Using mechanical models to investigate the controls on fracture geometry and distribution in chalk

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Abstract: Chalk is an important reservoir rock. However, owing to its low permeability, fractures are key to producing hydrocarbons from chalk reservoirs. Fractures in chalk usually form one of three geometric patterns: localized fractures (commonly concentric rings) developed around tips, bends and splays in larger faults; regularly spaced regional fracture sets; and fracture corridors comprising narrow zones of closely spaced parallel fractures. Localized fracture patterns are likely to give only local permeability enhancement; regional fracture sets and, especially, fracture corridors may provide long, high-permeability flow pathways through the chalk. Field mapping shows that both localized fracture patterns and fracture corridors often nucleate around larger faults; however, the fracture corridors rapidly propagate away from the faults following the regional stress orientation. It is therefore not necessary to know the detailed fault geometry to predict the geometry of the fracture corridors, although the fault density can help to predict the spacing of the fracture corridors. Mechanical modelling shows that while localized fracture patterns can form under normal fluid pressure conditions as a result of local stress anomalies around fault tips, bends and splays, fracture corridors can only form under conditions of fluid overpressure. Once they nucleate, they will continue to propagate until they either intersect another fault or the fluid pressure in them is dissipated.

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Cretaceous chalk is an important reservoir unit in the North Sea, and hosts a number of large fields in the Norwegian, Danish and UK sectors. However, the matrix permeability of chalk is generally very low (typically \(<1 mD\)), and, hence, fractures are often important in controlling fluid flow in the subsurface and are key to producing these fields economically (Megson & Hardman 2001). Fortunately, chalk is often heavily fractured, both in the subsurface (e.g. Ekofisk: Troublanc et al. 2005; Kraka: Jørgensen & Andersen 1991; Valhall: Bauer & Trice 2004) and in outcrop. Therefore, understanding the distribution, orientation and connectivity of the fractures is important to produce the fields efficiently.

Aims and objectives

In this study we use mechanical modelling techniques to examine the controls on fracture development and resulting geometry in chalk, and compare these results with fracture patterns observed in two chalk outcrops from southern and NE England. In particular, we will examine the relationship between fractures, fracture corridors and larger scale faults, and the impact of fluid pressure on fracture development. We will use elastic dislocation and finite-element models to investigate the in situ stress patterns developed around larger faults under different conditions of deformation and compare these with the observed fracture patterns. The elastic dislocation models simply assume the chalk to be a continuum elastic material, and take no account of fluid pressure. However, the finite-element models allow us to incorporate the effects of fluid overpressure, leading to dilation of the faults and fractures. We model two end-member cases of overpressured chalk: a case with a permeable host rock, in which the fluid overpressure results in a decrease in the effective horizontal stress; and a case with an impermeable host rock, in which the pore pressure remains constant but an overpressured fluid is injected into the faults and fractures, generating an outward pressure on the fracture wall.

The aim of these models is to derive rules and methods for predicting the fracture geometry based on the conditions of deformation and the pre-existing structure. This demonstrates the importance of fully utilizing outcrops and modelling to...
enhance fracture understanding, and to reduce the risks associated with predicting fractures in the subsurface.

NB: In this paper we use the term fracture to refer to a discontinuity in the rock caused by brittle failure with small or no offset. It thus includes joints (Mode 1 dilatant fractures with no fill), veins (Mode 1 dilatant fractures filled with crystalline cement) and Mode 2 shear fractures with very low displacement (typically <1 cm). We use the term fault to refer to discontinuities with a large shear offset of 1 m or more.

Previous studies

There is an extensive body of previous research assessing the distribution and geometry of brittle fractures observed in outcrop in a variety of different lithologies, from sandstones to basalts and granites (e.g. Ameen & Cosgrove 1990; Koestler & Ehrmann 1991; Rawnsley et al. 1992; Ameen 1995; Gross & Engelder 1995; Gillespie et al. 2001; Rijken & Cooke 2001; Richard et al. 2002; Belayneh 2004; Peacock 2004; Belayneh et al. 2007; Gross & Eyal 2007). However, the fracture geometries developed in many of these studies are remarkably similar, and there are several key geometric patterns that recur in different lithologies and settings. Three of these geometric patterns, often recognized in chalk, are regularly spaced regional fracture sets, fracture corridors and local concentric rings or arcs.

Regularly spaced regional fracture sets comprise long, parallel, regularly spaced fractures with a consistent strike across a large area. This strike generally reflects a regional tectonic in situ stress field, although it may reflect a more local stress field caused by flexure on the outer arc of a large fold (Pollard & Aydin 1988; Lorenz et al. 1991), for example. Fracture sets are often restricted within a single stratigraphic unit (typically a brittle bed surrounded by more ductile layers: Gross & Engelder 1995; Rijken & Cooke 2001; Peacock 2004), and the spacing of the fractures is generally proportional to the thickness of that unit, so that the layer-bound fracture density in thin beds is much higher than in thick beds (Bai & Pollard 2000a, b). The fractures may be vertical or they may be inclined, forming conjugate pairs. Vertical fractures are generally interpreted to have formed as Mode 1 dilatant features, while inclined fractures are generally interpreted to have formed as Mode 2 shear features, although the displacement on them may be very small (a few mm at most) (Schultz 2000; Ferrill & Morris 2003).

In some examples there is only one set of fractures; this is likely to generate a highly anisotropic permeability. Examples include dolomite layers in the Monterrey Formation, California (Gross & Engelder 1995) and Carboniferous limestones in the Burren, Ireland (Gillespie et al. 2001). However, more commonly there are multiple fracture sets in different orientations. In some cases, there may be a dominant set of long, continuous fractures, and a subordinate set of shorter fractures that terminate against the longer set. In these cases the dominant set is inferred to have formed first, perpendicular to the minimum principal stress orientation, while the subordinate set formed later due to strain relaxation as the dominant fracture set dilated. Examples of this include joints in Lower Jurassic limestones in NE (Rawnsley et al. 1992) and SW England (Belayneh 2004; Peacock 2004), and Cretaceous limestones in the Halukim Anticline, Israel (Gross & Eyal 2007). In other cases we see multiple co-dominant fracture sets that cross-cut each other, forming rectangular, quadrilateral or triangular patterns. These are typically inferred to represent fractures that formed at different times, in different stress fields, where earlier sets were held closed either by normal stress or by cementation to allow later sets to propagate across them (e.g. Gillespie et al. 2001; Peacock 2004).

Fracture corridors (sometimes called swarms) comprise relatively narrow zones (ranging from 10 cm to 5 m wide) containing closely spaced fractures that are parallel or subparallel to each other and to the zone as a whole (Segall & Pollard 1983; Odling 1997). Fracture corridors are typically long (tens to hundreds of metres), although the fractures within them may be shorter. They are separated by zones of much lower fracture density, so that overall fractures have a clustered distribution.

Examples of fracture corridors in outcrop are described by Laubach & Tremain (1994), Becker & Gross (1996), Odling (1997), Cacas et al. (2001), Barr et al. (2004), Al-Kindi (2006) and Belayneh et al. (2007). They are commonly found in chalk; examples include Upper Cretaceous chalks from the east of Paris, France (Richard et al. 2002), the Laegerdorf quarry near Hamburg in Germany (Koestler & Ehrmann 1991) and SE England (Ameen & Cosgrove 1990; Ameen 1995; Belayneh et al. 2007). In some cases they form damage zones around larger faults, or may represent early stages of incipient fault development (e.g. Pollard et al. 1982; Peacock 2002) but in many cases (e.g. Belayneh et al. 2007) they do not appear to be associated with any fault. Fracture corridors will clearly act as major conduits in the subsurface and, indeed, have been associated with mud-losses during driling (Dyke et al. 1992; Brown et al. 1999).

Olson (2004) showed, using numerical models of fracture propagation, that the fracture geometry can be dependent on the rate of fracture propagation
or, more accurately, on the subcritical index, which relates the propagation rate to the stress concentration at the fracture tip). He showed that a low subcritical index (slow fracture propagation) gives a relatively random fracture distribution (i.e., a random distribution of fracture spacings and lengths). A slightly higher subcritical index (moderate fracture propagation) gives rise to regularly spaced fractures; at higher propagation rates individual fractures propagate rapidly across the layer, without interacting with other fractures, and then develop a stress shadow that inhibits the growth of any other fractures within a certain distance, which is proportional to the fracture height. However, if the subcritical index is increased still further the fracture geometry switches to a clustered distribution with the development of fracture corridors. The reason for this is that, with very high subcritical index, the stress anomaly that develops at the tip of a propagating fracture is sufficiently large to induce other optimally oriented pre-existing microfractures around the tip to start propagating, and, hence, a narrow zone of subparallel propagating fractures develops.

Local fracture patterns develop around larger structures such as faults and folds, and particularly around fault tips, bends and splays. These localized fracture patterns comprise closely spaced parallel or concentric fractures that either loop around the fault or bend into it and truncate against it. Fractures from regional fracture sets are often subverted, changing orientation to fit the local pattern, although the fracture density near the fault tends to be higher than in the surrounding rock. The scale of these fracture patterns can vary considerably, from the cm-scale to the 100 m-scale, depending on the scale of the larger structure with which it is associated. Large-scale changes in orientation of fractures (more than several hundred metres) was recorded by Rawnsley et al. (1992) around a splay point in a 150 m-throw normal fault (the Peak Fault) in Jurassic limestone exposed at Ravenscar, NE England. Smaller scale localized fracture patterns are observed around a series of cm- to m-scale throw strike-slip faults in Jurassic Limestone at Nash Point, South Wales (Rawnsley et al. 1998; Bourne & Willemse 2001).

These patterns are usually interpreted to form due to the local perturbations in the in situ stress field that can develop around faults, and especially around fault bends, tips and splays (Segall & Pollard 1980; de Joussineau et al. 2003). The fracture density in these localized patterns is often high (up to one fracture per cm), but as they have a restricted lateral extent they will probably cause only a local increase in permeability, and will have little impact on long-distance fluid flow as they are unlikely to be connected.

**Outcrop analysis of fractures in chalk**

The aim of this study is to identify which of the fracture patterns described above can be recognized in chalk, to determine what are the key controls on the fracture patterns formed under different conditions and to develop methods of predicting the fracture geometry in chalk in the subsurface. The controls on fracture patterns will be studied using mechanical models, but in order to calibrate these models we must examine the fracture patterns in chalk outcrops. For this reason, field mapping for this study was carried out in two localities where chalk crops out on the coast of England: Pegwell Bay in Kent, SE England; and Flamborough Head in Yorkshire, NE England (Fig. 1). In both of the selected localities, the chalk is exposed in both cliff sections and wavecut platforms, allowing us to examine the geometry of the fractures in both cross-section and map view. Both localities also contain larger faults that affect the pattern of fractures. This allows us to evaluate the impact of faulting on fracture development.

It has been well documented that there is a marked contrast in properties between the chalks of southern and NE England, with the latter...
consistently denser and less porous than the former, and also typically having higher unconfined compressive strength (UCS), tensile strength and Young’s modulus (e.g. Carter & Mallard 1974; Bell 1977; Clayton 1983; Bell et al. 1999). This distinction has been borne out by this study: the mean porosity of 11 chalk samples collected from Pegwell Bay is 43.2%, while three samples collected from Flamborough Head have porosities ranging from 8 to 17% — although Bell et al. (1999) suggested that porosity in the NE chalks can be up to approximately 25%. Carter & Mallard (1974) suggested that this distinction is due to a greater maximum depth of burial for the NE chalks; comparison of the measured porosities against porosity–depth trends for normally compacted chalk from the North Sea (Mallon & Swarbrick 2002) suggest a maximum depth of burial of the Pegwell Bay chalk of around 500 m, compared with approximately 2 km for the Flamborough Head chalk (Fig. 2). It should be noted, however, that other studies (e.g. Clayton 1983) suggest that this contrast in porosity between the chalk of southern and NE England may be a diagenetic effect, unrelated to burial history. It is also worth noting that the chalk in the producing fields of the North Sea tends to have similar porosity to the chalks in southern England, even though these are currently buried to depths of >2 km; in these cases it is thought that the hydrocarbons in the chalk have inhibited normal compaction (Mallon & Swarbrick 2002).

Fig. 2. Compilation of porosity–burial depth relationships for North Sea chalks (data from Mallon & Swarbrick 2002), showing the measured porosity values for samples taken from Pegwell Bay and Flamborough Head.
Pegwell Bay, SE England

Pegwell Bay lies on the southern coast of the Isle of Thanet, an approximately 6 × 8 km peninsula in NE Kent that represents the topographical expression of the east–west-oriented Thanet anticline. The Thanet anticline is an asymmetric chalk cored anticline with a steep southern limb (Shepherd-Thorn 1988). It is thought to have formed passively by draping and differential compaction of the chalk over a pre-existing basement fault scarp (Ameen 1995). Mansy et al. (2003) suggested that this basement fault was subsequently reactivated by late Cretaceous–Tertiary inversion, although it has not propagated up through the chalk.

Pegwell Bay lies on the steeper southern limb of this anticline, which dips at approximately 7° southwards. Chalk from the Newhaven Chalk Member of the Upper Chalk Formation (Santonian–Campanian age) is exposed along an approximately 500 m east–west-oriented coastal section. The chalk here is exposed both in 20–30 m-high cliffs and in a horizontal wave-cut platform up to 80 m wide, which allowed us to map the structures in the chalk in both in cross-section and plan view (Fig. 3).

This outcrop is cut by a series of small, regularly spaced, subparallel north–south– to NNW–SSE-striking faults, typically around 10 m apart (marked in red on Fig. 3). These faults dip both east and west, and are generally steep, with dips ranging from 60° to vertical. In the cliff section they are seen to have normal displacements of up to 20 cm, but the slickenside orientation suggests that these faults have also undergone a significant amount of strike-slip deformation. This is difficult to quantify, although the offset of a small east–west-striking fault in the wavecut platform suggests left-lateral displacement of approximately 1 m. This is thought to have occurred in response to the Neogene episode of NE–SW extension (Souque et al., pers. comm.).

The outcrop is also heavily fractured away from the faults. Most of the fractures observed in the field contain no evidence of cataclasis and are interpreted as Mode 1 dilatant fractures. They can be subdivided into three main types on the basis of geometry:

- **Through-going background fractures** – these isolated fractures do not intersect the faults, although they are often truncated at bed boundaries. They are relatively evenly spaced but have a low density (typically spacing of c. 25 m). They have steep dips and form two distinct sets, one striking north–south and a second striking NW–SE. These orientations are consistent with the inferred regional stress orientation during the late Cretaceous and the Neogene inversion episodes, respectively (Vandycke 2002).
- **Fracture corridors** – these comprise narrow zones, typically 1–5 m wide, containing a high density of subparallel fractures (marked in green in Fig. 3). Within a fracture corridor, the typical fracture spacing varies between 2 and 20 cm; the individual fractures within the corridor may be short and are all unfilled. The fracture corridors show a fairly consistent NW–SE strike in all parts of the Thanet anticline; however they sometimes show local changes in orientation, particularly where they intersect a fault. Often they appear to abut against a fault bend or tip, either in cross-section or in plan view. These fracture corridors are inferred to have nucleated next to the fault in response to the localized stress perturbation generated by the fault bend or tip and to have subsequently propagated away from the fault, changing orientation to reflect the regional stress orientation. It is possible that all of the fracture corridors formed in this way, but in many cases the fault at which they nucleated is not exposed.
- **Concentric fracture rings** – these localized fracture patterns comprise concentric rings of closely spaced fractures, typically a few metres in diameter. They mostly occur around bends in the faults. Typical fracture spacing within the ring patterns again varies between 2 and 20 cm, and the fractures are also unfilled. The rings are not necessarily circular but may be oval or elliptical in shape, and usually form partial rather than complete rings; unlike the fracture corridors, these fractures are always short and do not propagate far from their origin.

It is likely that the fracture corridors will have the greatest impact on fluid flow through the chalk: the background fractures, although often long, are too few to transmit significant volumes of fluid, whereas the concentric fracture rings are localized and isolated. The fracture corridors, however, contain a sufficient density of interconnected fractures and extend over sufficiently long distances to act as major fluid flow pathways. Furthermore, since they appear to nucleate on the faults, they are likely to provide connectivity between them. A key aim of the modelling work will, therefore, be to understand the controls on the nucleation and geometry of these fracture corridors.

Flamborough Head, NE England

Flamborough Head is an approximately 6 × 3 km headland jutting into the North Sea on the east coast of Yorkshire. It lies along the approximately
east−west-trending boundary between the Jurassic−
Lower Cretaceous Cleveland Basin, to the north,
and the relatively stable Market Weighton Block
to the south. The hinge zone between these two
features is formed by the Howardian Hills−
Flamborough Fault Belt, an approximately 10 km-
wide east−west-trending fault zone that passes
directly underneath Flamborough Head (Kirby &
Swallow 1987). The Upper Cretaceous chalk was
deposited as a post-rift blanket deposit over both
the Cleveland Basin and the Market Weighton
Block, but subsequent tectonic activity has caused
reactivation of the underlying faults and several
episodes of deformation in the overlying chalk
(Starmer 1995).

Strata from the Welton, Burnham and Flambor-
ough Chalk formations (Turonian−Santonian age)
are exposed along the north coast of Flamborough
Head, and are accessible in three bays (Thornwick
Bay, North Landing and Selwicks Bay), where
they are exposed both in cliff section and wavecut
platforms (Fig. 4). Selwicks Bay also contains a
large (20 m-throw) normal fault formed during
one of the late tectonic reactivation phases, charac-
terized by an approximately 10 m-wide core of
breccia and crystalline calcite veins up to 30 cm
thick (Starmer 1995). The chalk is heavily fractured
at all localities around Flamborough Head (Peacock
& Sanderson 1994).

Fracture corridors are also observed in Flam-
borough Head, although only in Selwicks Bay. However, these differ from those in Pegwell Bay
in two respects: first, they are narrower (typical-
ly 30−60 cm wide) and, secondly, they mostly

Fig. 3. Aerial photograph showing the cliff section and wavecut platform at Pegwell Bay, Kent. In the annotated
photograph, the red lines indicate the north−south- to NNW−SSE-oriented strike-slip faults, and the green shading
indicates fracture corridors.
comprise veins filled with crystalline calcite cement. Within the corridors, the vein spacing is typically 0.5–3 cm and the veins can be up to 5 mm thick, although most are about 1–2 mm thick (Fig. 5a).

In Thornwicks Bay and North Landing, the majority of fractures belong to regional fracture sets, comprising regularly spaced open, uncemented fractures with a consistent orientation across the study area. Fracture spacings vary from 10 cm to 2 m. Often two to three sets of oblique, cross-cutting regional fractures can be seen, forming a rectangular or triangular pattern (Fig. 5b). In the cliff section, we see that there are two distinct styles of regional fractures that appear to be controlled by stratigraphy (see Fig. 6):

- The lowermost Welton Formation is characterized by large, inclined fractures, typically dipping at around 60°, and often forming conjugate pairs. These fractures are generally long, cutting through bed boundaries, and relatively widely spaced (1–3 m apart). They appear to have formed by Mode 2 shear failure, although the shear displacement is very small (<2 mm).
The Burnham Formation is characterized by smaller, more closely spaced vertical bed-bound fractures (typically 20–50 cm apart). These are interpreted to have formed by Mode 1 tensile failure.

Outcrop studies: review

There are a number of key differences between the fracture patterns observed in the chalk at Pegwell Bay and those observed at Flamborough Head:

- The majority of the fractures observed at Pegwell Bay form fracture corridors represented by narrow zones containing closely spaced subparallel fractures. At Flamborough Head, however, fracture corridors are only locally important; in most areas the majority of fractures are single-stranded, regularly spaced fractures.
- At both Flamborough Head and Pegwell Bay, the fracture corridors appear to nucleate around larger faults. However, at Pegwell Bay these larger faults are strike-slip faults with throws of about 1 m, spaced at around 30–40 m intervals along the outcrop, while at Flamborough Head the fracture corridors appear to have nucleated exclusively around a normal fault with an approximately 20 m throw at Selwicks Bay. The fractures that comprise...
the fracture corridors at Pegwell Bay are unfilled, whereas at Flamborough Head the fractures that comprise the fracture corridors contain crystalline calcite cement, as does the fault around which they nucleate. The regional fractures at Flamborough Head are mostly unfilled.

- At Pegwell Bay, the fracture corridors are likely to form the principal fluid-flow pathways through the chalk. However, at Flamborough Head, the restricted distribution and calcite cement fill in the fracture corridors suggests that they will be less important in controlling long-distance fluid flow; this is likely to occur predominantly along the unfilled regional, regularly spaced fractures.
- Almost all of the fractures observed on the south coast of England are oriented normal to bedding, and are thus interpreted as predominantly Mode 1 dilatant fractures (although they may also have a small component of strike-slip shear displacement). However, a significant number of the regional fractures at Flamborough Head are inclined, often forming conjugate sets, and are thus interpreted as having formed by Mode 2 shearing (although they have negligible throw). These fractures are thought to have formed at the same time as the Mode 1 dilatant fractures.

Fig. 5. Photographs showing: (a) an example of a vein corridor from Selwicks Bay; and (b) cross-cutting regional fractures exposed on the wavecut platform at Thornwick Bay.
at the same locality; the variation in failure mode appears to have been largely controlled by a contrast in mechanical properties between different stratigraphic units.

There are two potential reasons for the contrast in fracture styles and geometries, and thus in the inferred fracture development, described above:

- It may result from a difference in mechanical properties between the chalk in southern England and the chalk in NE England due to the contrast in porosity (Bell 1977; Bell et al. 1999). This may be related to different burial histories (Carter & Mallard 1974) or to different diagenetic processes (Clayton 1983).

- It may result from different regional structure and tectonic history. The chalk at Flamborough Head was deposited over the boundary between a Mesozoic basin to the north (the Cleveland Basin) and a high to the south (the Market Weighton Block). This boundary was reactivated several times during the various Tertiary deformation episodes following chalk deposition, as both a strike-slip and a compressional feature (forming the Howardian Hills–Flamborough Fault Belt). The fractures at Flamborough Head are inferred to have formed during one of these deformation episodes, as is the 20 m-throw normal fault at Selwick’s Bay. By contrast, the chalk at Pegwell Bay was deposited on the London–Brabant structural high, where there was relatively little inversion; any inversion that did occur on the underlying basement fault was accommodated in the chalk by folding rather than faulting (Mansy et al. 2003). Therefore at Pegwell Bay, unlike at Flamborough Head, we see no faults with throws greater than a few metres.

In both cases it appears that larger faults are key to controlling the development of fracture corridors, although the size of these larger faults varies. At Pegwell Bay, the fracture corridors tend to nucleate around tips, bends or splays in the metre-scale strike-slip fault, but then propagate outwards, oblique to the faults, typically until they intersect an adjacent fault. At Flamborough Head, the clustering of the observed vein corridors around the 20 m-throw normal fault at Selwick’s Bay, combined with the presence of crystalline calcite cement in both the fault and the vein corridors (but not in the regional fractures at North Landing and Thornwicks Bay), suggests that the vein corridors were formed by a high-pressure calcite-rich fluid that was migrating along the fault. They were thus able to propagate a short distance (c. 50–100 m) from the fault before fluid leak-off or viscosity caused the fluid pressure at the tip of the veins to drop below the level required to drive propagation.

Fig. 6. Photographs from North Landing showing the contrast in style between the regional fractures in the Welton Chalk, characterized by through-cutting conjugate inclined fractures, and the Burnham Chalk, characterized by vertical bed-bound fractures. The former are inferred to have formed as Mode 2 shear fractures and the latter as Mode 1 dilatant fractures.
In the following sections we will use numerical modelling techniques to investigate the stress patterns formed around larger scale faults under different conditions and in chalk with different mechanical properties. The aim is to derive rules for predicting fracture distribution and orientation around faults in the subsurface.

Elastic dislocation modelling and the impact of mechanical properties

The elastic dislocation model is a mechanical model that calculates the equilibrium displacement, strain and stress fields in an elastic medium resulting from localized displacements on one or more planar fault (Okada 1992; Ma & Kusznir 1993). It therefore differs from finite-element models in that it requires the fault displacements to be specified as initial boundary conditions rather than calculating these as a result of stress boundary conditions. The results can be used to predict the distribution and geometry of Mode 1 fractures: these will tend to form in areas where the minimum stress is most tensile (or least compressive), and will be orientated perpendicular to the minimum principal stress. This technique has been successfully used to predict fracture distribution and geometry in a number of previous studies (e.g. Healy et al. 2004).

In this study we use elastic dislocation models to calculate local stress fields around different geometric features (e.g. tips, bends and relay zones) in metre-scale vertical strike-slip faults of the type mapped in Pegwell Bay. We will compare the results with the fracture patterns observed around similar features in outcrop. We will also run several series of models with identical geometry but varying mechanical properties or conditions of deformation to investigate the impact of these on the resulting fracture geometry.

Derivation of the elastic dislocation model

In this study we use the elastic dislocation formulae developed for strike-slip faults by Okada (1992). The Okada formulae give the full 3D displacement vector and strain tensor at any point in a 3D space around one or more planar strike-slip fault. They take account of the effect of the earth’s surface (which is modelled as a free surface), but assume the earth is a homogeneous medium extending infinitely in other directions (i.e. they are suitable for modelling deformation in strata that are fairly uniform on the scale of the faults, such as the thick chalk sequences we are modelling here, but would be less suitable for layered or heterogeneous strata). The faults are modelled as rectangular planar discontinuities of uniform displacement. There is no limit to the number of discontinuities, so that faults with lateral variations in displacement can be modelled by multiple adjacent discontinuities of different displacement.

Although the Okada formulae are fully 3D, in this study we use them to calculate the localized stress field on a horizontal section through the structures we are modelling. We use the 3D Hooke’s law to calculate the overall stress tensor by combining this local stress field with a regional stress field (comprising a vertical effective lithostatic stress, a uniform horizontal confining stress and a horizontal differential stress representing a tectonic regional stress). We then project the 3D stress tensor onto the horizontal section to obtain the maximum and minimum horizontal stress, and the orientation of the maximum horizontal stress.

Mode 1 fractures will form where the fluid pressure exceeds the minimum principal stress plus the tensile strength of the rock (where compressive stress is positive). They are therefore most likely to form where the minimum principal stress, \( \sigma_{\text{min}} \), is low, and will preferentially form normal to the minimum stress axis. In this study we model the faults as vertical planar discontinuities with a purely horizontal displacement. As a result the principal stress axes will remain approximately orthogonal to the vertical, and the fractures will be vertical and strike parallel to the maximum horizontal stress \( \sigma_{\text{h,max}} \).

By normalizing the minimum horizontal stress plus the tensile strength of the rock against hydrostatic pressure, we can also define fracture resilience, \( R \), as:

\[
R = \frac{\sigma_{\text{h,min}} + T_0}{\rho_w g Z}
\]

where \( T_0 \) is the tensile strength of the rock, \( \rho_w \) is the density of water, \( Z \) is the depth and \( \sigma_{\text{h,min}} \) is the absolute (not effective) minimum horizontal stress. The lower the value of \( R \), the more likely fracturing is to occur; where \( R < 1 \) the effective stress under normal fluid pressure conditions is tensile and we would expect fractures to form without overpressure.

Model geometry

In this study we are particularly interested in the local stress perturbations around tips, bends and kinks in metre-scale strike-slip faults similar to those mapped at Pegwell Bay. All of the models therefore incorporate a single vertical fault 60 m long by 100 m high. Six different geometric features were modelled (Fig. 7):

- the tip of a planar fault;
- a single bend in the centre of the fault;
a soft-linked releasing relay zone (i.e. a zone of lateral fault offset where the offset segments are unconnected) located in the centre of the fault;

a hard-linked releasing relay zone (i.e. a zone of lateral fault offset where the offset segments are connected) located in the centre of the fault;

a triangular kink in the centre of the fault;

a trapezoidal kink in the centre of the fault.

The relay zones and kinks are given dimensions to be typical of those observed in the field. We also ran a series of models varying the size and aspect ratios of the triangular kink model. For each model, we calculate the local stress field on a horizontal planar section extending $10 \times 10$ m around the feature of interest.

A planar fault in an elastic medium will adopt an elliptical displacement profile in response to a
Fig. 7. Continued.
regional stress, and the maximum displacement on the fault, \( D_{\text{max}} \), can be calculated using the equation:

\[
D_{\text{max}} = \frac{\pi L \sigma_{\text{h, max}} \sin \theta \cos \theta (1 - \mu \cot \theta)}{8E\gamma}
\]

where \( L \) is the fault length, \( \mu \) the fault friction coefficient, \( \theta \) the angle between the fault and the \( \sigma_{\text{h, max}} \) orientation, \( E \) is the Young’s modulus, and \( \gamma \) is a parameter based on Poisson’s ratio (Welch et al. 2009). In these models, the maximum stress was oriented at 75° to the faults to give a left-lateral displacement (this is consistent with the regional stress field inferred by Vandycke 2002). To simulate the displacement gradient in the fault tip model, the fault was subdivided into a series of vertical panels, 0.6 m wide near the tip and 6 m wide away from the fault tip. Each panel was then given a displacement consistent with an elliptical displacement profile, with a maximum displacement in the centre given by equation (2).

Note that the displacement given by equation (2) represents the maximum elastic strain that can build in the rock. Over geological time, non-elastic processes such as creep and pressure solution will lead to the slow accumulation of non-elastic displacement on the fault. The present-day fault displacement measured in the field (c. 1 m) is therefore much greater than the elastic displacement predicted by equation (2). We can check whether the predicted displacement on a fault is in equilibrium with the regional stress field by looking at the stress orientation adjacent to the fault – for a planar frictionless fault, the principal stresses should be orthogonal to the fault (except around tips, bends or kinks, where there may local stress perturbations caused by the feature in question). In the base-case models, we set the fault friction coefficient to zero. However, we also ran a series of models in which the fault was frictional; friction on the fault allows it to support a shear stress, and reduces the displacement.

It should be noted that Bourne & Willemse (2001) showed that for a network of intersecting faults, interference between the faults means that the assumption of an elliptical displacement profile and the maximum displacement given by equation (2) are not valid. However, the spacing of the faults in Pegwell Bay is large compared with the scale of the features that we are modelling, so we have ignored interactions between faults in these models.

**Mechanical properties**

There is an extensive body of literature detailing the results of laboratory measurements of different mechanical properties on samples of chalk taken from outcrop or boreholes, including Carter & Mallard (1974), Bell (1977), Mortimore (1979), Jones et al. (1984), Clayton (1990), Clayton & Saffari-Shooshtari (1990), Flexer et al. (1990), Kronieger (1990), Mimram & Michaeli (1990), Mortimore & Fielding (1990), Bell et al. (1999), Katz et al. (2000), Gommesen & Fabricius (2001), Bowden et al. (2002) and Mortimore et al. (2004). These studies all show that the key control on mechanical properties of chalk is the porosity: as this decreases with burial and compaction, the tensile strength, UCS and Young’s modulus of chalk will all increase (the relationship between porosity and Poisson’s ratio is less clear so we have used a constant value for Poisson’s ratio in this study).

We cross-plotted the results from these various studies against the porosity of the chalk samples in each case to derive a porosity–mechanical property relationship for tensile strength, UCS, Young’s modulus, friction coefficient and cohesion. We then compared these trends with the chalk porosity–depth trend given by Mallon & Swarbrick (2002) (Fig. 2) to estimate typical mechanical property values for chalk at different depths of burial (see Table 1). For the base-case model, we have assumed a depth of 2000 m, since this is more typical of the depth of producing North Sea chalk fields. However, we have also run models for deformation at burial depths of 500 m and 5000 m to study how different depths of deformation may affect the distribution and orientation of fractures. It is important to note that there are several different factors that vary with depth, and which may affect the distribution and orientation of the fractures (Table 1):

- the Young’s modulus of the chalk;
- the tensile strength of the chalk;
- the magnitude of the effective vertical stress, \( \sigma_v' \), and hence the horizontal stress;
- the depth of the controlling fault below the free surface;
- the equilibrium displacement of the fault, which will change as a result of variations in the above factors (see equation 2).

**Results**

The results from these models are illustrated in Figures 7–10. The colours on these diagrams indicate fracture resilience, \( R \) (calculated using equation 1). Hot colours indicate areas of lower resilience where fractures are most likely to form. The black lines indicate \( \sigma_{\text{h, max}} \) orientation (i.e. the likely fracture orientation). Different sets of models were run to investigate the impact of different fault geometries, friction coefficient and depth of burial.
on the resulting stress patterns. In Figure 7 we compare the stress patterns predicted by the models with the fracture patterns observed in outcrop at Pegwell Bay.

**Stress fields around different fault bends.** The distribution of fracture resilience and \( \sigma_{h_{\text{max}}} \) orientation around the six modelled geometric features are shown in Figure 7. For comparison, we also show the fracture patterns developed around faults with similar geometries as mapped in Pegwell Bay. The results can be summarized as follows:

- **Fault tip** (Fig. 7a) – as would be expected, there is a reduction in fracture resilience on the tensile side of the fault tip (the far side in this model) and an increase on the compressive side (the near side). On the tensile side of the fault tip, \( \sigma_{h_{\text{max}}} \) is oblique to the fault. We do, indeed, see oblique fractures developed around the tips of several of the faults mapped in Pegwell Bay.

- **Fault bend** (Fig. 7b) – with left-lateral fault displacement, this forms a releasing bend with a tensile stress (\( R < 1 \)) around the eastern limb. As would be expected, we see a large reduction in the fracture resilience on the outside of the bend but, more surprisingly, we also see a reduction in the fracture resilience on the inside of the bend. The \( \sigma_{h_{\text{max}}} \) orientation forms a complex semi-circular pattern around the fault bend. We do, indeed, see heavy fracturing around many of the fault bends observed in Pegwell Bay, with complex patterns developed (although, in many cases, these are made more complex by the presence of splay faults).

- **Soft-** (Fig. 7c) and **hard-linked relay** (Fig. 7d) – with left-lateral fault displacement, these left-lateral relay zones are releasing bends. As would be expected, they generate local tensile stresses that reduce the fracture resilience, not just

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| Depth (m) | Porosity (%) | Tensile strength (MPa) | Young’s modulus (GPa) | Poisson’s ratio | \( \sigma_{v}’ \) (MPa) | \( \sigma_{h_{\text{max}}}’ \) (MPa) | \( \sigma_{h_{\text{min}}}’ \) (MPa) | Fault slip (m) |
|-----------|--------------|------------------------|-----------------------|----------------|----------------|----------------|----------------|--------------|
| 500       | 45           | 0.5                    | 5                     | 0.3            | 0.76           | 0.34           | 0.31           | 0.0002       |
| 2000      | 17           | 2.5                    | 15                    | 0.3            | 14.5           | 6.54           | 5.89           | 0.0012       |
| 5000      | 2            | 2.5                    | 21                    | 0.3            | 57.0           | 25.71          | 23.14          | 0.0034       |

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**Fig. 8.** Results from the elastic dislocation model for different geometries of triangular kink. In (a) the kink is larger, but maintains the same aspect ratio as the base case (Fig. 7e); in (b) the kink is the same width as the base case but the aspect ratio is reduced. Colour shading indicates fracture resilience, with warm colours indicating areas where fractures are most likely to form, and cold colours indicating areas where fractures are least likely to form. Lines indicate \( \sigma_{h_{\text{max}}} \) orientation.
within the relay zone but over a large area extending out from the relay zone for 2–3 times its size. $\sigma_{h\text{max}}$ forms an ovoid pattern around the relay zone, so we might expect to see concentric fractures centred on this zone. The stress fields generated for the soft- and hard-linked relays are almost identical, suggesting that the development of a through-cutting fault across the relay zone has little impact on the local stress field.

- **Triangular kink** (Fig. 7e) – this geometry comprises adjacent releasing and restraining bends.

**Fig. 9.** Results from the triangular kink elastic dislocation model for frictional faults. (a) shows a fault with friction coefficient 0.025 and a $\sigma_{h\text{min}}/\sigma_{h\text{max}}$ ratio of 0.9, giving 0.0005 m displacement; (b) also shows a fault with friction coefficient 0.025 but a $\sigma_{h\text{min}}/\sigma_{h\text{max}}$ ratio of 0.28, giving 0.0012 m displacement. Colour shading indicates fracture resilience, with warm colours indicating areas where fractures are most likely to form, and cold colours indicating areas where fractures are least likely to form. Lines indicate $\sigma_{h\text{max}}$ orientation.

**Fig. 10.** Results from the triangular kink elastic dislocation model for a fault forming at depths of: (a) 500 m and (b) 5000 m. The Young’s modulus and tensile strength of the rock, the regional stress field, the hydrostatic stress and the fault displacement all vary with depth (see Table 1). Colour shading indicates fracture resilience, with warm colours indicating areas where fractures are most likely to form, and cold colours indicating areas where fractures are least likely to form. Lines indicate $\sigma_{h\text{max}}$ orientation.
A high compressive stress develops around the restraining bend and a high tensile stress develops around the releasing bend, which generates a complex pattern of $\sigma_{h\text{max}}$ orientation, with two ovoid 'stress rings' (one of which surrounds the releasing bend) separated by a narrow, convergent 'stress neck' that cuts through the restraining bend. The two stress rings develop in areas of reduced fracture resilience, while the stress neck, around the restraining bend, lies in an area of increased fracture resilience. In the outcrop examples we do see well-developed concentric rings of fractures developing around the releasing bend, as would be expected. However, we also see fracture corridors propagating away from the restraining bends; these are harder to explain because, although they are consistent with the predicted $\sigma_{h\text{max}}$ orientation, the models predict this to be an area of high fracture resilience where fractures are unlikely to form.

- **Trapezoidal kink** (Fig. 7f) – this geometry generates a similar stress pattern to that seen around the triangular kink, except that there is a greater separation between the tensile, concentric stress rings around the releasing bend and the compressive, convergent stress pattern through the restraining bend.

**Stress field as a function of kink geometry.** Increasing the size of the triangular kink appears to have little or no effect on the magnitude of the tensile stress or geometry of the stress field, although the size of the stress perturbation does increase in proportion to the size of the kink (Fig. 8a). This is probably because the magnitude of the force generated by the kink increases in proportion to the kink size, but so does the area over which this force acts, and therefore the net additional stress is independent of the kink size.

Reducing the aspect ratio of the kink reduces both the size of the stress perturbation and the magnitude of the tensile stress as the offset of the releasing and restraining components of the bend are reduced. However, $\sigma_{h\text{max}}$ orientation still follows the same general pattern, with concentric rings around the releasing bend and convergent stress lines cross-cutting the restraining bend (Fig. 8b).

These results are consistent with the observations from Pegwell Bay. The faults mapped here contain kinks and bends with various different sizes and aspect ratios. Typically, the size of the localized fracture patterns is proportional to the size of the fault bends and kinks, but the kinks and relay zones with lower aspect ratios tend to have less prominent fracture patterns (Fig. 3).

**Stress field as a function of friction on fault.** In the models described above, the faults are all assumed to be frictionless. As frictionless faults are unable to support a shear stress, away from the tips, bends and kinks, the $\sigma_{h\text{max}}$ orientation curves into the fault so that it is either parallel or perpendicular to it (see Fig. 7). However, if there is friction on the fault, it will be able to support a shear stress, and $\sigma_{h\text{max}}$ will not be orthogonal to the fault but inclined towards regional $\sigma_{h\text{max}}$ orientation. The equilibrium elastic displacement on the fault will also be reduced (note that the elastic displacement reflects the maximum elastic strain that can build up in the rock, and will be much lower than the total displacement built up over geological time, which will incorporate inelastic processes such as creep and pressure solution; however, it is the elastic displacement that will control the local stress field around the fault).

With a $\sigma_{h\text{min}}/\sigma_{h\text{max}}$ ratio of 0.9, even a small amount of friction gives a large reduction in the equilibrium displacement on the fault. This causes a significant reduction in both the size and the magnitude of the stress perturbation around the kink. In Figure 9a, which shows the stress pattern around a fault with a friction coefficient of 0.025, the elastic displacement on the fault is reduced to 0.0005 m, compared with 0.0012 m for a frictionless fault. The circular stress pattern disappears completely and we simply see a slight deflection in the regional stress field. In this case there is no tensile stress around the kink and, hence, no significant reduction in fracture resilience. We can increase the elastic displacement on a frictional fault by increasing the differential stress. In the model shown in Figure 9b, the fault also has a friction coefficient of 0.025, but the elastic displacement on the fault has been increased to 0.0012 m by reducing the $\sigma_{h\text{min}}/\sigma_{h\text{max}}$ ratio to 0.28. Increasing the elastic displacement increases the magnitude of the stress anomalies developed around the kink but has little effect on the $\sigma_{h\text{max}}$ orientation, which still mainly reflects the regional stress orientation. This suggests that the magnitude of the stress anomaly is proportional to the displacement on the fault but the $\sigma_{h\text{max}}$ orientation is mostly controlled by the amount of shear stress the fault can support, which is dependent on the fault friction coefficient. Note that the relationship between internal friction coefficient and porosity for chalk (Flexer et al. 1990) suggests an internal friction coefficient of around 0.3–0.4 at a depth of 2000 m. However, if the faults are dilatant due to fluid over-pressure there will be no friction acting on them – this scenario will be examined in the next section using finite-element models.

Most of the faults mapped at Pegwell Bay are surrounded by either parallel or perpendicular fractures (Fig. 3). This suggests that the faults were effectively frictionless during fault slip.
Stress field at different depths of faulting. As explained above, a number of parameters are depend-ent on the depth of burial of the chalk at the time of faulting: the mechanical properties (especially Young’s modulus and tensile strength) of the chalk; the magnitude of the effective lithostatic stress and regional horizontal stress; the distance of the free surface above the fault; and the equili-brium displacement on the fault. The fluid pressure used to normalize the fracture resilience, \( R \), will also vary with depth.

As Figure 10 shows, the stress perturbation around the kink has a consistent geometry and size regardless of the depth of faulting. However, the magnitude of the stress perturbation increases with depth. This may be due to several factors:

- At greater depths, the Young’s modulus of the chalk increases. This means that the stress will be greater for a given strain, so that if fault displacement remains constant, the magnitude of the stress perturbation around the kink will increase.
- At greater depths, the effective lithostatic stress will increase. Since both the mean and differential horizontal stress components are proportional to the vertical stress, the magnitude of the horizontal regional stress will also increase. This explains the increase in fracture resilience with depth away from the fault.
- At greater depths, the equilibrium fault displacement increases as the increase in maximum horizontal stress overrides the effect of increasing Young’s modulus (see equation 2). An increase in the equilibrium fault displacement will increase the magnitude of the stress perturbation around the kink.

The consistency of the stress field geometry at different depths of burial suggests that the geometry of local fracture patterns is not very sensitive to the depth of burial at the time of deformation. This is borne out by comparing the model results with the fractures observed at Pegwell Bay: as noted above, the models illustrated in Figure 7 were run assuming a depth of burial of 2000 m, while the fracture patterns observed at Pegwell Bay are likely to have formed at a depth of burial of around 500 m, but the two match very closely. However, the elastic dislocation models do not allow for dilatation of the fault planes, which may be likely at shallow depths of burial. The impact of this will be considered using finite-element models.

Discussion

The elastic dislocation models are able to explain most of the localized fracture patterns observed around fault tips, bends and splays in Pegwell Bay – especially the oblique fractures observed around fault tips and the concentric rings of fractures observed around releasing bends and relay zones in the faults. However, they are less able to account for the fracture corridors also mapped at Pegwell Bay (and likely to be the predominant control on fluid flow through this chalk).

Most of these fracture corridors tend to nucleate around bends, tips and kinks in the strike-slip faults; however, they then propagate outwards, oblique to the faults. They extend long distances away from the faults – typically 50–100 m, often connecting to adjacent faults and extending much further than the localized stress anomalies predicted by the elastic dislocation models. Furthermore, although the ‘stress necks’ developed around restraining bends and relay zones most closely match the geometry of the fracture corridors, these are areas of high fracture resilience where fractures are least likely to nucleate (see Fig. 7e).

However, these models all assume a normal fluid pressure gradient; the fluid pressure is not modelled explicitly but by reducing the vertical and horizontal stresses to take account of normal fluid pressure. In practice, fracturing often occurs in overpressured rocks (where the fluid pressure is higher than the normal fluid pressure gradient), and is often driven by the pressure of the fluids in the rocks, rather than by the stresses within the rock matrix itself (Secor 1965; Engelder 1985). Fluid pressure can have an impact on fracture nucleation and propagation in two ways:

- Prior to fracture nucleation, high fluid pressures within the rock pore space or within the faults may alter the local stress patterns around the fault tips and bends where fractures nucleate. Overpressured pore fluids will have the effect of reducing the effective stress within the rock matrix, thus promoting fracturing, while high-pressure fluids within the fault plane may exert an outward pressure on the fault walls, which will alter the orientation and magnitude of the local stress field around the fault.
- As the fracture propagates, it will dilate and fill with fluid. This dilation will itself cause a local perturbation in the stress field, which will be superimposed on the perturbation caused by the larger fault. The outward pressure on the fracture wall exerted by the fluid in the fracture will further perturb the local stress field. Although this stress perturbation will be much smaller than that caused by the fault, it will be at a maximum around the fracture tip. Thus, the very process of fracture propagation may itself alter the stress magnitude and orientation at the fracture tip.
To model the effect of fluid pressure on fracture propagation we used finite-element models.

**Finite-element modelling and the impact of fluid pressure**

Unlike elastic dislocation models, finite-element models allow us to apply stress directly to the fault and fracture walls to model fluid-driven fracture propagation. Alternatively, we can apply a tensile stress as a boundary condition to the model to simulate fluid overpressure. We therefore use finite-element models to determine whether fluid overpressure can account for the development of fracture corridors in chalk.

**Description of the models**

Only two fracture geometries were modelled: a releasing bend in a vertical strike-slip fault; and a restraining bend in the centre of a 60 m-long vertical strike-slip fault. The releasing bend model was identical to the hard-linked relay zone geometry shown in Figure 7d; the restraining bend model was similar except that the direction of offset on the bend was reversed (Fig. 11). In all cases, the fault was frictionless and the chalk was assumed to be homogeneous with mechanical properties for a burial depth of 2000 m (see Table 1). The finite-element models were all created and run using the Elfen finite-element code, developed by Rockfield Software Ltd and the University of Swansea.

All of the models were 2D plane stress models, representing a horizontal slice through the centre of the fault. Instead of applying a displacement to the fault directly, stress loads were applied to the four edges of the model, and the fault was allowed to slip freely in response to the resulting differential stress. The fault was oriented at 45° to the maximum stress orientation (as opposed to 75° for the elastic dislocation models). Three sets of models were run for each of the two geometries, simulating a normally pressured host rock, an overpressured permeable host rock and an overpressure impermeable host rock:

- In the normally pressured model, horizontal stresses were applied to the edge of the model equivalent to the effective stresses that would be expected in a chalk reservoir at a depth of 2000 m with a hydrostatic fluid pressure gradient and a $\sigma_{\text{min}}/\sigma_{\text{max}}$ ratio of 0.9. The same horizontal stress values were used to calculate fault displacement in the elastic dislocation models, and are given in Table 1. Note that both horizontal stresses are compressive.

- In an overpressured permeable rock, we assume that at all times the fluid in the host rock pores remains in equilibrium with the fluid in the fault and the fracture. We can therefore incorporate the effect of fluid pressure into the model by reducing the vertical effective stresses by an amount equal to the fluid overpressure, and recalculating the horizontal effective stress assuming an equivalent horizontal strain to the normally pressured case (we assume that horizontal strain is controlled by the external tectonic strain and is independent of fluid pressure; for simplicity we assume a Biot’s coefficient of 1). For high fluid overpressure (fluid pressure twice the hydrostatic gradient), this gives tensile horizontal effective stresses of $-1.93$ and $-2.59$ MPa.

- In an overpressured impermeable rock, we assume that there is no interaction between the fluid in the host rock pores and in the fault and the fracture. The fluid in the pores remains on a hydrostatic fluid pressure gradient, while a high-pressure fluid is injected into the fault and fracture, which produces an outward pressure on the fault and fracture walls. Absolute (rather than effective) horizontal stress values of 35.1 and 33.6 MPa are applied to the edge of the model; these are counteracted by an outward fluid pressure on the fracture walls of 39.2 MPa, equivalent to twice the hydrostatic gradient.

Note that in all of these models the chalk is represented as a continuum elastic material. Fluid overpressure is simulated by either a reduction in effective horizontal stress (the overpressured permeable rock) or an outward pressure acting on the walls of the fractures and fault (the overpressured impermeable rock). These represent end-member cases; in reality, fluid leak-off and pressure reduction during fracture propagation would result in a situation somewhere between these end-member cases. However, to model this accurately would require a coupled fluid flow and finite-element simulator to accurately model poroelastic behaviour. This was beyond the scope of the present study. Likewise, we have not modelled time-dependent chemical or mechanical processes such as creep, pressure solution and grain sliding, which may cause relaxation of the local stress anomalies over time. This is the subject of continuing research.

One of the aims of the finite-element modelling is to show how the presence of the propagating fracture may itself affect the local stress field. The current release of the Elfen code is unable to model fracture propagation explicitly, so to do this we ran a series of static models for each of the scenarios and geometries described above to simulate fracture propagation as a series of incremental steps. The initial model in each series contains a representation of the main fault but no fracture. This is
used to calculate the stress perturbation around the fault bend prior to fracturing. In the next model in the series, an initial ‘seed’ fracture is added at the location most prone to fracturing (with the lowest fracture resilience), parallel to the $\sigma_{h_{\text{max}}}$ orientation at this location. In each subsequent model, this fracture is incrementally extended in a similar manner, based on the stress field in the previous model (Fig. 11). This is continued until either the fracture stops propagating or it has propagated well away from the initial fault bend.

Results of the finite-element models

Figures 12 and 13 show the predicted sequence of fracture propagation for the three modelled fluid pressure conditions for a releasing and a restraining bend, respectively. The colours on these diagrams indicate the magnitude of the minimum horizontal stress, $\sigma_{h_{\text{min}}}$; hot colours indicate areas of low (or more tensile) stress where fractures are most likely to form. The black tick marks indicate the $\sigma_{h_{\text{max}}}$ Orientation (i.e. the likely fracture orientation). For clarity, some of the intermediate incremental models of fracture propagation have been omitted.

Note that there are three distinct components to the stress field in each of the models (Fig. 14):

- **The local stress perturbation**, around the fault bend, which is generated by strain induced by slip around the fault bend, and also by the fluid pressure acting on the fault and fracture walls. This is very intense (i.e. it produces large variations in stress magnitude and orientation) but only over a fairly small area around the fault bend.
- **The mid-range stress field**. This will be influenced by the fault as a whole, and will reflect the internal elastic strain generated by slip on the fault. Since the fault bend is small relative to the length of the fault, the mid-range stress field is likely to approximate to that of a planar fault. For a frictionless fault, this stress field will be oriented orthogonal to the fault itself.
- **The regional (or far-range) stress field**, which will reflect the external applied stress (in finite-element models) or the regional horizontal strain (in actual field examples). In these models the applied $\sigma_{h_{\text{max}}}$ orientation is horizontal; the regional stress field will thus be inclined at 45° to the mid-range stress field for a frictionless fault.

Fig. 11. Diagrams showing the geometry and boundary conditions of the finite-element models for the releasing bend (top) and the restraining bend (bottom), with close-ups showing the geometry of the fault bend in the initial model, and in the first iteration of the series of models simulating fracture propagation.
Fig. 12. Elfen models showing the initiation and propagation of a fracture around a releasing bend in a frictionless fault. (a) shows fracture propagation at normal fluid pressure; (b) shows fracture propagation with fluid pressure $= 2 \times$ hydrostatic pressure in a permeable host rock; and (c) shows fracture propagation with fluid pressure $= 2 \times$ hydrostatic pressure in an impermeable host rock. The top row of models show the initial stress state, prior to fracturing, in each case; the models beneath represent progressive growth of the fractures. Note that in (b) and (c) the fault and fractures are both dilatant, although the fracture aperture is small. Contours mark the magnitude of the effective minimum horizontal stress (in the normally pressured and permeable host rock) or the total minimum horizontal stress (in the impermeable host rock); positive values correspond to compressive stress, so that hot colours represent areas of most tensile stress. The black tick marks show the local $\sigma_{h\text{max}}$ orientation.
Fig. 13. Elfen models showing the initiation and propagation of a fracture around a restraining bend in a frictionless fault. (a) shows fracture propagation at normal fluid pressure; (b) shows fracture propagation with fluid pressure $= 2 \times$ hydrostatic pressure in a permeable host rock; and (c) shows fracture propagation with fluid pressure $= 2 \times$ hydrostatic pressure in an impermeable host rock. The top row of models show the initial stress state, prior to fracturing, in each case; the models beneath represent progressive growth of the fractures. Note that in (b) and (c) the fault and fractures are both dilatant, although the fracture aperture is small. Contours mark the magnitude of the effective minimum horizontal stress (in the normally pressured and permeable host rock) or the total minimum horizontal stress (in the impermeable host rock); positive values correspond to compressive stress, so that hot colours represent areas of most tensile stress. The black tick marks show the local $\sigma_{\text{max}}$ orientation. In the normally pressured example, the main fracture length refers to the longest continuous fracture segment.
propagation of a fracture around a releasing bend in a frictionless fault, at normal (hydrostatic) fluid pressure. The key features to note here are:

- The initial model shows a similar stress perturbation around the fault bend, prior to fracture propagation, as that seen in the elastic dislocation model of the releasing relay zone (Fig. 7d). In both cases, the point of lowest minimum horizontal stress (i.e. the point of fracture nucleation) lies adjacent to the fault, in the middle of the fault bend, and the $\sigma_{h\max}$ orientation here is parallel to the fault bend. The initial ‘seed’ fracture segment was therefore placed close to the middle of the fault bend in an orientation parallel to the fault bend segment. Note that only one fracture was modelled, on the upper side of the fault bend. However, since the initial models have rotational symmetry around the centre of the fault bend, a fracture nucleating on the lower side of the fault bend would propagate in an identical manner.

- As the fracture propagates, it does not alter the orientation of the initial stress field but, instead, follows the $\sigma_{h\max}$ stress lines shown in the initial (unfractured) models. It therefore loops back towards the fault, eventually intersecting this at the two corners surrounding the fault bend. We can thus conclude that fractures formed around a releasing bend in normal fluid pressure conditions are unlikely to propagate outwards, but will form short concentric fractures around the fault bend.

Fracture propagation around a restraining bend at normal fluid pressure. Figure 13a shows the propagation of a fracture around a releasing bend in a frictionless fault at normal (hydrostatic) fluid pressure. The key features to note here are:

- Although the stress around the restraining fault bend is higher than the background stress, an area of low stress forms on the outer shoulders of the fault bend, a short distance away from the corners, with $\sigma_{h\max}$ perpendicular to the fault. However, once we place an initial fracture segment at this point, the point of lowest stress migrates towards the fault bend rather than to the tip of the fracture. This implies that the initial fracture segment will not propagate but that a new fracture will form closer to the fault bend. Only once the seed fracture is placed at the fault bend can it start to propagate in a stable manner (i.e. with the point of lowest stress following the fracture tip). Note that, as with the releasing bend, the initial model has rotational symmetry around the centre of the fault bend, so fractures could propagate out from either of the fault corners.

- As with the releasing bend at normal pressure, the propagating fracture does not alter the orientation of the initial stress field but instead follows the $\sigma_{h\max}$ stress lines shown in the initial (unfractured) models. These loop back towards the fault, and thus the fractures will eventually intersect the fault rather than propagating outwards. Again, we can conclude that fractures formed around a restraining bend in normal fluid pressure conditions are unlikely to propagate outwards, but will form short concentric fractures.

Fig. 14. Diagram showing the local stress perturbation, the mid-range stress field and the regional stress field for the initial (unfractured) model with a releasing bend in a 60 m-long frictionless fault, with an overpressured impermeable host rock.
Fracture propagation around a releasing bend with fluid overpressure. Figure 12b, c shows the propagation of a fracture around a releasing bend in a frictionless fault, with fluid pressure $= 2 \times$ hydrostatic pressure, for both a permeable host rock (B) and an impermeable host rock (C). The key features to note here are:

- In the impermeable host rock, overall pressures are much higher than in the equivalent case at normal fluid pressure (Fig. 12a) but fractures are still able to propagate since the fluid pressure driving fracturing is correspondingly higher. In the permeable host rock, the overall pressures are much lower than in the equivalent case at normal fluid pressure and, indeed, are often tensile, as a result of the high fluid pressure.

- The geometry of the initial stress perturbation in both models is very different from that at normal fluid pressure. Instead of looping around the fault bend, the $\sigma_{h_{\text{max}}}$ orientation is perpendicular to the fault at all points, and the $\sigma_{h_{\text{max}}}$ stress lines lead away from the fault, gradually rotating towards the mid-range stress orientation. Similarly, the point of lowest minimum horizontal stress lies not in the centre of the fault bend but on the outer shoulders, centred on the two fault corners. As in previous models, only one propagating fracture is simulated here, although, in practice, fractures could form just as easily at either corner of the fault.

- Once a fracture forms at the point of lowest stress, it propagates outwards along the $\sigma_{h_{\text{max}}}$ stress lines. As it propagates, the point of lowest stress follows the fracture tip, and, indeed, the magnitude and size of the stress perturbation increases as the fracture grows longer. As the stress perturbation at the fracture tip grows independently of the stress perturbation around the fracture bend, it can continue to drive fracture propagation indefinitely.

- Although the orientation of the stress field behind the fracture tip is altered by the presence of the fracture, the stress field ahead of the fracture tip is not affected, and so the fracture propagates along the $\sigma_{h_{\text{max}}}$ stress lines as they appear in the initial (unfractured) model, gradually rotating towards the mid-range $\sigma_{h_{\text{max}}}$ orientation. This is an important result because it means that we can predict the geometry of these outward propagating fractures easily, using finite-element or elastic dislocation models of the initial fault, without needing to consider the impact of the fracture itself on the stress field.

- As the fracture grows, a ‘stress shadow’ (i.e. a zone of high minimum stress) forms around the fault bend. This is likely to suppress any further nucleation of fractures around the fault bend.

- Both the impermeable and the permeable host rock models show a similar initial stress perturbation around the fault bend, prior to fracture propagation, and a similar pattern of fracture propagation. Thus, the permeability of the host rock has no effect on the final geometry of the fractures as long as the fluid pressure is the same (although, in practice, overpressure may be limited in permeable rocks when compared with impermeable rocks).

Fracture propagation around a restraining bend with fluid overpressure. Figure 13b, c shows the propagation of a fracture around a restraining bend in a frictionless fault, with fluid pressure $= 2 \times$ hydrostatic pressure, for both a permeable host rock (B) and an impermeable host rock (C). The key features to note here are:

- As in the case of the releasing bend, overall pressures in the impermeable host rock are much higher than in the equivalent case at normal fluid pressure (Fig. 13a) but fractures are still able to propagate since the fluid pressure driving fracturing is correspondingly higher. In the permeable host rock, the overall pressures are much lower than in the equivalent case at normal fluid pressure and, indeed, are often tensile, as a result of the high fluid pressure.

- Again, the geometry of the initial stress perturbation in both models is very different from that in the equivalent cases at normal fluid pressure: the $\sigma_{h_{\text{max}}}$ orientation is perpendicular to the fault at all points, and the $\sigma_{h_{\text{max}}}$ stress lines lead away from the fault, gradually rotating towards the mid-range stress orientation. The point of lowest minimum horizontal stress lies on the outer shoulders of the fault bend but is centred on the two fault corners. As before, the rotation symmetry of the initial model implies that fractures could form at either of the two fault corners, although only one propagating fracture is simulated here.

- Once a fracture forms at this point, it again propagates outwards along the $\sigma_{h_{\text{max}}}$ stress lines. As it propagates, the point of lowest stress follows the fracture tip, and the magnitude and size of the stress perturbation increases as the fracture grows. Again, this stress perturbation is able to drive fracture propagation indefinitely. As before, the orientation of the stress field behind the fracture tip is altered by the presence of the fracture but the stress field ahead of the fracture tip is not affected, so the fracture propagates along the $\sigma_{h_{\text{max}}}$ stress lines as they appear in the initial (unfractured) model, gradually rotating towards the mid-range $\sigma_{h_{\text{max}}}$ orientation.
As before, a ‘stress shadow’ (i.e. a zone of high minimum stress) forms around the fault bend as the fracture grows, which will suppress any further nucleation of fractures here.

As in the previous examples, both the impermeable and the permeable host rock models show a similar initial stress perturbation around the fault bend, prior to fracture propagation, and a similar pattern of fracture propagation. Thus, the permeability of the host rock has no effect on the final geometry of the fractures as long as the fluid pressure is the same (although overpressure may be limited in permeable rocks compared with impermeable rocks).

Discussion

The results presented here clearly show that the geometry of the fractures formed around faults is controlled by the fluid pressure at the time of fracturing. We are able to replicate both the short, concentric localized fractures and the long, outward propagating fracture corridors observed around fault bends in the Pegwell Bay chalk if we assume that the two styles of fracturing formed at different fluid pressures (normal fluid pressure for the former, and high fluid overpressure for the latter). This is illustrated in Figure 15.

Another key point to note is the large stress perturbations formed around the tips of the outward propagating fractures in the overpressured cases (Figs 12b, c & 13b, c). The minimum horizontal stress within this zone is considerably lower than that in the surrounding rock, while $\sigma_{h,max}$ is near-parallel to the fracture. This will not only aid propagation of the initial fracture but may also cause new fractures to start propagating (particularly if the rock already contains a population of small microfractures). These new fractures will form close to the initial fracture tip (where the stress is lowest) and will be parallel to it. This process will thus generate a narrow corridor of subparallel fractures similar to those observed in the chalk outcrops in both Pegwell Bay and Flamborough Head. This mechanism for generating fracture corridors is similar to that proposed by Olson (2004), although he was modelling fracture propagation in a homogeneous rock in response to a uniform stress field, and so was unable to pinpoint the location at which the corridors would nucleate. Interestingly, he implied that fracture corridors would only form where the fracture propagation rate is near-critical or critical (i.e. rapid propagation driven by a regional stress field or a pulse of high-pressure fluid rather than slow propagation driven by chemical processes).

![Fig. 15. Sketch diagram illustrating the fracture geometry we would expect to form around a releasing and a restraining bend in a vertical strike-slip fault, under different fluid pressure conditions, based on the results presented in Figures 12 and 13.](image-url)
Other important conclusions that we can draw from the results here are:

- Outward propagating fractures can be generated from both releasing and restraining bends in the faults.
- As they propagate away from the fault bend, these fractures rotate towards the mid-range stress field. Thus, along most of their length, the orientation of the fracture corridors will reflect the orientation of mid-range stress field around the faults rather than the local stress perturbation around the fault bend. This implies that it is not necessary to know the detailed geometry of the fault bend to predict the orientation of the fracture corridors. It is, however, necessary to know the general orientation of the faults as a whole, and also the friction coefficient on the faults, in order to predict the mid-field stress orientation.
- Since the initial orientation of the fracture corridors reflects the local stress around the fault, this suggests that the faults were active at the time the fracture corridors formed. However, this does not mean that the fracture corridors formed at the same time as the faults since the fluid overpressure that triggered the development of the fracture corridors is also likely to have caused reactivation of the faults and, hence, regeneration of the local stress field around them.
- In an overpressured rock, once the fractures have propagated away from the fault bend, the stress perturbation at the fracture tip can continue to drive fracture propagation indefinitely. Two factors are likely to limit the length of the fracture corridors: they may intersect another fault; or the fluid driving fracture propagation may be dissipated (by expanding fracture volume or fluid leak-off into the host rock), causing the fluid pressure to drop. The former mechanism is consistent with the observations at Pegwell Bay, where the fracture corridors mostly connect adjacent faults. The latter mechanism is more likely to be applicable to the vein corridors at Flamborough Head – the propagation of these corridors appears to be driven by high-pressure fluid migrating along the normal fault at Selwicks Bay, and the vein corridors are clustered around this fault, suggesting that there is a maximum distance that they can propagate away from it before the fluid pressure drops below the pressure required to drive continued fracture propagation.
- The geometry of the fracture corridors in permeable and in impermeable overpressured host rocks is very similar. In reality, chalk is likely to lie in between the two end-member (permeable and impermeable) cases presented here: there will be some interaction between the fluid in the host rock pores and the fluid in the faults and fractures, but they are unlikely to be in complete equilibrium during the process of fracture propagation.

Conclusions

The key conclusions that we can draw from this study are:

- The fractures observed in the chalk at Pegwell Bay and Flamborough Head can be classified into one of three main geometric patterns: localized fracture geometries around fault tips, bends and splays (often forming concentric rings around fault bends); regional fracture sets comprising parallel, regularly spaced fractures (often with multiple intersecting sets in different orientations); and fracture corridors, comprising narrow bands of closely spaced parallel fractures.
- Of these three patterns, the latter two are likely to have the main impact of fluid flow through the chalk: the localized fractures are likely to provide only local enhancement of the permeability around the faults, while fracture corridors may provide highly permeable long-distance flow pathways, especially if they are uncemented and they connect adjacent faults, as they do at Pegwell Bay.
- The localized fracture patterns are likely to be the result of local stress anomalies developed around bends, tips and splays in larger faults. The size of these fracture patterns is likely to be proportional to the size of the faults they develop around. They can form in normally pressured rocks since the stress anomalies developed around bends, tips and splays in faults can be sufficient to generate local tensile stresses.
- Regional fracture sets probably form in response to either regional (tectonic) strain or to strains developed around very large folds and faults.
- Fracture corridors can only form under conditions of fluid overpressure; they may form in permeable rocks (where the fluid overpressure generates a tensile effective horizontal stress) or in impermeable rocks (where the high-pressure fluid exerts an outward pressure on the fracture walls that drives propagation). They tend to nucleate around tips and bends in larger faults but once they start to propagate they will rotate to reflect the mid-range or regional stress orientation, rather than the local stress anomalies developed around the fault tip or bend. It is therefore not necessary to know the exact geometry of the fault to predict the geometry of the fracture corridors. Once they start propagating,
they will continue to propagate until they either intersect adjacent faults (as appears to be the case at Pegwell Bay) or the fluid pressure that is driving their propagation is dissipated (as appears to be the case at Flamborough Head).

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