In fine- and medium-grained rocks, this type of fracture has a high potential of preservation (Maltman 1988). Unfortunately, such textures also characterise tectonic fractures, and there are obvious problems in separating fractures related to compaction from those related to tectonism because any rock exposed to tectonic stresses will simultaneously be affected by stresses related to the overburden. These difficulties may perhaps be overcome by mapping the spatial distribution of fractures relative to larger faults.

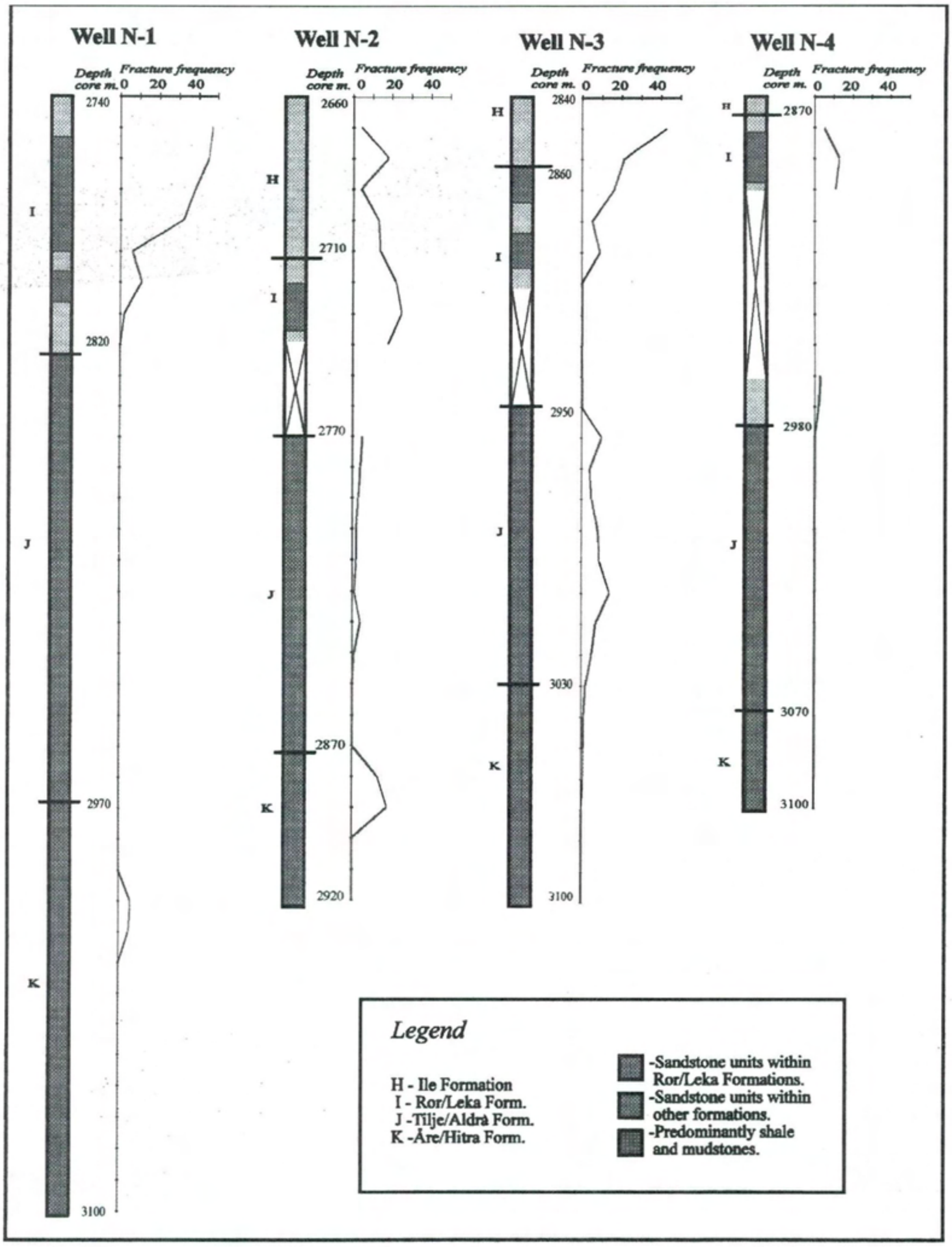
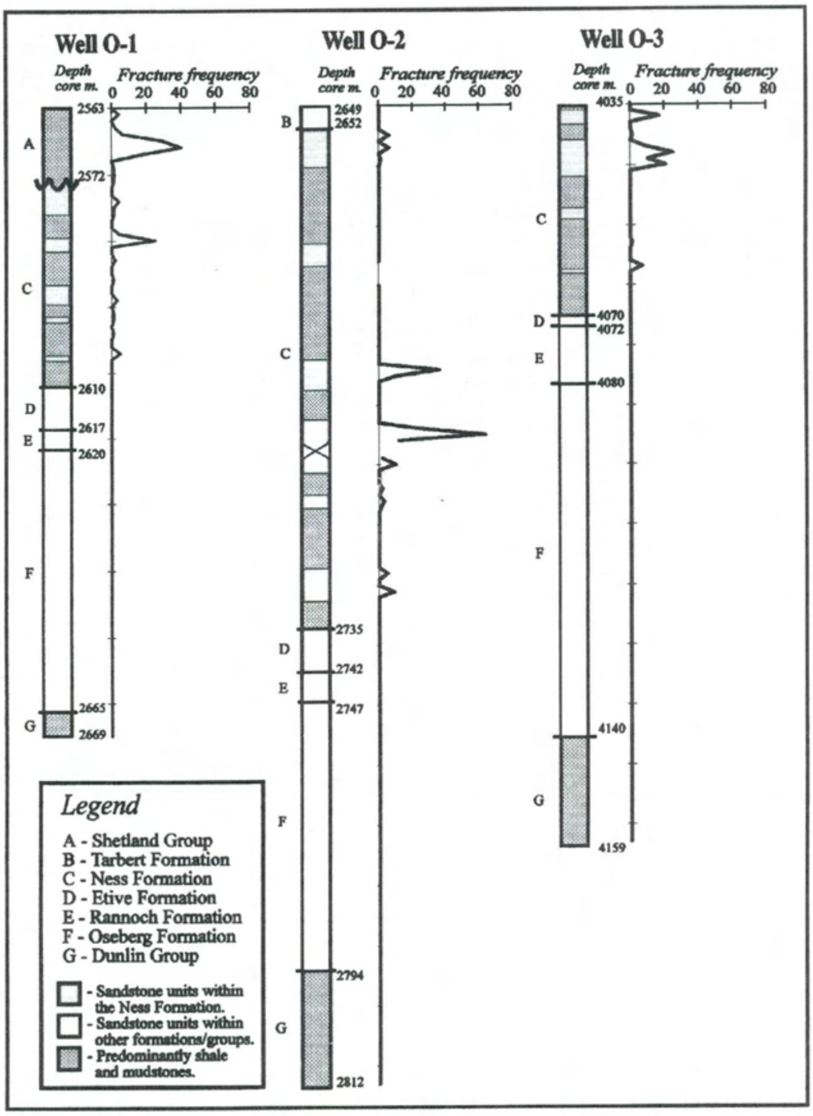


Figure 25. Examples of fracture vs. depth curves in sediments: On the scale of a few 100 metres (see core depth in diagram), no systematic variation can be seen. From Gabrielsen et al. (1997).

Figure 25 cont.



Water-escape fractures may be seen as a type of compactional fractures, preferentially active during the early stages of burial. In siltstones and sandstones these are typically high-angular, planar features, usually a few millimetres wide. They are often associated with symmetrical upward- or downward-pointing drag. Outside the zones affected by drag, there is commonly no relative offset of strata (Burbridge et al. 1988). These structures may further be subdivided into two classes (Gabrielsen & Aarland 1990); fractures (joints) through which water has escaped (Lowe 1975, Cheel & Rust 1986, Guiraud & Séguret 1987), and fractures (faults) associated with collapse due to volume loss associated with water escape (Burbridge et al. 1988, de Lange et al. 1988). The deformational style associated with water escape may vary considerably with lithology, from single fractures in sandstones to fragmentation in shales. Texturally, water-escape fractures in fine to coarse-grained, siliciclastic sediments are characterised by lamination parallel to the fracture walls, laminae being defined by clay particles and other tabular mineral grains (Maltman 1988). Internal, wall-parallel laminar elutriation of mineral grains may occur (Lowe 1975, Burbridge et al. 1988), and there are indications that packing of grains takes place in this process. The textures may depend upon degree of water flux, grading from hydroplastic through liquefaction to fluidisation, the latter representing the highest degree of grain sorting.

In conclusion, fractures related to water escape in fine- to coarse-grained siliciclastic sediments are most commonly strata-bound, and are usually associated with other structures related to liquefaction. Hence, they are generally easily identified on the scale of a well core where they happen to be more abundant. Consequently, water escape fractures do commonly not contribute significantly to fracture frequency, and are not believed to represent a great problem in the study of scaling relationships of fracture populations in fine- to medium-grained siliciclastic sediments. In mudstones and shales, however, the problems may be significant (Guiraud & Séguret 1987, Petit & Laville 1987).

Tectonic fractures affect the entire crust and may be generated near surface as well as in rocks at a deep level of burial (Fig. 23). In many published studies, tectonic fractures are implicitly considered to dominate natural fracture populations. Under moderate stresses, morphologies and texture of the fracture fill of tectonic fractures probably have characteristics that are similar to those of compaction fractures, and identification of the two types may be difficult. However, the textures related to tectonic fractures will frequently reflect strains that exceed those of consolidation fractures. Comprehensive descriptions of tectonic fractures are given in the literature (Stearns & Friedman 1972; Dunn et al. 1973; Pittman 1981; Nelson 1981; Narr & Currie 1982; Tillman 1983; Tillman & Barnes 1983; Watts 1983; Aydin & Johnson 1983; Nelson 1985; Hancock 1985; Wilke et al. 1985; Gabrielsen & Koestler 1987; Groshong 1988; Lorenz et al. 1988; Knipe 1992; Antonellini & Aydin 1994; Antonellini et al. 1994).

We consider the major faults (those which are above the limit of resolution in reflection seismic data) and those fractures which have first-order relations to the major faults, to be related to the contemporary (regional) and the secondary (local) stress-field derived hereof. Where σ_h and σ_H were significantly different during deformation, such structures may provide predictable fracture orientations and relations. In most extensional systems investigated by us, fault populations are characterised by relatively uniform fracture frequencies within each fault block, with the fracture frequency increasing both in the hangingwall and the footwall in the proximity of major faults. Widths of the zones with enhanced fracture frequencies ('the damage zone') in the hangingwalls and footwalls of extensional faults with vertical throws in the order of some tens of metres, is frequently recorded to be less than 20-30 metres. This is in general accordance with several previous studies (Jamison & Stearns 1982, Chester & Logan 1986), although the width of the damage zone is reported to vary (Knott et al. 1996; Beach et al. 1998). It is also evident that strain, and hence fractures, commonly are concentrated in hangingwalls adjacent to irregularities like flats or splays, so that abnormally wide damage zones can be expected at such sites (Koestler & Ehrmann 1991, Aarland & Skjerven 1998).

The present data may seem to confirm the general model for growth of fractures (deformation bands) in siliciclastic sediments from single fractures, widening into zones of deformation bands as described by Aydin & Johnson (1978). This may be in general agreement with the model of Walsh & Watterson (1988, 1989) which predicts that fractures grow concentrically from a point of nucleation into a surface with elliptic circumference. It is noted, however, that in plaster (Fossen & Gabrielsen 1995) and sand (Dahl 1985) analogue modelled faults are frequently seen to nucleate at the surface or at the contact between the deforming 'sediment' and basement. The development of single deformation bands into deformation zones also suggests that strain hardening plays an important part in the development of shear fractures in porous siliciclastic sediments. It may also be questioned as to whether there is an upper limit for such mechanism to prevail in the development of larger faults.

5.5 Quantification of sealing capacity of faults in siliclastic sedimentary rocks

Meso-scale faults in siliciclastic reservoirs will, in general, be permeability-reducing or sealing more often than not. There often seems to be a relation between sealing capacity, structural history, tectonic regime and orientation relative to the present stress system (Knott 1995). Thus, it is well established that faults frequently provide hydrocarbon seals in basins where rotated fault blocks define hydrocarbon traps (Hardman & Booth 1991). Also, flow in many fluid reservoirs are influenced by fracture systems where small-scale or even microscopic fractures are involved, where the fractures are acting either as permeability-enhancing elements (Nelson 1981; Narr & Currie 1982; Tillman 1983; Tillman & Barnes 1983; Watts 1983), or as permeability barriers (Stearns & Friedman 1972; Dunn et al. 1973; Pittman 1981; Aydin & Johnson 1983; Wilke et al. 1985; Gabrielsen & Koestler 1987; Lorenz et al. 1988; Antonellini & Aydin 1994). This emphasises the need to predict fracture frequencies and fracture distribution in fluid reservoirs, and to evaluate permeability characteristics of fractures at all scales.

In general, the complex geometries and deformational conditions which are associated with meso-scale faults, make analysis of sealing capacities difficult. Although estimates of sealing capacities based upon general characteristics (orientation, amount of throw, stress system and intensity) may be useful for specific fault sets (Knott 1994), a full analysis of the individual fault is usually still necessary to provide a solid fundament for reservoir modelling in-put (Knipe 1995, 1998).

To overcome the obstacle of fault zone complexity, it has been proposed to define faults as swarms of deformation bands in in-put for reservoir evaluation (e.g. Gabrielsen & Koetsler 1987). It is realised that this is an oversimplification, but it may be the only alternative when hard data (direct observations, cores or wireline logs) are lacking.

Several recent works have demonstrated that small-scale faults (deformation bands) in siliciclastic sediments frequently suppress permeability with two orders of magnitude, and up to five orders of magnitude in some cases (Antonellinei & Aydin 1994, 1995; Bjørnevoll 1996; Lothe 1998).

6. MODEL FOR THE INTRINSIC ARCHITECTURE OF LINEAMENTS AND FRACTURE PROFILES

The importance of the intrinsic architecture of lineaments is well established, and a considerable variance has been reported (e.g. Isachsen 1981) and description above. Thus, detailed analyses of the fracture distribution within and in the immediate surroundings of lineaments reveal a variety of fracture frequency profiles, fracture styles and distribution of fault rocks. Several parameters are expected to vary in the development of lineaments (master joint and fault zones), including lithology, depth of deformation (PT-conditions), strain intensity, strain rate and the nature of the actual local and far field stress systems (relation between s_v, s_h and s_H).

In the following discussion, the influences of these factors are considered in an attempt to explain the different lineament architectures resulting from the analysis of the available data, as categorised in Fig. 1. In this scheme, fracture distribution and orientation, fracture types, deformation products, and strain intensity are applied as variables, and used to determine fracture environment, stage of development, and the potential influence of strain-hardening and strain-softening. These data are applied to generate fracture profile types which are characteristic for each lineament type.

It is hoped that the resulting model may increase the ability to predict and evaluate fracture distribution and the potential of fluid flow in lineaments.

In accordance with the descriptions given in the first part of the present report, lineaments are subdivided into three families: Steep master joint zones, steep (immature/master) fault zones, and dipping (extensional) master fault zones.

Although all three lineament families display fracture zoning, the fracture distribution and orientation as well as the dominant fracture type and deformation is distinctive for each family. Typically, the lineaments are surrounded by an 'external' fracture set or system.

6.1 Steep master joint zones

Steep master joint zones are dominated by mode I fractures. Internal zonation is defined by fracture frequency alone, termed 'sub-zones' in the following to avoid confusion with the joint zone itself. In mature, steep master joint zones, five internal sub-zones (A - E) are frequently identified, of which E is an 'external' fracture set (see also description above). However, all of the sub-zones are not always present, so that steep master joint zones missing e.g. sub-zones A and B are not uncommon (Fig. 3). The sub-zones themselves and the fractures within each sub-zone are dominantly parallel to the regional trend of the lineament. A general exception to this is sub-zone D, which commonly displays a bimodal, internal fracture orientation. The fracture distribution of the steep master joint zones is symmetrical around the central sub-zone A.

The fracture surfaces of the steep master joint zone are commonly smooth, and sometimes display fringes and plumose structures (in the context of Hodgson 1961), strongly suggesting that the fractures are mode I-structures which initiated and grew in the brittle regime, in a relatively shallow position in the crust. The strong general parallelism of the individual fractures as well as the fracture sub-zones is indicative of $\sigma_3 = s_h$ and $\sigma_1 = s_v$, or $\sigma_3 = s_h$ and $\sigma_1 = s_H$. A discussion of these relations follows below.

The bimodal orientation of sub-zone D is almost ubiquitous in steep master joint zones, and is most easily explained as meso-scale F-joints and cross-fractures (Kulander et al. 1979, Kulander & Dean 1985) associated with the growth of the master joint zone (Fig. 26). Accepting that sub-zone D represents a system of interfering meso-scale F-joints and cross-fractures, and applying the concept of Dunn et al. (1994), it may be possible to analyse sub-zone D in more detail, but this is considered to be beyond the scope of the present report. Nevertheless, since the steep master fault zone according to the definition given above, is dominated by mode I-fractures, fault rocks, which requires shear (mode II or mode III fractures), should be absent in such zones.

It is tempting to ascribe the varying internal distribution of fractures and the presence or absence of sub-zones to differences in total strain or 'stage of development' of the master joint zone. Before doing so, however, the potential effects of mechanical strain modification of the bulk strength (in principle strain softening and strain hardening) of the zone should be considered. Strain hardening is most commonly associated with shear and the development of fault rocks. The development of steep master joint zones can most probably be described by the Mohr-Coulomb-Navier fracture criterion for a cohesion-less (or very weak) rock (Fig. 26). Hence, a strain-hardening scenario for steep master joint zones can principally be ruled out.

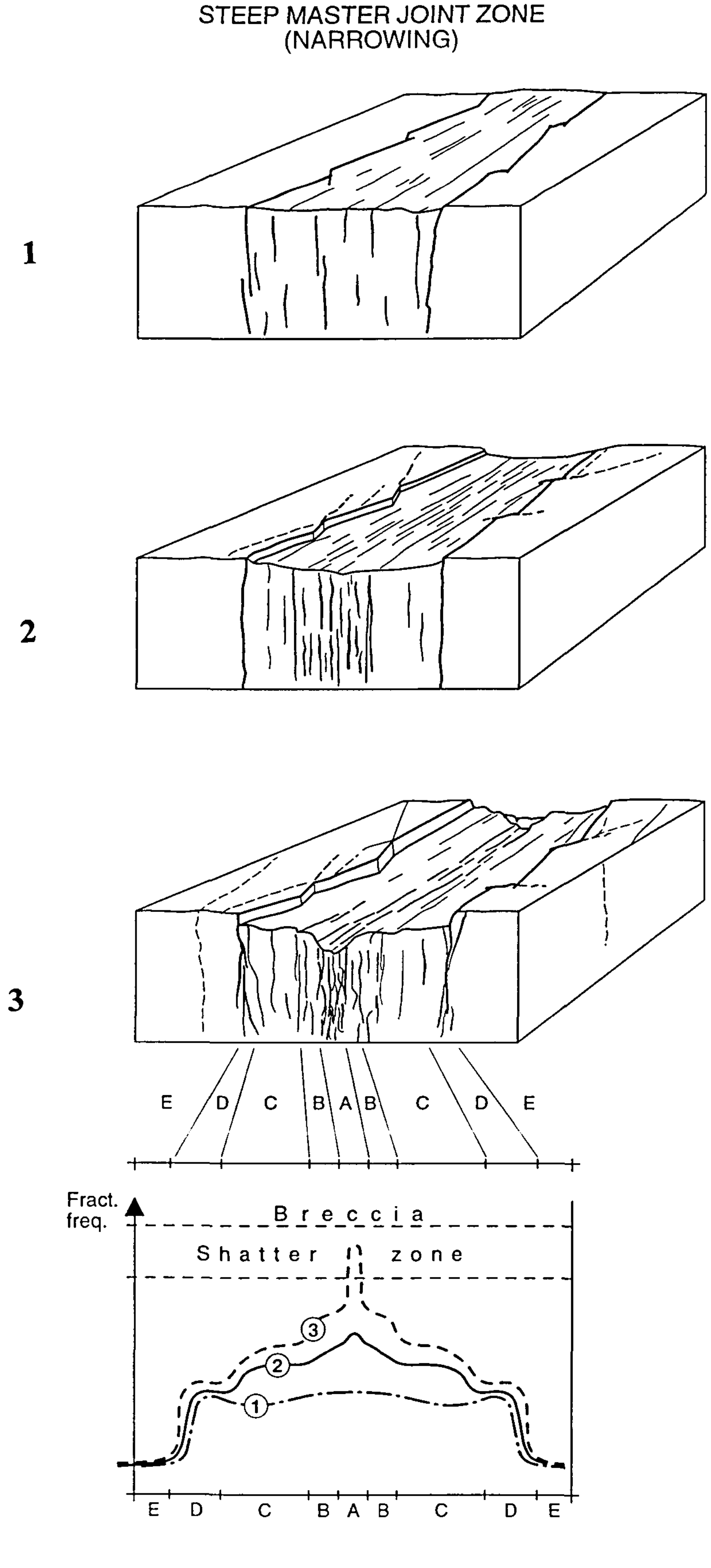


Figure 26A. *Three-stage scheme for development of **steep, narrowing master joint zone**, and associated schematised fracture profiles.*

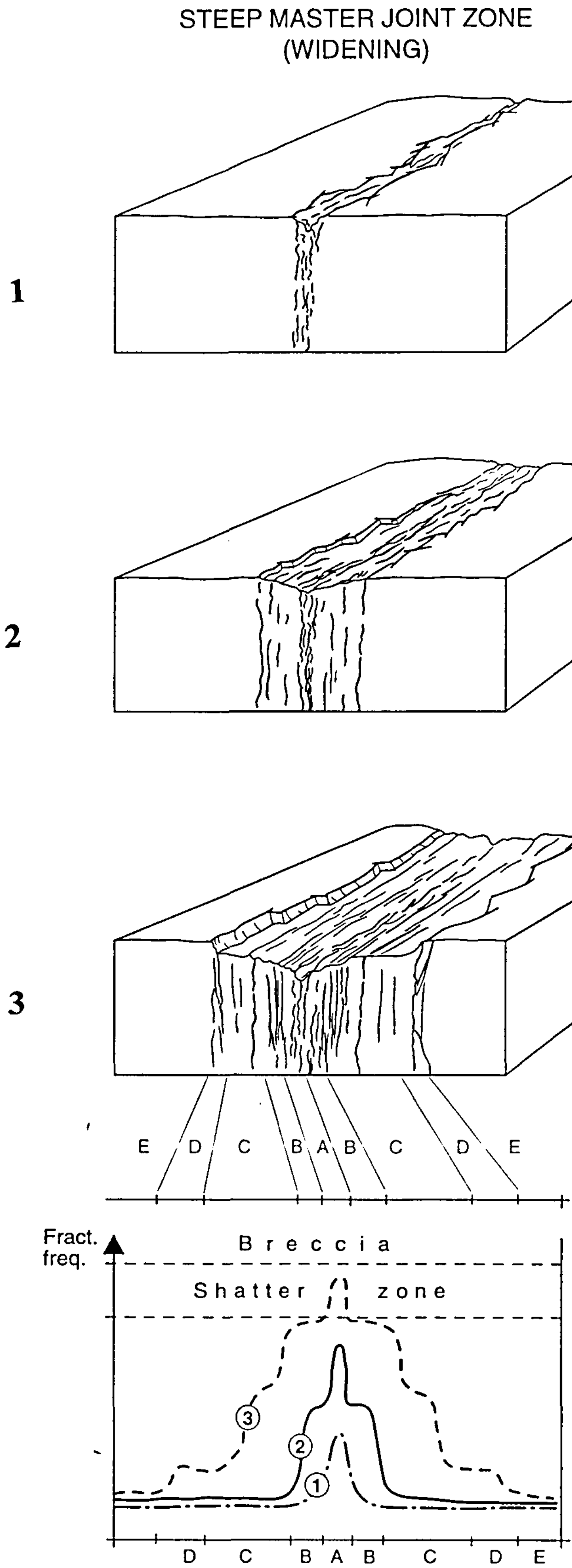


Figure 26B. Three-stage scheme for development of ***steep, widening master joint zone****, and associated schematised fracture profiles.*

Two models for the development of steep master joint zones can be envisaged (Figs. 26a and 26b):

In the first case, the zone spreads out laterally from an initial sub-zone A, which consists of a steep, high-frequency zone of mode I-fractures (Fig. 26b). Since no strain hardening is possible, this can only happen where the country rock is very weak, and in cases where the fracturing process is too fast (energy in-put to high; Bahat 1991) to be accommodated by the growing fracture swarm within the central part of the lineament. In this model, several fracture fronts, and hence, associated fringe zones on several scales (sub-zones D) should be expected. Although such complexities are seen in some cases, they are rare in the areas studied by the present authors, and outward-growing joint zones are generally not considered to be very common.

In contrast, the strain-softening model shown in Fig. 26a is considered more likely and much more common in nature. In this case, a wide fracture zone is initially present. The initial deformation zone (sub-zone E) includes the future sub-zones A-D. It is noted that, applying the outline with meso-scale F-joints and cross-fractures as described above, it is reasonable to ascribe the initiation of the bimodal fracture orientation of sub-zone D to the earliest stage. During continued development, the jointing concentrates in a narrower (sub-zones A-B) and finally in a central sub-zone (A).

Although the final fracture frequency profiles of the two types of master joint zones as described above are similar, some characteristic differences may be found. It should also be noted that these differences are perhaps greater in the incipient and immature stages of the joint zone development (Fig. 26):

Since the bimodal orientation of the fractures in front of the widening joint zone is outward-moving (strain hardening; first case above), joint zones of this type should be characterised by several fracture sets of bimodal orientation throughout the lineament, whereas such geometries are found only in the inner part in the joint zone where jointing concentrates in he central part at the mature stage (strain softening). The expected stage-wise fracture frequency profiles of these two lineament types can be studied in Fig. 26a,b.

In both cases, the country rock will be less resistant to erosion after (and during) fracturing, and will therefore define a negative topographic element, with the most intensely jointed central part (sub-zone A) as the topographically deepest element, commonly filled by lakes, rivers and creeks.

6.2 Steep master fault zones

Steep master fault zones are distinguished from the steep master joint zones basically in that they are dominated by mode II-fractures, have more complex inner (sub-zone) architecture, and contain fault rocks in central parts of the mature lineaments. Except for this, the fracture frequency distribution and zonation may be very similar to that of the steep master joint zones, and the complete sequence (sub-zones A-E) may be distinguished, with the fault rocks usually restricted to sub-zones A and B. The fracture distribution of the steep master fault zones is symmetrical around the central sub-zone A.

The fracture orientation within the sub-zones of the steep master fault zones are frequently bi- or multi-directional, and the single fractures will often have the characteristics of shear fractures, displaying Riedel-shears and sometimes P-shears. Mineralisation is more common, and the fracture surfaces are frequently striated. Thus, the steep master fracture zones bear all the characteristics of shear, implying that they were generated in a situation where $s_3 = s_h$ and $\sigma_1 = s_v$, or where $\sigma_3 = s_h$ and $\sigma_1 = s_H$ (see below for a more thorough discussion).

Following the description given above, it is likely that fracture populations of steep master fault zones encompass primary elements resulting from tensile fracture propagation (F-joints and cross-fractures; see description of steep master joint zones), conjugate fracture sets, and sets originating from primary shear in shear zones (Riedel shears). This is consistent with the bimodal fracture orientation associated with the steep master faults.

Strain-softening and strain-hardening are more likely to occur in steep master fault zones than in steep master joint zones, and two equally deformation paths are proposed. In the stain softening scenario (Fig. 27a), the total width of the zone is defined at an early stage of development. The fracture frequency is even throughout the fault zone, and the deformation products of the early stage may include fault rocks. During continued strain, the active deformation zone narrows, and the areas of the most intense deformation (fault rocks and high frequent fracture sub-zones; A-B) are concentrated in the centre of the fault zone.

In the strain hardening scenario (Fig. 27b), the development is the opposite of that of strain softening, so that the deformation zone widens with time, even causing brecciation outside the central sub-zones (Fig. 27b).

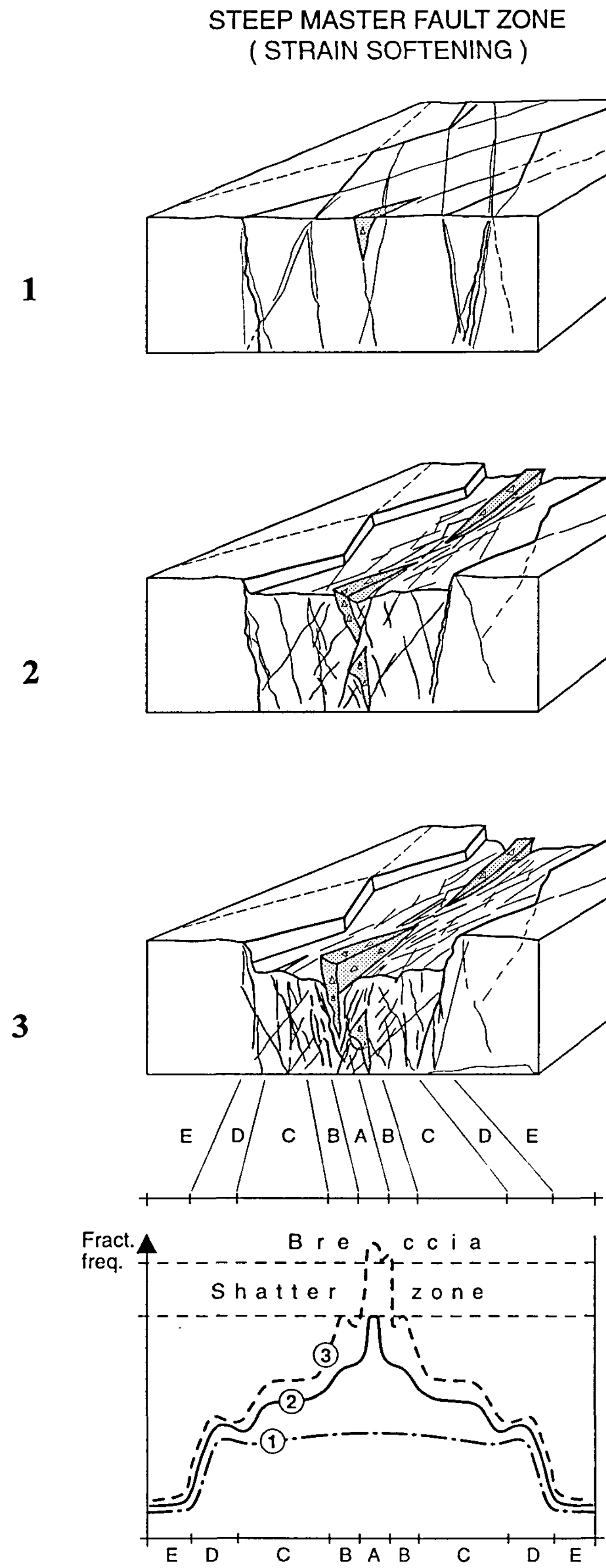


Figure 27A. Three-stage scheme for development of a ***steep master fault*** *during* ***strain softening****, and associated schematised fracture profiles.*

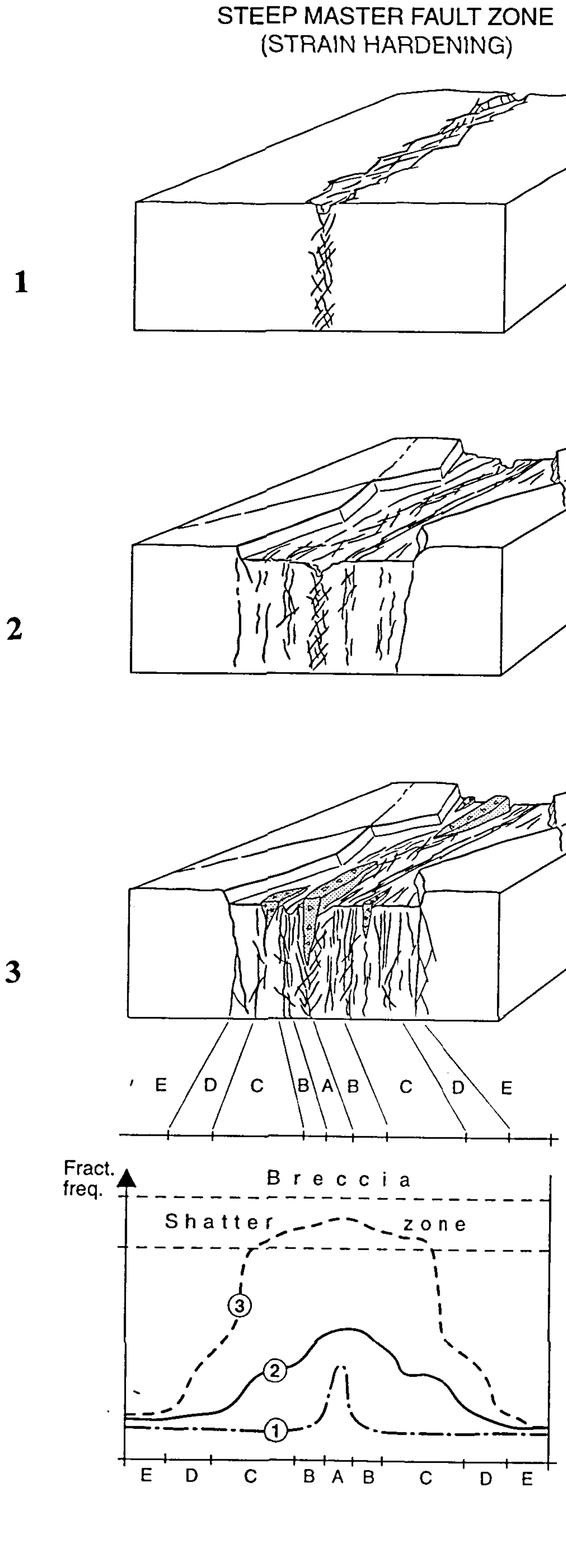


Figure 27B. Three-stage scheme for development of a ***steep master fault*** *during* ***strain hardening****, and associated schematised fracture profiles.*

6.3 Inclined master fault zones

Inclined master fault zones are characterised by simple shear, causing an asymmetrical fracture distribution in the footwall and hangingwall fault blocks (Fig. 28a,b). The development of inclined master fault zones follows that of the steep master fault zones above, in that a strain hardening and a strain softening scenario may be envisaged. A big difference, however, is the asymmetrical fracture distribution, commonly with most strain in the hanging wall, when the fracture distribution on both sides of sub-zone A is considered.

Also, according to the Andersonian criterion for fault orientations, the principal stress situations are that of $\sigma_3 = s_h$ and $\sigma_1 = s_v$ for the more steeply (60°) dipping, extensional faults, and $\sigma_3 = s_h$ and $\sigma_1 = s_H$ for shallowly dipping (30°) contractional faults.

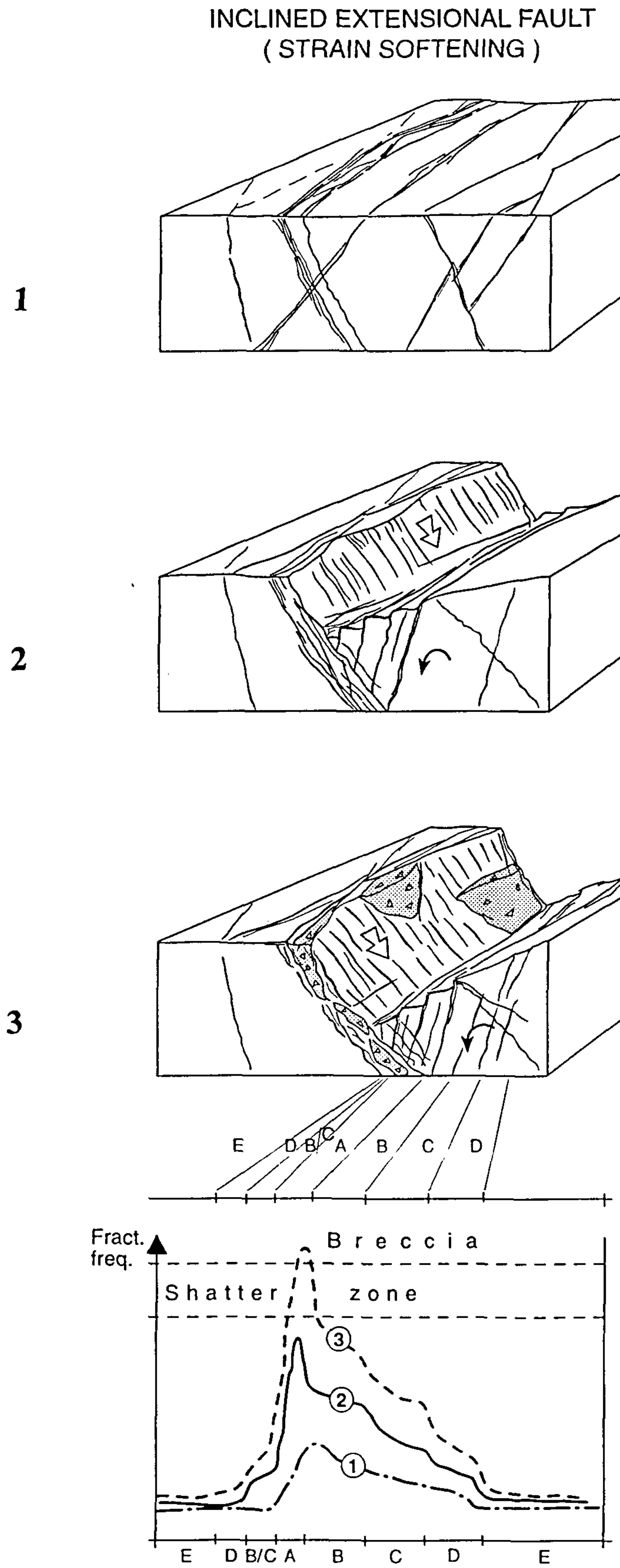


*Figure 28A. Three-stage scheme for development of a **inclined master fault** during **strain softening**, and associated schematised fracture profiles.*

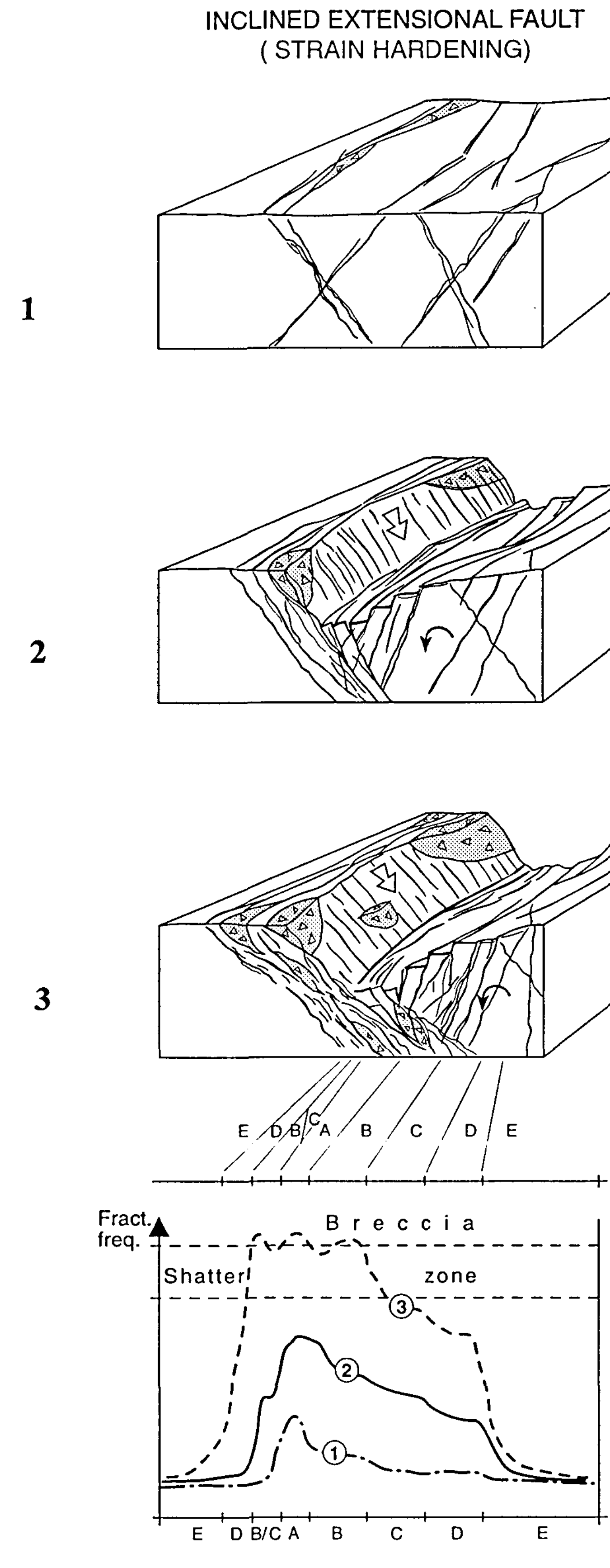


Figure 28B. *Three-stage scheme for development of a* ***inclined master fault*** *during* ***strain hardening****, and associated schematised fracture profiles.*

7. CONCLUSION

(1) Lineaments can be regarded as zones in the crust with enhanced fracture intensity (fracture zones) which have a distinct architecture. Fracture lineaments may be divided into three main types, namely:

- (i) *inclined fault zones*, which can be shown to offset lithological or other markers to a significant extent
- (ii) *steep fault zones*, which can be further subdivided into:
  - (a) master fault zones, with significant offset of markers,
  - (b) immature fault zones, which show minor or no apparent displacement, but contain breccia,
- (iii) *joint zones*, which are without displacement and breccia, or where criteria for identifying significant displacement are not available.

These structures are found both in metamorphic and sedimentary rocks.

(2) The spatial distribution of fractures varies in a systematic fashion, where the common fracture network can be ascribed to:

- (i) The *central part* of the lineament, 0-50-m wide, which consists of a dense network of short fractures and fault rocks (*sub-zones A-B*), and where secondary minerals are common, sealing most fractures.
- (ii) The *marginal part* (*sub-zones C-D*), commonly 20-50-m wide, revealing lower fracture intensity, pronounced lineament-parallel fracturing, and absence of clogging minerals.
- (iii) The *distal part* (*sub-zone E*; < 250 m) .
- (iv) The *general background fracturing*.

(3) Steep joint/fault zones reveal a more symmetric fracture pattern than inclined fault zones (normal and reverse), which tend to have asymmetric patterns with increased strain in the hangingwall.

(4) Fracture zones in metamorphic rocks typically show a more symmetric fracture distribution, more distinct zonation, and wider deformation zone than fracture zones in sedimentary rocks.

(5) Fracture systems have significant influence on conductivity in both metamorphic and sedimentary rocks: Lineament-perpendicular conductivity in fracture zones can be regarded as potential low in the central part, high in the marginal part (especially zone C), and intermediate in distal parts. The lineament-parallel symmetry of fracture sub-zones suggests that conductivity along the fracture zone in general is uniform.

(6) Several parameters have been applied to generate fracture profile types which are characteristic for each lineament type (master joint and fault zones), including lithology, depth of deformation (PT-conditions), strain intensity and strain rate and the nature of the actual local and far field stress systems. These influences are taken into account in a model to explain the different lineament architectures. Fracture distribution and orientation, fracture types, deformation products, and strain intensity are applied as variables, and used to determine fracture environment, stage of development and influence of strain-hardening and strain-softening.

8. ACKNOWLEDGEMENT

The authors would like to express their gratitude to the following enthusiastic group of geologist and geology-students participating in the work on fracture systems: S. Berg, M. Fejerskov, H. Henriksen, A. Lothe, K. Lyslo, E. Midtbø, A.K. Midtgård, A. Myrvang and E. Skurtveit.

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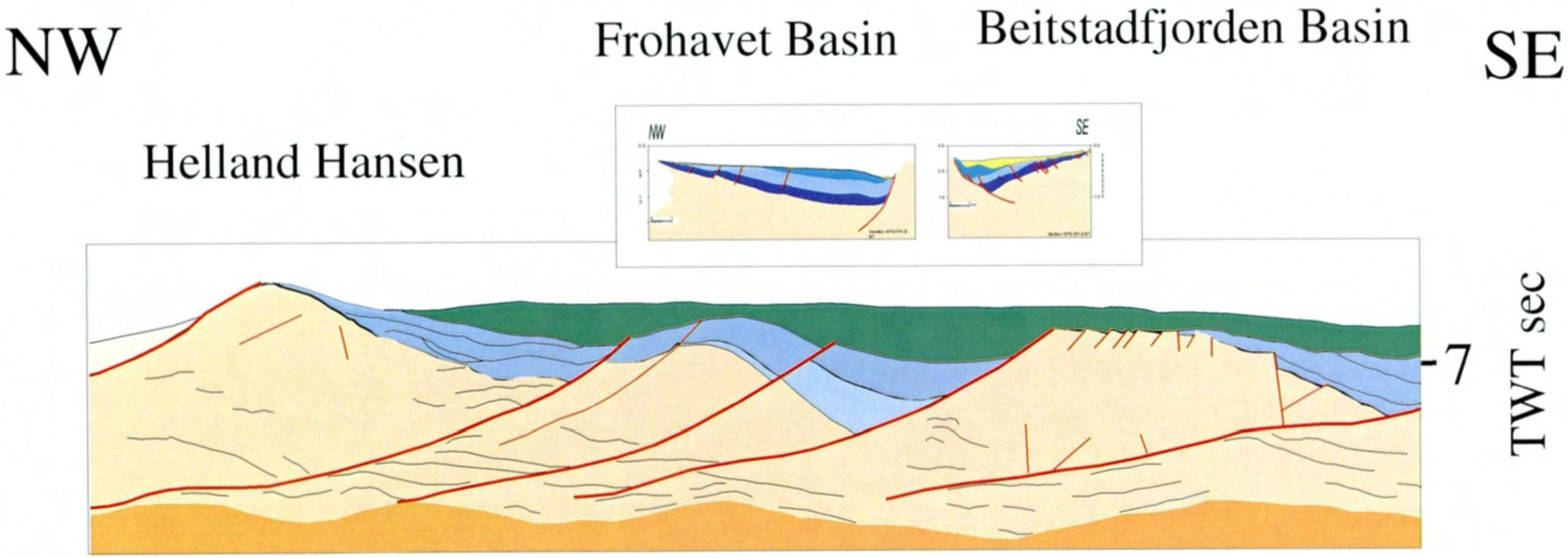

Jurassic Basins: Size and Geometry