The role of structural geology in reservoir characterization

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Abstract: In this brief review of the role of structural geology in reservoir characterization a comparison is made between the original role of structural geology, which focused on the three-dimensional geometry and spatial organization of structures and which involved a statistical treatment of the spatial arrangement of structures such as faults and folds and on static analyses, and the more recent trend in structural geology which is concerned with the dynamics of structure formation and the associated interplay of stress and fluid migration.

Traditionally the role of structural geology in the location and definition of hydrocarbon reservoirs has been to define the geometry and spatial organization of structures such as folds and fractures. Field observations, theoretical analyses and analogue modelling of these various structures has led to a sound understanding of their likely three-dimensional geometry and their spatial relationships. For example field observations and analogue modelling of buckle folds has shown that these structures have a periclinal geometry, i.e. have the form of an elongate dome, basin or saddle, and that this geometry is characteristic of buckle folds on all scales. Figure 1 shows three examples which illustrate this point. The folds range in scale between folds with a wavelength of less than a centimetre to folds with wavelengths in excess of 10 km.

It is clear from Fig. 1a that in plan view the folds are arranged statistically in an en echelon manner. Analogue modelling (e.g. Dubey & Cobbold 1977; Blay et al. 1977) has shown how this pattern of distribution emerges as the folds are initiated and amplify into finite structures. Initially isolated folds form and as deformation continues these amplify and extend along their hinges. As they do so they may approach other folds. The way in which they interact is found to depend on the separation between the hinges of the two folds (Fig. 2). If the hinges of two interfering folds are off-set but the amount of off-set is only a small fraction of the wavelength, the folds link to form a larger fold with a deflection in the hinge line (Fig. 2a). Experimental observations also show that if the hinges of two folds are separated by more than about half their maximum wavelength but are still close enough to interact, they lock up, each preventing further propagation of the other, so that they become arranged in an en echelon fashion (Fig. 2b).

In the above discussion the linking and blocking of individual folds is considered. However, as an individual fold amplifies, it may cause the initiation of other folds on either side and so begin the formation of a wave-train. The work of Dubey & Cobbold (1977) has shown that as adjacent wave-trains propagate, they too may interact with each other by the process of 'linking' or 'blocking' depending on the relative wavelengths and positions of the two trains. Some of the different ways in which they can interact are shown in Fig. 3. It can be seen that linking of folds from two fold-trains can also give rise to structures which, in plan view, have deflections in their hinges (Fig. 3c(ii)). Alternatively folds may bifurcate (Figs 3c(iii) & d(ii)).

As well as having a limited extent along their hinges, folds often die out relatively rapidly in profile section. A typical profile section through a fold in a multilayer is shown in Fig. 4a. The random initiation of folds within a multilayer would give rise to the type of fold distribution shown in Fig. 4e. It can be seen therefore that the folds are arranged in an en echelon manner both in plan view and in plan. For an extended discussion of this the reader is referred to Price & Cosgrove (1990).

Exactly the same spatial arrangement can be found for faults. For example normal faults are often arranged in an en echelon manner in both profile and plan view and considerable work is at present being carried out in an attempt to understand the stress distribution and fracture patterns likely to develop in the relay zones between these overlapping fractures.

It can be seen from the above discussion that the shortening along any line normal to the fold hinges or the extension along any line normal to the strike of the faults will be constant regardless of where the line is drawn, but the exact location of the individual folds or faults along any particular line cannot be determined.

Although an understanding of the three-dimensional geometry and spatial organization
Fig. 1. Three examples of folds displaying periclinal geometry and an en echelon spatial organization. (a) A crenulated mica schist from the Lukmanier Pass, Switzerland. Coin 1 cm. (b) En echelon periclines in a folded metasediment near Luarca, North Spain. Coin 2 cm. (c) Large pericinal folds in the Zargros mountains. The pericline in the centre of the image is 20 km long.
the location of potential reservoirs, it does not provide any insight into the possible fluid migration paths that may have operated during the initiation and amplification of these structures. More recent studies (e.g. Sibson 1990; Sibson et al. 1975, 1988; Cosgrove 1993) have been concerned with dynamics of deformation and the interplay of deformation, stress and fluid flow.

In the following section, brittle failure and hydraulic fracturing are briefly considered together with the stress fields that generate them in order to illustrate this approach and show its possible relevance to the migration and concentration of hydrocarbons.

The discussion is then extended to include the links between brittle structures and fluid flow and ductile structures and fluid flow.

Brittle failure and hydraulic fracturing

In this section the geological expression of hydraulic fracturing is considered with particular reference to fracturing of a sedimentary succession during burial and diagenesis. The theory of brittle failure and hydraulic fracturing is discussed in most structural texts (e.g. Price 1966; Phillips 1972; Price & Cosgrove 1990), to which the reader is referred, and only a brief summary of these concepts relevant to the ideas discussed in this paper are presented here.

Figure 5a is a summary diagram showing the failure envelope for brittle failure. This is determined in part by the Navier–Coulomb criteria for shear failure and in part by the Griffith criteria of extensional failure. Figure 5b and c show extensional fractures and shear fractures respectively, and the principal stresses with which they are associated. These stress states are represented as Mohr circles on Fig. 5a where it is clear that in order for shear failure to occur the Mohr circle must be sufficiently large to touch the shear failure envelope. It follows from the geometry of the failure envelope that this can only occur if the diameter of the Mohr circle (i.e. the differential stress $\sigma_1 - \sigma_3$) is greater than four times the tensile strength of the rock ($T$). It also follows from the geometry of the failure envelope that the angle $2\theta$ between the normal stress axis and the line joining the centre of the Mohr circle to the point A where it touches the failure envelope, is the angle between the two conjugate shear fractures (Fig. 5c). For extensional failure to occur the Mohr circle must touch the failure envelope at point B. This can only occur if the differential stress is less than four times the tensile strength of the rock.

Thus the type of brittle failure indicates whether the differential stress during fracturing was greater or less than $4T$. Occasionally it is clear that both extensional and shear failure occurred together during a single deformation event. An example where this has occurred is illustrated in Fig. 6, which shows a line drawing of incipient boudinage in a relatively thick sandstone layer in the Carboniferous turbidites at Millook, north Cornwall. The individual boudins are separated from each other by quartz veins; some of these are single extensional veins which cut completely across the layer and are an expression of extensional failure; others
form part of an en echelon array and are a manifestation of shear failure. It can be concluded therefore that the differential stress during boudin formation was around four times the tensile strength of the sandstone at that time.

It is interesting to note that the boudin necks defined by shear failure and extensional failure zones are both made up of tensile fractures and that the only difference between the two is the spatial organization of these fractures. In general, in a rock undergoing extensional failure, the individual tensile fractures, although aligned, are randomly distributed whereas during shear failure they are organized in such a way as to define either one or a conjugate pair of en echelon extension fractures (see Kidan & Cosgrove 1996).

The orientation of extensional fractures

The stress states represented by the Mohr circles shown in Fig. 7a all have a differential stress less than $4T$ and because they all touch the failure envelope they will all generate extensional failure. The differential stresses vary between just less than $4T$ (circle i, Fig. 7a) and zero (circle iv, Fig. 7a). When $\sigma_1 - \sigma_3$ is zero the stress state is hydrostatic and the Mohr circle collapses to a point.

The relationship between extensional fractures and the principal stresses generating them is shown in Fig. 7b (e.g. Anderson 1951; Price 1966). These fractures form parallel to the maximum principal compression $\sigma_1$ and open against the least principal compression $\sigma_3$. In the stress state represented by Mohr circle i (Fig. 7a), which has a relatively large differential stress, there is a definite direction of easy opening for the tensile fractures, i.e. parallel to $\sigma_3$. Thus the fractures that form will have a marked alignment normal to this direction (Fig. 7b(i)). However, the differential stress for the stress states represented by the Mohr circles ii–iv becomes progressively smaller until, for the hydrostatic stress represented by the Mohr circle iv, the differential stress is zero. In a hydrostatic stress field the normal stress acting across all planes is the same and therefore it is equally easy to open fractures in all directions. Thus the fractures will show no preferred orientation and if they form sufficiently close to each other they will generate a breccia texture (Fig. 7b(iv)). It can be argued therefore that as the differential stress becomes progressively lower the tendency for the extension fractures to be aligned will become less and less (Cosgrove 1995).

Hydraulic fracturing

The state of stress in the Earth’s crust tends to be compressional and true tensile stresses are thought to be uncommon. This will be particularly true for the stress states in a sedimentary pile undergoing burial and diagenesis in a tectonically relaxed basin. Nevertheless extensional fractures occur commonly and this
apparent contradiction has been satisfactorily explained by arguing that failure occurs by hydraulic fracturing (e.g. Phillips 1972). It is argued that the internal fluid pressure in the sediment or rock acts so as to oppose the applied stresses and that the rocks respond to the effective stresses $\sigma_1 - p$, $\sigma_2 - p$, $\sigma_3 - p$, where $p$ is the fluid pressure. Thus a state of lithostatic stress in a rock will be modified by the fluid pressure to an effective stress state $\sigma_1 - p$ and $\sigma_3 - p$ and the Mohr circle will be moved to the left by an amount equal to the fluid pressure. The three stress states represented by the solid Mohr circles shown in Fig. 8 are all stable stress fields in that they do not touch the failure envelope and therefore will not cause failure. However, all three circles can be driven to the left by a fluid pressure until they intersect the failure envelope, when hydraulic fracturing will occur. It can be seen that stress state i will cause shear failure, stress state ii will cause aligned extensional failure and stress state iii will cause brecciation of the rock by the formation of an almost random array of extension fractures.

It is a common misconception that the result of hydraulic fracturing in sediments and rocks is the formation of randomly oriented extension fractures and the generation of breccia textures (Fig. 7b(iv)). It is clear from the above discussion that the expression of hydraulic fracturing can vary, ranging from randomly oriented extensional fractures through aligned extensional fractures to shear fractures.

Having briefly considered brittle failure, the relationship between stress and fractures and the process of hydraulic fracturing, we can proceed to consider the factors that affect the state

Fig. 4. Typical three-dimensional geometry and spatial organization of multilayer folds.
of stress in a sedimentary succession and the type and orientation of the hydraulic fractures that might develop in them.

Stress variation with depth

In this section the state of stress in sediments being buried in a basin is briefly considered in order to predict the type and orientation of hydraulic fractures that might form. The stress state will depend on the material properties of the sediment or rock and upon the boundary conditions. Consider the relatively simple boundary conditions which affect sediments in a tectonically relaxed basin, i.e. one in which the main source of stress is due to the overburden. If the boundary conditions are such that horizontal strains are prevented by the constraints of the rock mass surrounding the area of interest, then it can be shown (e.g. Price 1966) that the vertical and horizontal stresses are related as follows:

\[ \sigma_H = \sigma_V / (m - 1) \]  

where \( m \) is Poisson's number, the reciprocal of Poisson's ratio. The vertical stress will be \( \sigma_V \) and its magnitude given by:

\[ \sigma_V = \sigma_1 = z \rho g \]  

where \( z \) is the depth, \( \rho \) the average density of the overlying rocks and \( g \) the acceleration due to gravity.

Fig. 5. (a) The Navier–Coulomb/Griffith brittle failure envelope. The two Mohr circles represent stress states that would give rise to extensional failure (the smaller circle) and shear failure. (b) and (c) show the relationship between the principal stresses and extensional failure and shear failure planes respectively.

Fig. 6. Line drawing of incipient boudins formed in a sandstone layer in the Carboniferous turbidites at Millook, north Cornwall. Some boudins are separated from each other by a single extensional gash (i) and others by an array of extensional fractures (ii).
gravity. These equations show that if the overburden has a constant density and Poisson’s number does not change with depth, then the vertical and horizontal stresses increase linearly with depth (Fig. 9).

It can be seen from Fig. 9 that the differential stress ($\sigma_v - \sigma_h$) will also increase linearly with depth. From the discussion of brittle failure given earlier it follows that the type of hydraulic fractures that will form in the upper section of the sedimentary pile, where the differential stress is less than four times the tensile strength of the rock, will be vertical extensional fractures opening against the least principal stress $\sigma_3$ and the fractures that form at depths where the differential stress is greater than $4T$ will be shear fractures dipping at around $60^\circ$ and striking normal to $\sigma_3$.

Thus when the vertical stress is the maximum principal compression and when the two horizontal principal stresses are not equal, the fracture patterns generated by hydraulic fracturing in the upper and lower zones indicated on Fig. 9 will be as shown in Fig. 10a and b. In the upper zone where extensional failure occurs, the fractures will be vertical and will be aligned normal to $\sigma_3$. In the lower zone where shear failure occurs, conjugate fractures will form dipping at $60^\circ$ and intersecting (and striking) parallel to $\sigma_2$. It can be seen from Fig. 10a and b that in

Fig. 7. (a) Four Mohr circles that represent four stress states that will give rise to extensional failure (i.e. $\sigma_1 - \sigma_3 < 4T$). (b) (i)-(iv) show the organization of the extensional fractures for each stress state.

Fig. 8. The influence of an increase of fluid pressure on three stable stress fields i, ii and iii (see text).
Fig. 9. Plot of variation of vertical and horizontal stress and fluid pressure with depths according to Eqns 1 and 2 which assume that the stresses are generated by the overburden in a tectonically relaxed basin. The expression of the hydraulic fractures is determined by the differential stress which increases with depth. At the depth when it exceeds $4T$ the fractures change from extensional to shear.

Field evidence for hydraulic fractures

The arguments outlined above indicate that as sediments undergo burial and diagenesis in a basin, conditions of stress and fluid pressure are likely to be encountered that will lead to the formation of hydraulic fractures. Indeed, these fractures would provide a transient increase in permeability that would facilitate the dewatering of relatively impermeable horizons. Unfortunately these fractures are generally not preserved, forming as they do at relatively shallow depths.
where they are unlikely to be preserved as vein systems and where the rock properties are such that barren fractures would tend to heal once the excess fluid pressure has been dissipated.

The author has examined a number of quarries and other outcrops of low permeability shales thought to have been overpressured during their burial. Remarkably little evidence was found of such fracturing. Occasionally, thin (1–2 mm thick) bedding-parallel veins of fibrous calcite occur, with the fibres oriented normal to the bedding fractures, as for example in the Kimmeridge shales at Kimmeridge Bay in Dorset. ‘Chicken wire’ texture has been recorded in cores from shales known to have been previously overpressured (Powey 1990, pers. comm.). These are polygonal arrays of fractures along which very thin veins of calcite have been precipitated. Hydraulic fractures are sometimes preserved as sedimentary dykes in the regions around sandstone bodies in shales if the fluid pressures are sufficiently high to fluidize the sands. Sedimentary dykes provide one of the most convincing demonstrations of hydraulic fracturing in sedimentary successions.

Recently, Cartwright (1994) has reported the occurrence of large polygonal fracture arrays in Early Cainozoic mudrock-dominated sequences from the North Sea. The fractures have been mapped using regional two- and three-dimensional seismic data. They occur in the stratigraphically bounded tiers in deep-water sequences bounded by regionally condensed sequences (seals). The faults are organized into cellular networks comprising polygonal prismatic and pyramidal forms and Cartwright suggests that these faults are the result of hydraulic fracturing. The polygonal cells are between 500 and 1000 m in diameter and a synoptic block diagram illustrating the structural framework of the faults is shown in Fig. 11.

Despite the impressive seismic images of hydraulic fracture patterns presented by Cartwright (1994), evidence for the occurrence of hydraulic fractures in low-permeability rocks known to have been overpressured at some stage in their history, is remarkably scanty.
Even when fractures are found in such rocks it is difficult to demonstrate convincingly that they are the result of hydraulic fracturing. For example, the barren fractures frequently exhibited by shales at outcrop often form polygonal arrays when viewed on the bedding plane and are normal to bedding. Exposures of such rocks are often characterized by polygonal prisms. This is a fracture pattern that might be expected to form during the burial and overpressuring of the rock (Fig. 10c). Nevertheless it is difficult to determine whether these fractures represent ancient hydraulic fractures generated during burial and dewatering which have been reopened as a result of the release of residual stresses during exhumation, or whether they represent desiccation fractures resulting from the drying out of the shales on exposure.

This similarity in the geometric organization of desiccation fractures and extensional hydraulic fractures reflects an underlying similarity in the mechanism of formation of both structures. This can be exploited when hydraulic fracturing is studied using analogue models.

The experimental investigation of the initiation and development of hydraulic fractures in overpressured shales and semi-lithified rocks is extremely difficult. However, if it can be argued that the process is directly analogous to the formation of cooling fractures or desiccation fracture, then the formation of these fractures under a variety of boundary conditions can be studied with relative ease.

The association of thrusts and high fluid pressures

In the above discussion the build-up of fluid pressure and the formation of hydraulic fracturing in a tectonically relaxed basin was considered. Such a fluid pressure build-up will also occur in tectonic regimes, particularly those associated with compressional tectonics. In this section the association of thrusts and high fluid pressures are considered.

The recognition of overthrusts, large blocks of rock up to 10 km thick and with lengths and widths often in excess of 100 km, which had been transported tens of kilometres along sub-horizontal fault planes, presented geologists with the challenge of understanding the mechanism by which these blocks were moved. An early study of this problem was carried out by Smoluchowski who represented the overthrust block as a single rectangular prism caused to move over a flat, dry surface by the application of a horizontal push from one end. By selecting the appropriate values for the coefficient of sliding friction and rock strength it can be shown that overthrusts with lengths greater than a few tens of kilometres cannot be moved.
by this mechanism unless the applied stress exceeds the strength of the rock.

Alternative mechanisms for the movement of large overthrusts were presented. For example, Oldham (1921) suggested that movement on a thrust plane may not be synchronous and that decoupling and displacement may occur locally and migrate along the thrust plane in a caterpillar-like manner. This mechanism would allow the overthrust block to move forward without having to overcome the total frictional resistance of the block to movement at any one time. In this way, it was argued, large overthrust blocks could be moved at stresses below the strength of the rock. Despite such innovative thinking it was not until 1959 that a widely accepted solution appeared to the mechanical problems associated with the movement of large-scale thrusts.

In their classic paper, Hubbert & Rubey (1959) argued that the existence of high fluid pressures would reduce the effective normal stress across a potential thrust plane and thus reduce the horizontal compressive stress necessary to move the thrust to below the brittle strength of the rock. This idea represented a major step forward in the understanding of the generation and movement of large overthrusts and, following this work, the association of overthrusting and high fluid pressures provided a great insight into the thrust to below the strength of the block to movement at any one time. In this way, it was argued, large overthrust blocks could be moved at stresses below the strength of the rock. This idea represented a major step forward in the understanding of the generation and movement of large overthrusts and, following this work, the association of overthrusting and high fluid pressures became almost axiomatic. More recently the overthrust 'paradox' has been reassessed (Price 1989) and this association questioned.

Although Hubbert & Rubey's work on high fluid pressures provided a great insight into the role that fluid pressures may play in thrust initiation, it says nothing about the migration of fluids during thrusting.

**Faults, folds and fluid migration**

The migration of fluids along major faults during and immediately after reshear has been discussed by Sibson et al. (1975), who introduced the idea of seismic pumping. This concept sprang from observations of hydrothermal vein deposits found in the upper, brittle regions of ancient fault zones. The textures of these deposits usually indicate that mineralization took place episodically and it was suggested that episodic injection of hydrothermal fluids could be accounted for by the dilatancy–fluid diffusion model for energy release in shallow earthquakes proposed by Scholtz et al. (1973). In later papers, Sibson et al. (1988) and Sibson (1990) proposed two other mechanisms, the suction-pump and fault-valve mechanisms which also predict periodic variations in fluid pressure along faults associated with fault reactivation.

In contrast, the migration of fluids in association with the development of folds has received comparatively little attention. It is, however, possible to determine the stress gradients within and around a fold from the equations that govern the buckling behaviour of anisotropic bodies such as a sedimentary sequence. They show that there is a difference between the mean stress inside and outside a fold and that this difference (i.e. the stress gradient) changes as the fold amplifies. For a box fold the gradient is initially such that fluids are drawn into the fold from the surrounding region. However, beyond a certain amplification the gradient is reversed and fluids are driven out of the fold. This process can be inferred from the geometric changes that accompany the amplification of a kink-band in a layered material where the layers remain a constant thickness during the deformation (Fig. 12a–c). Initially there is an increase in volume within the kink-band. This continues until the layering inside the kink-band is normal to the kink-band boundary, i.e. \( \omega = \theta \) (Fig. 12b), when it reaches a maximum. Up to this point fluids will be drawn into the fold from the surrounding unfolded region. As the kink-band amplifies beyond this point, the volume of the kink-band is reduced and fluids will be expelled from the fold. When \( \omega = 2 \theta \) (Fig. 12c), the volume of the kink-band is the same as it was before the fold was initiated. Any further amplification would require the layering inside the kink-band to thin and if there is no mechanism by which this thinning can be achieved, the fold locks up.

The magnitude of the stress gradient between the fold and its surroundings has been quantified by Summers, who shows that the difference in mean stress inside \( (\sigma_1) \) and outside \( (\sigma_2) \) a fold is given by:

\[
\sigma_1 - \sigma_2 = \tau_e \left[ (N/Q + 1) - (N/Q - 1) \cos (\theta - \omega) \right]
\]

where \( \theta \) and \( \omega \) are defined as in Fig. 12d, \( N \) and \( Q \) are measures of the resistance to compression and shear respectively, in the direction of the applied maximum principal compressive stress, and \( \tau_e \) is the maximum shear stress in the layering adjacent to (i.e. outside) the fold. The variation in mean stress gradient as the fold amplifies can be clearly seen by expressing Eqn 3 graphically (Fig. 12e). As the fold begins to amplify, there is an increase in the stress difference which would tend to draw fluids into the
Fig. 12. (a)–(d) Volume changes that accompany the amplification of a kink-band. During the early stages the volume increases and fluids are drawn into the fold (a–b). However, as the structure amplifies beyond the stage where \( \omega = \theta \) (b), the volume decreases and fluids are expelled. (d) Detail of (b) showing the dilation within the kink-band. (e) The graphical expression of Eqn 3 showing the relationship between the difference in mean stress inside and outside the kink-band and the orientation \( \omega \) of the foliation within the kink-band. Graphs for materials with anisotropies \( (N/Q) \) of 5 and 100 are shown ((e) after Summers).

Fig. 13. (a)–(d) Various accommodation structures that develop during multilayer buckling as layers with anomalous thicknesses adjust to fit into the overall wavelength and amplitude of the multilayer buckles. These may considerably increase the permeability of the hinge region of the fold ((a) and (b) after Ramsay 1974; (c) after Price & Cosgrove 1990; (d) after Herman 1923).
structures such as folds and faults, which are potential hydrocarbon traps, provides a fertile area of research which will doubtless have an important impact on the role that structural geology plays in hydrocarbon exploration.

Conclusions

Knowledge of the three-dimensional geometry and spatial organization of structural traps has for many years been one of the most powerful tools in hydrocarbon exploration. Nevertheless, these models of structure geometry and organization are essentially static and do not consider the interaction of fluids and structures during deformation when the structure is being initiated and amplified. However, recent advances in the understanding of the role of fluid pressure in the initiation of faults and folds and the subsequent influence of these structures on the migration and concentration of fluids, has opened a new chapter in the role of structural geology in reservoir characterization.

References


