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Sedimentary basin inversion and intra-plate shortening

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Abstract

Sedimentary basin inversion, the shortening of formerly extensional basins, is accommodated mainly by compressional reactivation of extant faults and fractures across a wide range of scales. As such, inversion is a large-scale manifestation of Byerlee friction, the dynamic criterion for fault reactivation governing the effective shear strength of the shallow crust. Basin inversion generates distinctive deformational architecture, and it is implicated strongly in sedimentary basin exhumation. As a principal source of horizontal stress, inversion drives significant sedimentary porosity reduction and resultant fluid flow. Upper crustal deformation is critically dependent on fluid overpressure (i.e., pore fluid pressures greater than would be calculated from a hydrostatic gradient), and, perhaps more than in any other tectonic setting, overpressures are potentially large during sedimentary basin inversion. This review therefore includes discussion of the role of fluid overpressure in inversion and the evidence for it. Collectively, inversion has profound implications—good and bad—for the prospectivity of many petroliferous sedimentary basins. Thus, recognizing the evidence for basin inversion, quantifying its magnitude and understanding the mechanisms that accommodate inversion and other phenomena affected by it, have become essential components of the basin analyst's remit.

Keywords: Basin inversion; Tectonics; Fault reactivation; Reverse fault; Rift basin; Fluid overpressure; Effective stress; Exhumation; Petroleum geology

1. Aims and scope

In this paper, sedimentary basin inversion describes the compressional or transpressional reactivation and shortening of formerly extensional basins. Inversion induces significant changes in the tectonic structure and dimensions of sedimentary basins, it has a major influence on thermal history and rock properties, and its impact on petroleum prospectivity is fundamental. Furthermore, as one of the principal mechanisms driving uplift and erosion, sedimentary basin inversion is often primarily implicated in the exhumation of basins—the exposure at the surface of formerly deeply buried rocks.

To date, much of our understanding of sedimentary basin inversion originates from seminal studies of Mesozoic extensional basins in the NW European Alpine foreland. They were inverted in response to compression generated by European–African plate collision in the Cenozoic (Ziegler, 1987). A principal achievement of this work has been to highlight characteristic structural geometry and erosional unconform-

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ities resulting from intracratonic basin inversion. Furthermore, without fully elucidating the mechanism(s) responsible for the transmission of stress through the lithosphere, these studies serve also to emphasize the potentially far-reaching influence of basin inversion (e.g., Cenozoic inversion affected basins some 1600 km from the Alpine front: Ziegler et al., 1995). However, whilst the European Alpine foreland is an excellent natural laboratory in which to investigate sedimentary basin inversion, it is by no means the only geodynamic setting in which inversion tectonics occurs. Moreover, the influence of the Iceland plume during North Atlantic opening means that it is often difficult to isolate the effects of inversion from epeirogenesis in diagnosing the nature of the uplift responsible for the exhumation of NW European basins.

Sedimentary basin inversion is, literally, a global process. For example, compressional stress originating from major changes in plate configuration during the Santonian–Maastrichtian interval are recorded in inversion episodes across Africa, Arabia, Australasia, Europe and the Americas (Guiraud and Bosworth, 1997; Davidson, 1997: fig. 31; P.F. Green, personal communication). Marshak et al. (2000) show that the inversion of North American Proterozoic extensional basins during the Ancestral Rockies and Laramide contractional events was largely controlled by the trends of the basement faults. Given that these faults formed during extensional episodes some 1.3–1.1 and 0.9–0.7 Ga that affected the entire Rodinia, we should expect to see evidence for their inversion across cratonic platforms worldwide.

This review integrates well-established ideas on the geometry and kinematics of inversion structures (Cooper and Williams, 1989; Buchanan and Buchanan, 1995; Ziegler et al., 1995) with modern concepts of deformational mechanics, including the role of fluid overpressure, exhumation and geodynamics. As such, it serves to emphasize the central role of inversion in the long-term evolution of many, probably most, sedimentary basins.

2. Definition

Although the term "inversion" was first coined by Glennie and Boegner (1981), inverted basins have been recognized for a long time (e.g., Lamplugh, 1920; Stille, 1924; Pruvost, 1930; Voigt, 1963). The Geological Society Special Publication on Inversion Tectonics (Cooper and Williams, 1989) included extensive discussion on the uses and abuses of basin inversion (Cooper et al., 1989). This suggested it should be confined to (1) basins whose extensional phase was actively controlled by faults and (2) scenarios in which changes in the regional stress field resulted in the extensive re-use, or reactivation, of pre-existing faults. Such a definition includes inversion in orogenic belts, but it implicitly excludes flexural, thermal and isostatic mechanisms of sedimentary basin uplift, such as the synkinematic uplift of normal fault footwalls (footwall uplift). The use of "negative inversion" to describe changes in the sense of fault displacement from reverse to normal is generally discouraged, though it remains a useful and relevant concept for understanding the processes of syn- and postorogenic extension.

3. Tectonic settings

Sedimentary basin inversion has been widely documented from most of the main types of basins, namely continental rifts (e.g., Daly et al., 1989), aulacogens (e.g., Hamilton, 1988; Bartholomew et al., 1993; Varga, 1993; Molzer and Eerslev, 1995; Knott et al., 1995), rifted continental margins (e.g., Boldreel and Andersen, 1993; Guiraud and Bosworth, 1997; Gasperini et al., 2001; Benkhelil et al., 2002; Turner et al., 2003), backarcs (e.g., Letouzey et al., 1990), orogenic foredeeps (e.g., Camara and Klimowitz, 1985; Gillcrist et al., 1987; Turner and Hancock, 1990), intracratonic basins (e.g., Tucker and Arter, 1987; Butler, 1998; Underhill and Paterson, 1998) and 'peri-orogenic' regions of continental escape tectonics (e.g., Morley et al., 2001). Furthermore, rapid switches between transtensional and transpressional strain experienced by pull-apart basins as they are translated between releasing and restraining bends along strike-slip fault systems means that they often exhibit classic inversion features (e.g., intra-continental strike-slip fault zones: Storti et al., 2001a,b; Crowell, 1976; Harding, 1985; transform margins: Basile et al., 1993; Mascle et al., 1996).

Whilst the 'inversion' of pull-apart basins will not feature significantly in this review, it is worth noting that most inversion episodes are non-coaxial, and hence, oblique-slip or strike-slip kinematics are important modes of fault reactivation during inversion. In this context, non-coaxiality describes scenarios in which the bearing of the axis of maximum horizontal stress during inversion does not coincide with that of the earlier axis of minimum horizontal stress. In such a setting, shortening strains will often be partitioned between faults displaying dip-slip, oblique-slip and strike-slip kinematics (e.g., Williams, 2002). A classification of inverted settings in terms of the relative magnitudes of compression and strike-slip components (Lowell, 1995) highlights several basins in South America whose inversion was highly non-coaxial.

4. Structural criteria for recognition

To varying degrees, inverted basins are characterized by reversals in the sense of dip-slip fault displacement, from normal to reverse (Fig. 1), a change in the polarity of structural relief (i.e., low-lying basin areas become structural culminations) and expulsion of the synrift fill of a basin (Fig. 2). Reversals in the sense of dip-slip fault displacement result in the formation of reverse faults. Terms "reverse" and "thrust" fault should be applied rigorously. Thrusts are dip-slip faults that originate in a horizontally compressive stress field, typically at angles of 30° to maximum principal stress. Whilst reverse faults also are a product of horizontal compression, they attain steeper angles (c. 60°) to the horizontal. Reverse faults owe their steep inclination either to: (1) back-rotation and steepening of a thrust fault after its formation; (2) transpressional settings in which shallow thrust faults steepen with depth as they converge on a strike-slip basement lineament from which they splay; or (3) initiation as normal faults, subsequently reactivated as reverse structures during sedimentary basin inversion. Thus, in non-orogenic settings and away from the comparatively unusual transpressional tectonics described above, the presence of reverse faults is a strong indication that inversion has occurred.

The diagnostic criterion for recognizing sedimentary basin inversion is identification of the null point (Williams et al., 1989), or, in three dimensions, the null line. Fig. 1 shows a reactivated fault in an inverted basin along which net displacement changes

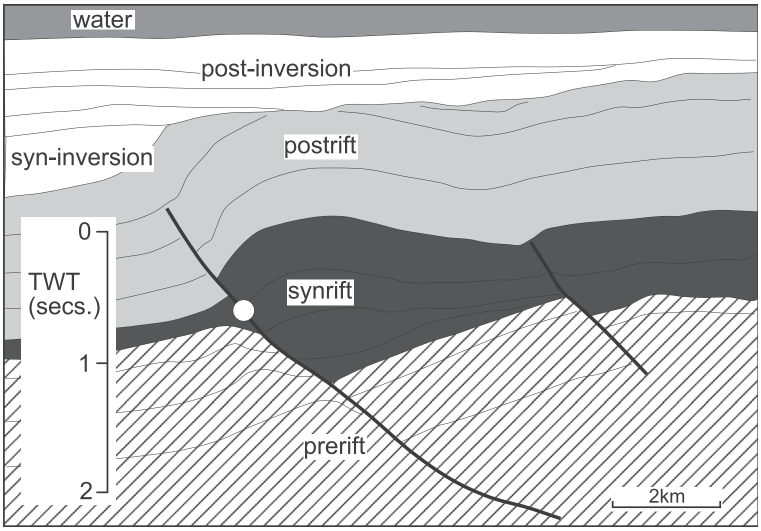

Fig. 1. Cross-section showing typical inversion geometry in an inverted half-graben from the East Java Sea basin, Indonesia. After Goudswaard and Jenyon (1988). White polka dot signifies the position of the null point, marking the divide between reverse displacement above and normal displacement below. Thus, progressively more severe inversion leads to migration of the null point (or, in three dimensions, the null line) from the tips of a fault toward its centre, where pre-inversion normal displacement was greatest.

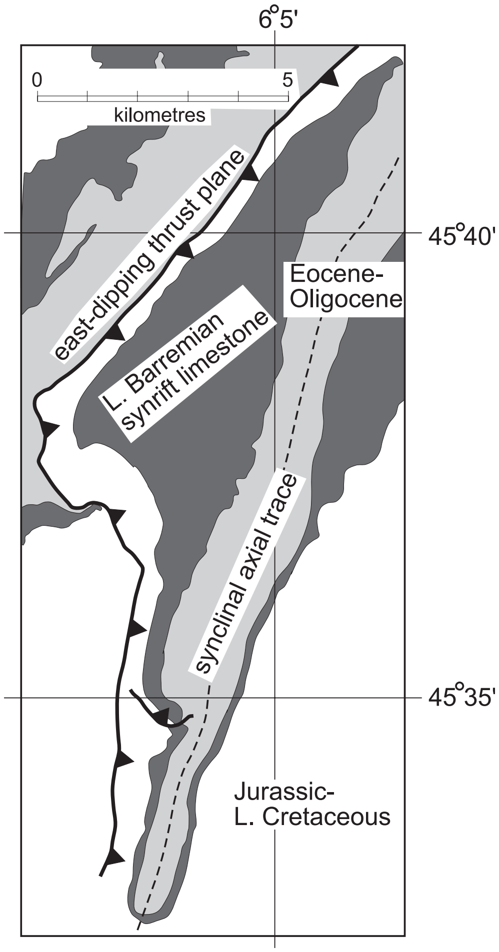

Fig. 2. Geological map of an inverted half-graben near Chambery, external French Alps, SE France. The cuspate outcrop pattern of the Lower Barremian synrift limestone, exhibiting marked along-strike variation in thickness, records its synkinematic deposition in the hanging wall of an east-dipping listric normal fault during regional Cretaceous extension. Subsequent reactivation of the fault during Alpine shortening led to expulsion of the synrift succession in the hanging wall of a major thrust (shown with barbs in hanging wall) such that it now forms a major topographic culmination. Map from Carte Geologique Detaille de la France (1960).

from normal in the lower reaches of the fault, to reverse in its upper reaches. Such fault displacement patterns are best explained by invoking either:

1. Reverse reactivation of a normal fault. Fault systems evolve during many increments of coseismic displacement. Consequently, fault displacement will increase progressively from its tip, where displacement is by definition zero, to its centre where the fault (system) will have undergone the greatest number of increments of displacement. Subsequent reversal of the sense of fault displacement, from normal to reverse, means that the fault tip regions will display net reverse offset before the centre. Consequently, segments of net reverse and net normal displacement will be separated by the so called null point at which fault displacement is zero.
2. Scissor-like fault kinematics in which each side of the fault essentially pivots about a 'false' null point positioned more or less halfway between the fault tips. This pivoting motion may take place about horizontal, vertical or obliquely inclined axes. Such faults have been mapped from 3D seismic surveys and may be interpreted as large-scale Riedel shears that develop within torsional strain fields.

As the magnitude of inversion increases, that is, the ratio of shortening strain to initial extensional strain, the null point(s) will migrate progressively along a fault, toward its centre. Consequently, sedimentary basin inversion will often generate shortening in the cover sequence before the net extension at basement level has been cancelled. As an elegant explanation for apparently bizarre structural geometries, such as faults that change from normal to reverse along strike (Fig. 3), seamless vertical transitions from hanging wall rollover monocline to compressional anticline (Fig. 1), upthrown hanging wall depositional 'thicks' (Figs. 2 and 4) and compressional folds in the hanging walls of normal faults (Fig. 5), the null point concept is an essential tool for seismic interpreters and field geologists working in inverted basins.

An inevitable consequence of null points is that reactivated faults in inverted basins are likely to confound the wealth of empirical data on displacement–length relations (cf. Walsh et al., 2002). This mainly reflects the fact that, except along the neoformed segments of reactivated faults (i.e., the near-tip segments that propagated during a fault's reactivation), the measured total displacement between footwall and hanging wall markers is apparent. Following an inversion episode, therefore, the thickest area of synrift sediment accumulation will not necessarily be

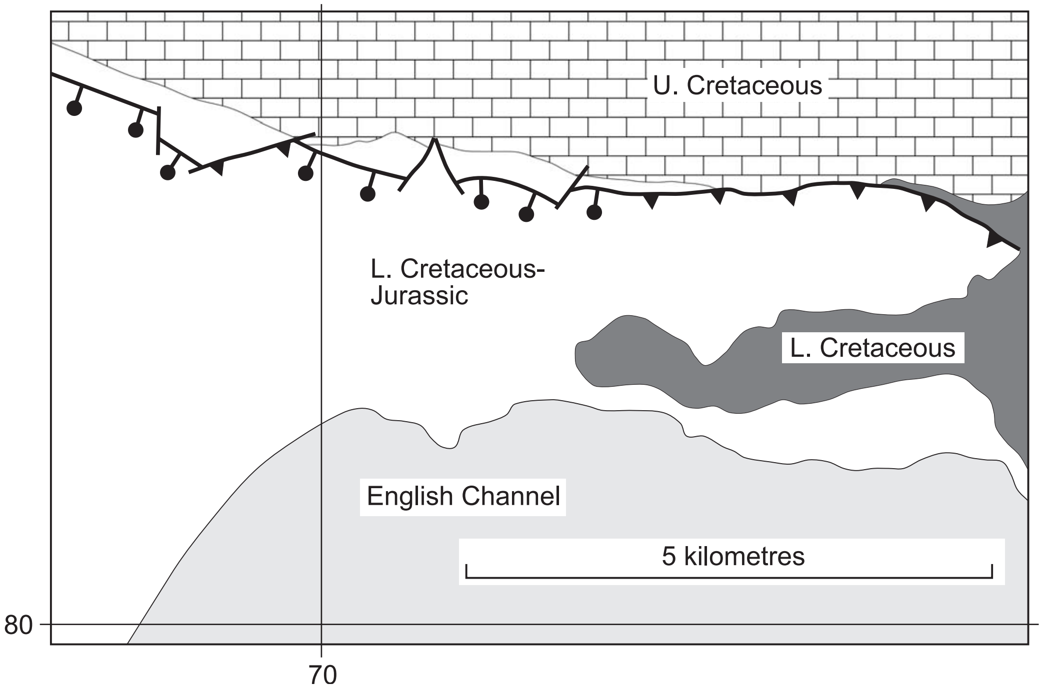

Fig. 3. Geological map of an onshore segment of the Abbotsbury–Purbeck fault system (in bold), Wessex basin, southern England. According to the concept of the null point, the switch from normal to reverse displacement (shown respectively by lines with a polka dot and barbs in the hanging wall) is interpreted as the result of compressional reactivation of a normal fault, with the magnitude of bulk reverse displacement lying within the range of the variation in pre-inversion normal displacement. Grid line labels from the UK National Grid, map from Geological Survey of England and Wales (1974).

sited adjacent to that part of a fault where normal offset was greatest.

Breakdown of displacement–length relations in inverted settings also reflects greater partitioning of strain between major faults and their wall rocks. We consider this to be a consequence of: (1) more rapid strain rates (Ziegler et al., 1995: Table 1); (2) non-coaxial fault kinematics; and (3) greater brittleness of

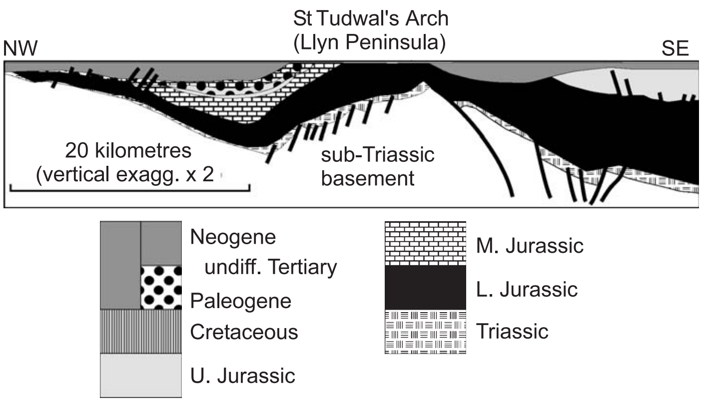

Fig. 4. Cross-section through the St. George's Channel basin, offshore North Wales, UK. It exemplifies the inversion of relief that often accompanies basin inversion, with pronounced thickening of the syntectonic Lower Jurassic succession toward what is now an uplifted structural culmination (the St. Tudwal's Arch). From authors' own data.

Fig. 5. Folded layers in the hanging wall of a normal fault juxtaposing Triassic rocks in its footwall with a hanging wall sequence of Lower Jurassic strata. Upright, near-isoclinal folds in the Jurassic layers bear witness to the horizontal shortening of these rocks. Retention of net normal displacement on the fault indicates that either (1) this outcrop is located on the normal-displacement side of the null point or (2) the fault is unreactivated but has acted as a steep buttress against which the hanging wall succession was flattened.

lithologies due to cooling and smaller confining pressures in inverting basins (attributable to the lithostatic pressure reduction that accompanies exhumation; see Corcoran and Dore, 2002). Whilst there are few published data to test this hypothesis at present, the author's observations from the Wessex and Bristol Channel basins, southern England indicate that a host of wall rock structures proximal to reactivated faults accommodate significant strain during their inversion:

1. 'Pinnate' joints and vein systems (cf. Hancock, 1985).
2. Pinnate stylolitization and fracture cleavage.
3. Buckle folds (Fig. 4).
4. Diffuse thrust fault zones comprising many closely spaced slip surfaces, each accommodating up to a few centimetres displacement.

The above description of the null point concept implies that compressional reactivation during inversion utilizes the same fault surface, or fault zone, as was active during the earlier extensional phase. However, Andersonian theory suggests that normal faults are unfavourably steeply inclined to be reactivated easily in compression. Case studies from inverted basins indicate that this theoretical problem is circumvented by the development of the so called 'breakback' or shortcut fault (e.g., Powell, 1989), and/or by reduction of effective stress along faults undergoing reactivation. Shortcut faults are low-angle thrusts splaying into the footwall of a reactivated normal fault. Characteristically, they incorporate a slice of the footwall of an inverted basin in their hanging wall and therefore result in inliers of basement bounded on one side by a thrust, and on the other by a normal fault (Fig. 6). Like the null point, they can provide diagnostic evidence of sedimentary basin inversion in areas with minimal seismic or well data.

5. Mechanics of inversion and the role of fluid overpressure

Upper crustal faulting is dominated by reshear or fault reactivation under conditions of so-called Byerlee friction. Based on a considerable body of experimental rock mechanics data, Byerlee (1978) showed that nearly all rocks share the same frictional proper-

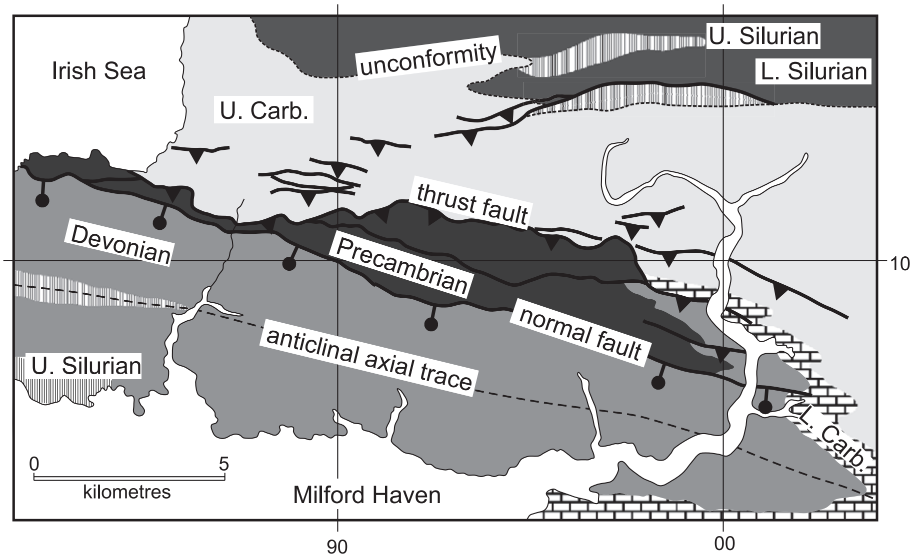

Fig. 6. Geological map showing details of fault geometry and kinematics at the northern margin of an inverted half-graben, SW Wales, UK. A fault-bounded slice, or horse, of Precambrian basement rocks separates the Devonian fill of the half-graben from its footwall (now covered by Upper Carboniferous rocks). The northern margin of the Precambrian basement horse is defined by a low-angle thrust (barbs in hanging wall; note pronounced southward V-ing into a N–S valley), whereas its southern margin is marked by a steeper normal fault (lines with polka dots in hanging wall). The normal fault defined the margin of the Devonian half-graben but, during late Carboniferous inversion, shortening was accommodated by a shortcut thrust that propagated into the footwall of the half-graben. This thrust is interpreted to have brought the Precambrian horse to the present surface from some 4 km depth where initially it formed the edge of the footwall to the Devonian half-graben (Powell, 1989; Hayward and Graham, 1989). Fault reactivation is evident also along the Silurian/Carboniferous unconformity, in the NE part of the map. Here, several pre-Carboniferous faults truncated by the unconformity were reactivated during late Carboniferous inversion such that they now offset the unconformity by several hundred metres. Grid line labels from the UK National Grid, maps from Geological Survey of England and Wales (1911, 1912).

ties with a failure criterion which may be approximated by Amonton's Law:

$$\tau = \mu(\sigma_n - P_F) \tag{1}$$

where τ and σ_n are respectively shear and normal stress acting on the plane, P_F is formation fluid pressure and the frictional coefficient μ = c. 0.84 (range = 0.60–0.85). On a Mohr diagram (Fig. 7), Byerlee friction may be understood in terms of an imaginary volume of brittle rock at Coulomb failure throughout, containing a network of identical faults in all possible orientations. Fixing σ_3, the minimum principal normal stress, as the stress ratio (σ_1/σ_3) is increased, the first rupture to occur will take place on a pre-existing fault oriented at an angle θ = c.25° to σ_1. Such a fault is oriented at the optimum angle for reactivation and, compared to faults oriented at all other angles to σ_1, stress ratio at the moment of its reactivation attains a minimum positive value.

For more realistic scenarios in which pre-existing faults lie within a restricted range of orientations, the differential stress required to induce reactivation increases rapidly. Thus, at angles of ± c. 15° to the optimum angle θ, stress ratio at the moment of fault reactivation increases by c. 50%, rising to infinity at angles of 2θ (Sibson, 1985). However, evidence from modern seismicity indicates clearly that severely misoriented faults (i.e., fault planes lying outwith Sibson's envelope of orientations within ± c. 15° to θ) are reactivated persistently. In an inversion setting for example, Chung et al. (1995) document M_S = 5.5 and 5.7 earthquakes with large components of reverse kinematics on pre-existing faults beneath a Mesozoic to Late Tertiary extensional sedimentary basin in eastern China. Given the roughly E–W to ENE–WSW bearing of the present-day compressional stress field in their study area, fault strikes of 126° and 41° obtained from their solutions suggest θ angles of 40–45°.

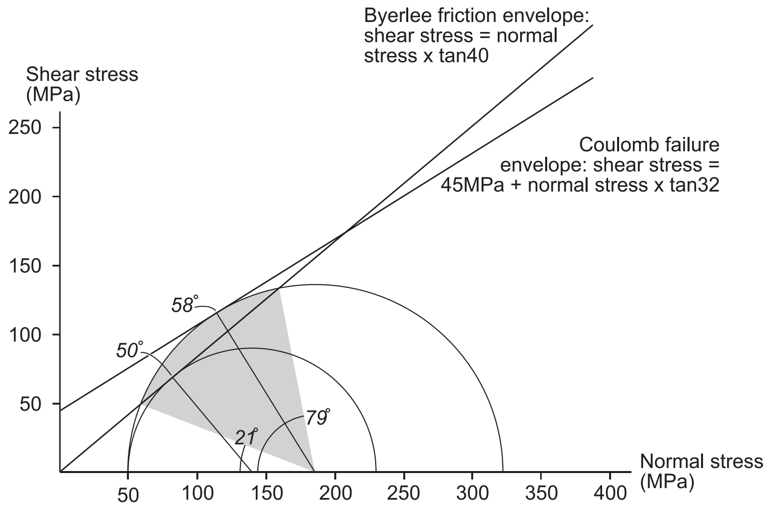

Fig. 7. Byerlee friction and Coulomb failure envelopes on a Mohr diagram. Fixing σ_3, the minimum principal stress, at 50 MPa, the graph shows that the maximum differential stress ($\sigma_1 - \sigma_3$) that this rock can sustain is 270 MPa (represented by the larger Mohr circle). When differential stress attains this magnitude, the rock will fail along either (1) a neoformed fracture oriented at an angle $\theta = 29°$ to the maximum principal stress axis (i.e., $2\theta = 58°$) or (2) an extant fracture lying within the range $\theta = 10.5° - 39.5°$ indicated by the grey-shaded sector of the Mohr circle. Thus, whether failure takes place according to Coulomb or Byerlee criteria depends on the availability of extant fractures oriented at favourable angles to the axis of maximum principal stress. Note that, in principle, this rock may fail at the reduced differential stress ($\sigma_1 - \sigma_3$) = 180 MPa represented by the smaller Mohr circle. However, this would happen only in the unlikely event that an extant fracture was oriented at the specific angle $\theta = 25°$.

In terms of epicentral depths, evidence from modern earthquakes also confound rock mechanical theory. Taking the parameters used to construct the envelopes in Fig. 8, the depth of the cross-over of the friction and failure envelopes is approximately 7 km (assuming a rock density of 2600 kg m^{-3}). Yet, compressional fault reactivation is common to depths of 20 km (Sibson and Xie, 1998: Table 1). It is clear, therefore, that fluid overpressure must play an important role in reducing effective stress such that stress ratio at the moment of fault reactivation approaches infinity (i.e., σ_3 is zero or tensile).

Virtually no upper crustal deformation takes place in the absence of fluid overpressure. Defined simply as pore fluid pressures greater than would be calculated from a hydrostatic gradient (c. 10 MPa/km), overpressure exercises profound influence over the mechanics of deformation through the reduction it brings about in effective stress (difference between pore fluid pressure and applied stress). Thus, in Hubbert and Rubey's famous experiment, best repeated many times, an emptied beer can straight from the freezer, when inverted on a gently inclined smooth surface, will initially not slide down it. Subsequently, the air trapped inside gets warmer and expands sufficiently to reduce effective stress at the interface between the beer can and the inclined surface, at which point the can slides down.

As well as its importance in structural geology, overpressure also inhibits siliciclastic compaction, possibly leading to the retardation of vitrinite reflectance (Carr, 1999), the primary tool for assessing thermal maturity of petroleum source rocks. Overpressure is most usefully expressed in terms of the fraction of lithostatic pressure it is capable of supporting. Hubbert and Rubey (1959) defined their λ factor to describe magnitude of overpressure as a fraction of lithostatic pressure:

$$(P_F - P_H)/\rho g z \qquad (2)$$

where P_F is formation fluid pressure, P_H is the hydrostatic fluid pressure and $\rho g z$ is the product of rock density, gravitational acceleration and burial depth. A λ of zero signifies hydrostatically pressured fluids, whilst $\lambda = 1.0$ is not sustainable since it implies that

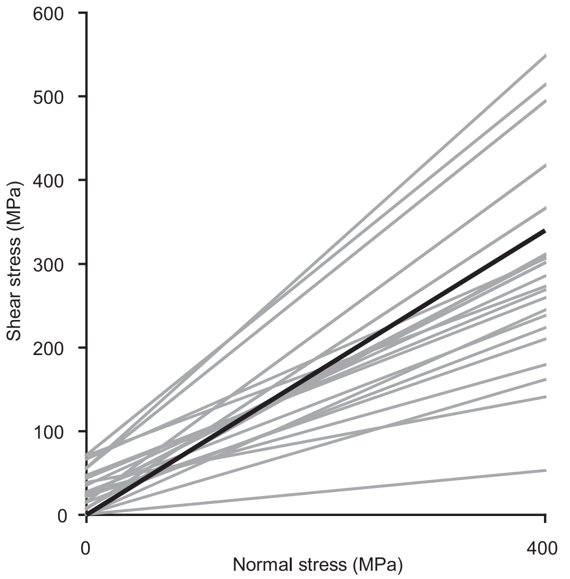

Fig. 8. Envelopes defining the range of conditions for Coulomb failure (grey lines) and Byerlee friction (bold black line). Failure envelopes are based on data from experiments on a variety of sedimentary and igneous–metamorphic lithologies given in Goodman (1989: Table 3.3). The graph demonstrates that below normal stresses of c. 200 MPa (i.e., uppermost c. 8 km of the crust), Byerlee friction (i.e., reactivation of extant faults and fractures) exerts the dominant control over failure in the crust.

fluid pressure exceeds rock strength. However, the existence of hydrothermal breccias—triaxial networks of closely spaced, contemporaneous vein systems that resemble 3D jigsaw puzzles—and vertical mineral fibres in horizontal veins (Fig. 9; see also Cosgrove, 1993) attest to brief moments when fluid pressure must have attained lithostatic pressure, at least locally.

Perhaps more than in any other setting, overpressures are potentially large during sedimentary basin inversion. In what has been coined the ‘sponge effect’, basin inversion can be likened to wringing out a sodden sponge, liberating unknown volumes of water previously trapped in sedimentary pore spaces. The 200-km-long Cretaceous Mother Lode vein system of California provides a graphic example of the volumes involved. Given the low solubility of quartz in water, more than 10^6 m^3 of fault-hosted vein fill per kilometre strike length suggests that something like 10^9–10^{10} m^3 (6.3–63 billion barrels!) of aqueous fluid flushed through the vein system per kilometre strike length (Sibson, 1995).

Field observations lending support to the association of overpressure with rock deformation in general, and sedimentary basin inversion in particular, include the following:

1. Textural characteristics of hydrothermal vein systems associated with ancient reverse fault zones exhumed from depths comparable to the base of the seismogenic zone (Sibson, 1990).
2. Synkinematic vein arrays in inverted basins in which vertical mineral fibres indicate vertical dilation, i.e., fluid pressure momentarily exceeds lithostatic load (e.g., Fig. 9; Cosgrove, 1993).

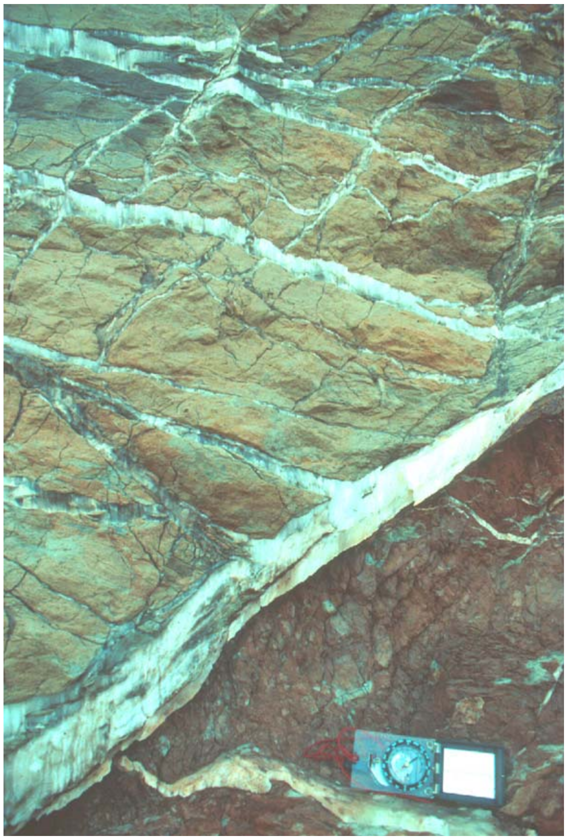

Fig. 9. Contemporaneous gypsum veining and a reactivated normal fault juxtaposing latest Triassic mudstone in its hanging wall with late Triassic sandstone in the footwall. Individual generations of vein fill may be traced continuously from the fault plane to the hanging wall succession into which the veins propagated at sub-horizontal attitudes. Vertical mineral fibres in all the veins testify to supra-lithostatic fluid pressures at the moment of crack propagation.

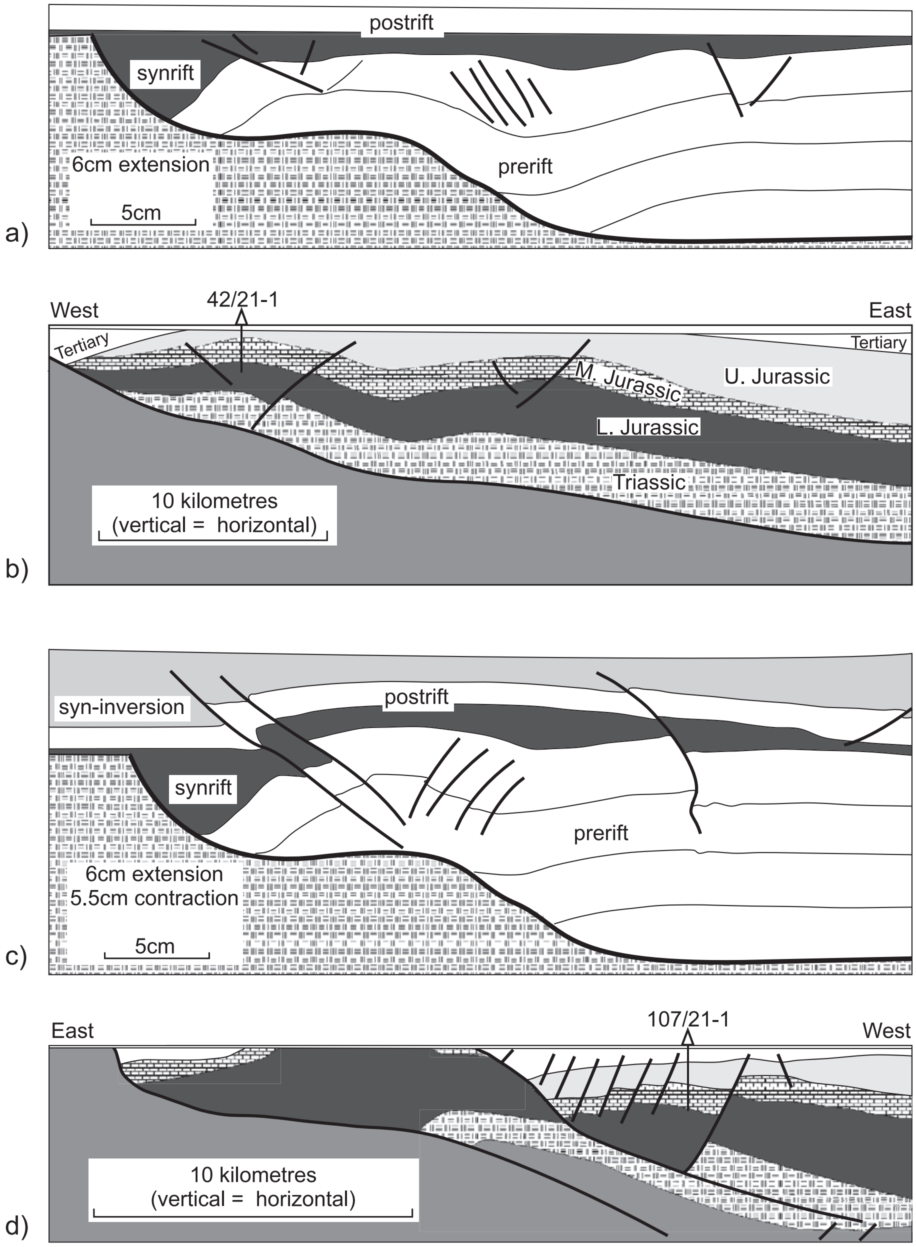

Fig. 10. Comparison of the results of experimental models of fault reactivation (a, c) with cross-sections from opposing margins of the UK–Irish St. George's Channel basin (b, d; same stratigraphic fill in each). The model and cross-section in (a) and (b) exemplify typical deformational geometry resulting from translation of a hanging wall succession down a ramp-flat detachment. The model in (c) displays deformational geometry resulting from compressional reactivation of the detachment in (a), with shortening strain of some 90% initial extension (6 cm extension, 5.5 cm shortening). Cross-section (d) is interpreted also as a product of significant compressional reactivation of a ramp-flat detachment. It shows pronounced tightening of a thick Lower Jurassic syntectonic sequence that accumulated in a rollover monocline above a major inflection in the underlying detachment. Experimental models after McClay (1995); cross-sections from authors' own data (b) and (d) (Turner, 1996).

3. Transitory surface effusions following earthquake activity, such as CO_2-rich springs (Irwin and Barnes, 1975), sandblows (e.g., Yeats et al., 1997: Figs. 9-34, 12-6, 12-9), hydrocarbon discharge (e.g., Hamilton et al., 1969) and the sometimes dramatic outpourings of warm groundwater along active fault lines (e.g., Briggs and Troxell, 1955; Bureau of Reclamation, 1976).
4. Brecciation processes in fault zones (Sibson, 1986).
5. Spatial distribution of aftershocks (e.g., Wetmiller et al., 1984; Eberhart-Phillips, 1989; Sibson, 1990).
6. Association of active reverse faults with overpressured basins (e.g., Western Transverse Ranges, California: Yeats, 1983; Yeats et al., 1997).
7. Theoretically unfavourable inclination and plan-view orientation of many strongly reactivated faults, and the 'selectivity' of reactivation of only one or a few faults within sets of geometrically and kinematically similar fault planes (e.g., Abbotsbury–Purbeck Fault, southern England: Fig. 10; see also Sibson, 1995).
8. In situ fluid pressures equivalent to $\lambda \approx 1.0$ inferred from the toes of actively deforming accretionary wedges in Barbados and the Makran (e.g., $\lambda = 0.97$: Davis et al., 1983).

Fault valve behaviour (Sibson et al., 1988) was formulated initially as an explanation for the incremental nature of mesothermal gold–quartz deposits. It describes cycles of:

1. overpressure build-up and progressive reduction in effective stress across a pre-existing fault,
2. fault reactivation, rupture and breach of a permeability seal,
3. rejuvenation and neoformation of fracture permeability in the wall rocks leading to sudden and rapid diminution of fluid pressure and
4. re-establishment of fault seal followed by the next cycle of overpressure build-up.

The idea of faults acting as pressure valves has found wide acceptance in inversion studies because it provides a plausible mechanism for the episodic occurrence of near-lithostatic fluid pressures, a prerequisite for the compressional reactivation of steeply inclined faults. Implicit in the model is that faults act as permeability seals during aseismic intervals, but that they are very effective conduits for vertical fluid transport during coseismic episodes. Support for these implications has since come from studies of fluid transport in inverted settings (e.g., Sullivan et al., 1994; Andrews et al., 1996; Gluyas et al., 1997) and the sealing capacities of intra-reservoir fault rocks (e.g., Ottesen Ellevset et al., 1988; Fisher and Knipe, 1998; Gibson, 1998). Gudmundsson et al.s' (2001) analysis of Mode 1 fracture networks within the damage zone of an active fault on Iceland indicate that the veins propagated under conditions in which fluid pressure exceeded the minimum principal stress by 20 MPa. Furthermore, their modelling suggests that vertical fluid flow is favoured strongly during coseismic episodes, with the rate of fluid flow along vertical veins exceeding that along horizontal veins of the same dimensions by a factor of 6.

Basin inversion is promoted by large differential stress, low frictional strength and/or processes that serve to reduce effective frictional strength (e.g., overpressure), and the presence of weak layers like salt and unlithified shale along which the shortening cover can detach itself from basement. In the inverted basins of the UK, salt is omnipresent and exerts strong influence on faulting in the supra-salt cover succession. Stewart et al. (1996) present a large number of seismic-based cross-sections through inverted UK salt basins that exemplify its role in vertically partitioning deformation between the sub-salt and supra-salt successions.

6. Insights from analogue modelling

The familiar results of analogue models demonstrate the impressive success achieved by modern modelling methods in replicating a diverse range of deformational geometries (Figs. 10 and 11). Problems of scaling are acknowledged explicitly. For example, laboratory strain rates of c.2.8×10^{-4} s^{-1}, or 10% strain/60 min (calculated from strain rate of 4.16×10^{-3} cm s^{-1} in a 150-cm-long deformation rig: McClay, 1995), contrast with geological strain rates of 10^{-13}–10^{-14} s^{-1} (10% strain/30–300 kyears). Also, in spite of the introduction of different grades of sand and mica to the models, it is impossible to

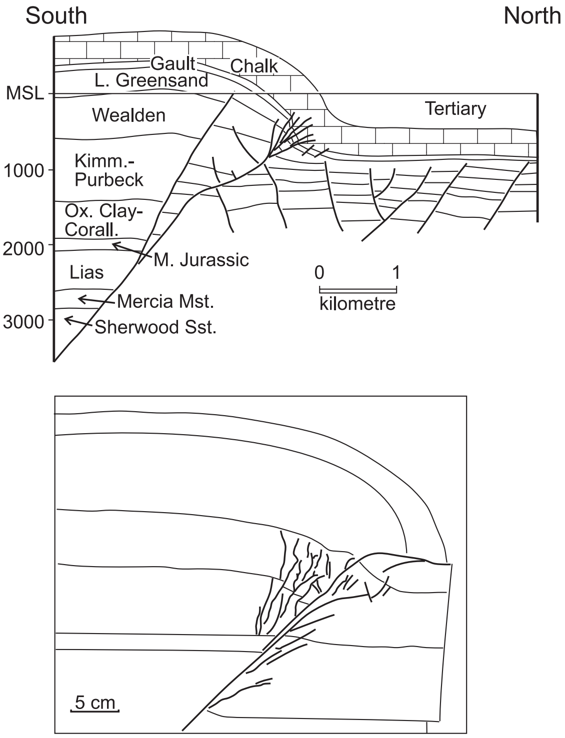

Fig. 11. Comparison of a cross-section through the Abbotsbury–Purbeck fault system of the Wessex basin, southern England with the results of an experimental model. The model simulates the compressional reactivation of an extensional fault-propagation fold on a planar fault dipping 45°. Shortening strain is some 350% with 12cm shortening subsequent to an initial 3.5-cm extension. Note the selectivity of fault reactivation in the Wessex basin—extensional faults in the northern half of the section remain more or less unreactivated with most of the shortening in this part of the basin accommodated by a major shortcut fault in the footwall of the Abbotsbury–Purbeck fault. The Wytch Farm oilfield occupies the footwall of one of these unreactivated normal faults. Wessex basin cross-section after Butler (1998); vertical scale in metres below mean sea level (MSL). Experimental model after Mitra (1993).

simulate fully the orders of magnitude range in scale on which the crust exhibits anisotropic behaviour. To deal with the strain rate problem in their models of whole-lithosphere deformation, Vendeville et al. (1987) employ silicone putty and honey to replicate the near-Newtonian behaviour of mantle lithosphere and aesthenosphere at geological strain rates.

Many analogue models employ a rigid footwall in experiments simulating listric faulting and/or fault reactivation. This reflects the inability of the dry sand and mica layers to undergo displacement along low-angle surfaces unless the fault planes are near-frictionless. The assumption of a rigid footwall is, however, a poor approximation to reality in most inverted basins. Footwall deformation during inversion is variously incorporated in experiments that simulate non-rotational rigid basement blocks (Koopman et al., 1987), arrays of 'domino' faults that decrease and increase their inclination to the horizontal during extension and contraction (Buchanan and McClay, 1992), and reactivation of normal faults on which extension was terminated prior and subsequent to breakthrough of the major fault (Mitra, 1993). According to (1) the inclination of the faults at the onset of inversion and (2) the degree of decoupling of the hanging wall and footwall, these experiments reproduce a range of footwall and cover sequence (i.e., postrift) deformation, including flexural slip folding and en-echelon shortcut faults. The results closely match seismic and field examples from inter alia the southern North Sea (e.g., Nalpas et al., 1995), western UK (e.g., Figs. 4 and 10; Tucker and Arter, 1987; Bulnes and McClay, 1998) and the Wessex basin, southern England (e.g., Fig. 11; Butler, 1998; Underhill and Paterson, 1998), and they demonstrate how the extensional architecture exerts a profound influence over the subsequent inversion geometry.

Eisenstadt and Withjack (1995) use wet clay to replicate scenarios in which hanging wall and footwall are involved in the inversion. They examine the relationship between the magnitude of shortening strain in inverted basins and the spectrum of structural geometries that evolve with progressively more shortening. The rationale for this work comes from the extreme difficulty in placing even the most qualitative constraints on shortening strain in inverted basins. Seismic interpreters may refer to their basins having undergone "mild" inversion when, infact, seemingly modest amounts of buckling or reverse fault displacement may reflect substantial shortening. Consequently, most methods for quantifying shortening strain in inverted basins yield gross underestimates. The reasons for this are twofold:

1. If it is assumed that all shortening is accommodated by reverse-sense fault displacement, formerly extensional faults will need to have all their

normal-sense displacement cancelled before they begin to accumulate net reverse offset.
2. In reality, a plethora of small, sub-seismic scale structures will accommodate large amounts of shortening strain alongside major fault displacement (e.g., pinnate joints, veins, small faults, stylolites and cleavage; flexural slip folds). Furthermore, inverted basins will undergo 'horizontal compaction' in which significant shortening strain is accommodated by volume (porosity) reduction.

The apparent nature of fault displacement in inverted settings, and the partitioning of strain between major faults and small-scale structures, violates some of the fundamental assumptions of cross-section balancing (Dahlstrom, 1969; Hossack, 1979). Consequently, as a technique for quantifying bulk strain across inverted basins, balanced cross-section restoration has not found wide acceptance. Expressing inversion magnitude as a percentage of the original extensional strain, Eisenstadt and Withjack (1995) demonstrate that seismic scale manifestation of compressional deformation is mild even when inversion has attained 100%. Thus, classic inversion structures exemplified by seismic examples from the southern North Sea, the Irish Sea basins and SE Asia may record extreme inversion of 200% or more (cf. Song, 1997).

7. Role of inversion in sedimentary basin exhumation

This section demonstrates that whilst modern methods permit the cooling history and exhumation magnitude of deeply eroded basins to be quantified readily, identifying the large scale processes responsible for their uplift and erosion remains an outstanding problem. Exhumation describes the positive feedback between uplift and erosion during which formerly deeply buried rocks are exposed at the surface. Traditional geological training is founded on observing and interpreting evidence of Earth history preserved in the rock record. Perhaps the greatest challenge in investigating uplift and erosion is that such studies are usually attempting to make reconstructions of what is missing from the rock record.

The palaeobiology community has long grappled with conceptual difficulties posed by the inherent incompleteness of the palaeontological record. However, problems of 'missing' evidence are relatively new in other branches of geoscience. The problem is particularly acute in structural geology and tectonics—in the absence of direct dating methods (e.g., Ar–Ar dating of muscovite-bearing fault rocks), it is possible only to bracket the age of a deformational episode between the youngest rocks it effects and the oldest rocks that post-date it. Due to the difficulty in constraining an absolute datum, very few studies are able to measure ancient uplift (e.g., Abbott et al., 1997; House et al., 1998). Furthermore, in petroliferous basins, exhumation is actually a far more useful parameter to constrain because it leads directly to cooling and lithostatic pressure release, with attendant implications for petroleum generation and retention. In the absence of erosion, uplift on its own results in neither cooling nor pressure decrease.

The main processes driving the exhumation of sedimentary basins are epeirogeny and basin inversion. Epeirogeny describes the uplift of broad regions of continental interiors driven by plumes (e.g., Nadin et al., 1995), basaltic underplating (e.g., Brodie and White, 1994), mantle delamination (e.g., Platt and England, 1993), post-glacial isostasy (e.g., Lambeck, 1991), intra-plate stress (e.g., Cloetingh, 1988) and thermal-isostatic effects associated with rifting and/or oceanic opening (e.g., Gallagher and Brown, 1999).

The widespread use of thermal maturity proxies, particularly apatite fission track analysis, has facilitated major advances in reconstructing that part of the history of a sedimentary basin removed by erosion. Comprehensive review of the theory of fission track analysis, and examples of geological applications, is provided by Gallagher et al. (1998). Apatite fission track analysis yields detailed information in the 60–120 °C temperature window in which petroleum source rocks undergo catagenesis. Fission tracks are microscopically visible damage paths in the crystal lattice induced by spontaneous fission of ^{238}U. The tracks form at an initial length with little variability, and immediately begin to anneal (shorten) at a rate dependent on ambient temperature and apatite chemistry (particularly Cl content). If the temperature falls, an individual track is frozen at the shortened length it attained at maximum palaeotemperature. Thus, measurement of the lengths of fission tracks in crystals whose initial abundance of ^{238}U is obtained independently allows the absolute timing of cooling and the maximum palaeotemperature prior to cooling to be determined. Because new tracks are constantly being created throughout geological time, a sample that cools after having reached a high temperature in the past will have a population of both long and short tracks. The long tracks post-date cooling, whereas the short tracks formed before cooling and their length will be proportional to the maximum palaeotemperature achieved.

Sedimentary basins in which the fill accumulates in response to long-term subsidence and burial yield the simplest profiles of fission track length and age versus depth. Because the entire succession will be at its maximum temperature today, shallow samples exhibit relatively high track ages and narrow length distributions, with track length distributions becoming progressively younger and broader with depth (e.g., Otway basin: Gleadow and Duddy, 1981). By contrast, fission track ages and length distributions in samples from inverted settings indicate that the succession has been hotter in the past. Typically, inverted basins exhibit pronounced inflections in the rate of change of fission track age and distribution. These mark the base of the so called partial annealing zone, the transition from partial to total annealing of fission tracks formed prior to the onset of cooling (e.g., Green et al., 1995). Conversely, datasets displaying similar fission track ages over large depth intervals indicate small partial annealing zones consistent with rapid cooling (e.g., Arne et al., 2002).

Many basins experience palaeotemperatures in excess of apatite fission track stability. In these scenarios, vitrinite reflectance often provides a record of the higher temperatures (total annealing in apatite $\approx 0.7–0.9\%$ R_o: Bray et al., 1992) and numerous studies of inverted basins have integrated apatite fission track analysis with vitrinite reflectance to great effect (e.g., Bray et al., 1992; Green et al., 1995; Kamp et al., 1996; Arne et al., 2002). Furthermore, vitrinite reflectance provides an independent constraint on maximum palaeotemperature in the partial annealing zone, where mixing of pre- and post-cooling fission tracks leads to potentially complex track length distributions.

To date, therefore, studies of inverted basins show that the chief impact of sedimentary basin inversion on apatite fission track data is erosion-induced cooling from higher temperatures generated by deeper burial. Despite the critical role of synkinematic fluids in promoting fault reactivation by reduction of effective stress, their influence on kinetically dependent thermal maturity proxies like apatite fission track analysis and vitrinite reflectance has, until recently, eluded detection. However, the thermal signature of synkinematic hydrothermal fluid pulses have recently been detected in studies from the British continental shelf and Atlantic margins (e.g., Parnell et al., 1999; Middleton et al., 2001; Green et al., 2001). Here, short-lived episodes of migration of anomalously hot reservoir fluids are recorded in the relatively shallow subsurface.

Apatite fission track and vitrinite reflectance data from the 0–1500 m interval in the Apley Barn borehole, southern England are typical of the evidence for hot fluids (Green et al., 2001). They show a highly non-linear Tertiary geothermal gradient, reminiscent of the temperature versus depth profiles modelled by Ziagos and Blackwell (1986), with a thermal 'spike' centred on a major permeability boundary (unconformity spanning the hiatus Westphalian to Rhaetian, c. 300 m subsurface). Similar effects are reported by Lampe et al. (2001) from the Rhine Graben. They ascribe locally elevated vitrinite reflectance anomalies to episodic fluid flow through shallow (1000–1500 m) aquifers during Alpine-related basin inversion. Fluid inclusions data from the Rathlin basin, British Atlantic margin (Middleton et al., 2001), indicate that Tertiary fluid pulses attained temperatures of between 144 and 177 °C, up to 58 °C higher than ambient temperatures. The Rathlin basin data suggest that, for a large discrepancy to exist between temperature obtained from vitrinite reflectance versus that from fluid inclusions, fluid-driven heating must have been of very short duration (< 1 ka).

In the settings from which the examples cited above are taken, the thermal effects of hot fluids emanating from deep basins onto adjacent flanking highs are well preserved, due probably to the relative stability of these regions during the Mesozoic–Tertiary. In inverted basins, shallow thermal effects are obviously unlikely to be preserved following their exhumation. However, results from numerical models of lithosphere rheology show that marginal basins adjacent to inversion highs are an integral product of the inversion process (Nielsen and Hansen, 2000).

Given the high preservation potential of marginal basins, it is within them that we should expect to find the evidence for hot fluid pulses. In this light, Tertiary thermal pulses in southern England (Green et al., 2001) are associated directly with the Alpine inversion of the proximal Wessex and Weald basins. They are interpreted as the record of so called seismic pumping whereby fluids are expelled from inverting basins during increments of regional compression and fault valve activity.

On their own, the thermal maturity proxies discussed above rarely provide diagnostic evidence of sedimentary basin inversion. Only with an understanding of the structural context of sample localities can the results be interpreted in terms of inversion. The coincidence of the timing of cooling with known inversion episodes (e.g., Green et al., 1995), and concentration of maximum cooling in samples taken from the hanging walls of reactivated faults (e.g., Arne et al., 2002), are typical observations that allow the finger to be pointed firmly at sedimentary basin inversion.

The cryptic signature of basin inversion means that interpretation of thermal maturity data in basins where the effects of epeirogeny and inversion are superimposed on each other is particularly challenging. The western UK basins provide a good example of the problem. Interaction between uplift related to the Cretaceous–Paleocene opening of the North Atlantic and Neogene basin inversion is illustrated by the complex denudation pattern of Britain. It is maximal over the submergent East Irish Sea basin, some 500 km from the Atlantic margin, where Late Mesozoic–Cenozoic exhumation was responsible for the erosion of up to 3 km of Upper Jurassic and Cretaceous section (Lewis et al., 1992; Ware and Turner, 2002). Given that the exhumation maximum is centred on the East Irish Sea basin, a long way south of the extrusive Tertiary Igneous Province, the regional pattern of exhumation is probably a product of the combination of (1) northward-declining Alpine-driven inversion and (2) southward-declining underplating conjectured beneath a wide region extending far beyond the area of extrusive volcanism (Brodie and White, 1994). To date, however, the emphasis has been on underplating as the chief process driving the exhumation of the western UK basins.

Brodie and White (1995) argue that observed shortening in the East Irish Sea basin is significantly less than the c. 30% strain they contend is needed to account for the 3-km missing section. However, as discussed above, measurement of inversion-related shortening strain in exhumed basins is problematic because (1) fault heaves in inverted basins are apparent and, therefore, not a good record of horizontal shortening and (2) strain partitioning during basin inversion means that significant shortening will be accommodated by porosity reduction and pure shear in the wall rocks of major faults. Yet, despite the promotion of the underplating hypothesis in the western UK basins, direct evidence (e.g., from wide-angle seismic studies of lower crustal velocities) of several kilometres of underplated basaltic melt beneath the Irish Sea basins is absent. Furthermore, apatite fission track data suggest that the main cooling episode in the East Irish Sea basin was Early Cretaceous, some 60 Ma prior to the activity of the Iceland Plume (Duncan et al., 1998).

Novel analysis and interpretation of sedimentary velocity data makes it possible to discriminate between inversion and epeirogeny in basins with complex histories, like those of the western UK. The long wavelength (>2 km) form of vertical sonic transit time versus depth profiles responds chiefly to compaction-driven porosity reduction. Consequently, it is an effective measure of the former maximum burial depth of exhumed sedimentary successions in basins where transient heating elevated thermal maturity proxies without concomitant burial increase. The chief control over sonic velocity in undeformed sedimentary successions is burial-controlled mechanical compaction due to porosity reduction:

$$\Delta t = \Delta t_0 \exp(-bx) + c \tag{3}$$

where Δt = sonic transit time at depth x, b is compaction coefficient per unit lithology, Δt_0 = surface transit time of uncompacted sediment and c is the sonic transit time of the rock matrix. Following Heasler and Kharitonova (1996), Ware and Turner (2002) use a single, average compaction coefficient to describe porosity reduction with depth in heterolithic sedimentary sequences in the East Irish Sea basin (see also Steckler and Watts, 1978; Tosaya and Nur, 1982; Castagna et al., 1985; Han et al., 1986; Marion et al., 1992). To compute the magnitude of exhumation, they extrapolate the best-fit transit time versus depth curve above the erosion surface (i.e., unconformity or the present basin floor) to the level at which transit time equals that of uncompacted sediment (591 μs/m or 180 μs/ft; sonic velocity = 1695 m/s: Magara, 1976).

Exhumation in the East Irish Sea basin varies significantly between adjacent fault blocks. This short-wavelength pattern of denudation is interpreted as a record of locally variable inversion, which tends to be focused in the hanging walls of major faults, superimposed on a much longer wavelength denudation pattern generated by epeirogenic processes during Atlantic opening (Fig. 12). Comparable methodology is applied by Hillis (1995) and Japsen (1998), using mainly Upper Cretaceous Chalk velocities to map burial anomalies of c. 1 km around the inverted margins of the North Sea and contiguous basins.

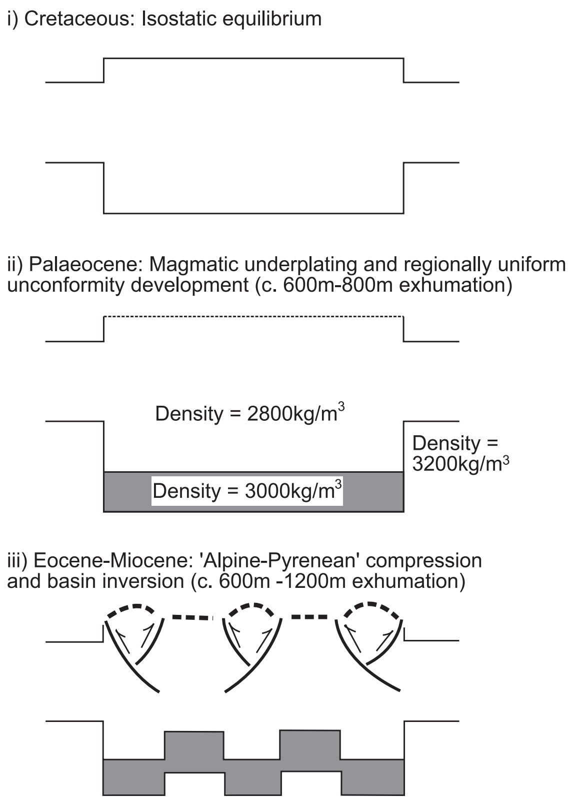

Fig. 12. Schematic model to explain the short-wavelength variation in exhumation recorded by sonic velocity data from exhumed Triassic rocks in the East Irish Sea basin, UK. Here, exhumation magnitude is too large to be accounted for by basin inversion on its own. Hence, the top Triassic unconformity is interpreted as a product of locally variable basin inversion superimposed on uniform epeirogenic uplift. After Ware and Turner (2002).

8. Implications for petroleum geology

Intuitively, sedimentary basin inversion is bad news for petroleum geologists:

- Sealing horizons are removed and/or their effectiveness is severely reduced by brittle failure (e.g., Kjemperud and Fjeldskaar, 1992; Sales, 1993).
- Reactivated faults may become conduits for hydrocarbon leakage to the surface (e.g., MacGregor, 1995).
- Potentially attractive reservoirs may be downgraded due to overcompaction (e.g., Olaussen et al., 1984) and/or synkinematic diagenetic processes such as stylolitization (e.g., Walderhaug, 1992) and secondary cementation (e.g., quartz: Berglund et al., 1986; anhydrite and/or baryte: Gluyas et al., 1997).
- Cooling of source rocks may cause hydrocarbon generation to cease (e.g., Jensen and Schmidt, 1993; Underhill and Stoneley, 1998; Arne et al., 2002) and/or they may have exceeded their generative capacity during former deeper burial (e.g., Theis et al., 1993).
- Lithostatic pressure reduction during subsequent exhumation may cause oil accumulations to exsolve gas, pushing down the oil–water contact and spilling oil (Dore and Jensen, 1996).
- Postulated oil reservoirs may previously have been buried at depths at which oil cracked to gas (e.g., Theis et al., 1993).
- Regional tilting during inversion results in changes to trap configurations, possibly destroying subtle one-way dip closures (e.g., Macgregor, 1995; Kjemperud and Fjeldskaar, 1992; Dore and Jensen, 1996; Francis et al., 1997).

Whilst inversion is but a subset of the range of processes that can trigger exhumation, the implications of regional exhumation for petroleum prospectivity are profound and they are reviewed comprehensively by Dore and Jensen (1996) and Dore et al. (2002). A host of different factors are shown to influence prospectivity during and following the exhumation of petroliferous sedimentary basins. Key processes include reservoir underpressuring due to pore dilatancy and fluid volume reduction, generation of overpressure due to tectonic compression and aquathermal effects, and brittle versus ductile behaviour of mudstone caprocks. Despite the potentially detrimental effects of inversion, many prospective sedimentary basins have undergone severe inversion and, in the Wessex basin of southern England, Europe's largest onshore oilfield at Wytch Farm is located in an extensional palaeostructure in the unreactivated footwall to a major compressionally reactivated fault zone (Fig. 11; Colter and Havard, 1981). Conversely, a survey of the five major gas accumulations in the exhumed European Atlantic margin basins (Corcoran and Dore, 2002) shows that four of them are significantly underfilled (i.e., the level of the hydrocarbon–water contact is above the spillpoint of the structure in which the hydrocarbons are trapped). Ultimately, whether exhumation enhances or suppresses petroleum prospectivity is not clear-cut and depends largely on the post-exhumation hydrocarbon budget available through secondary and tertiary migration.

Note that several of the processes listed above can have a positive impact on hydrocarbon prospectivity. Thus, source rocks presently at sub-mature depth may have been mature prior to inversion (e.g., Jensen and Schmidt, 1993; Arne et al., 2002), high-temperature/high-pressure gas accumulations may exsolve oil by the process of retrograde condensation (e.g., Piggott and Lines, 1991), pressure reduction due to exhumation can generate large accumulations of so called basin-centred gas by methane exsolution from formation waters (Masters, 1984; Maximov et al., 1984; Cramer et al., 2002) and brittle failure of reservoirs, especially carbonates, may enhance their permeability (e.g., Daniel, 1954; Aguilera, 1980; Bourne et al., 2001). In the Norwegian North Sea, a late dry gas charge that entered earlier formed oilfields (e.g., Troll, Sleipner) is attributed to the sudden drop in fluid pressure that resulted from Neogene exhumation (Horstad and Larter, 1997). It has been suggested that, during periods of fluid overpressure, source rock maturation is retarded due to the suppression of endothermic volume expansion reactions (Carr, 1999). Suddenly releasing the fluid pressure will stop retardation and, in spite of temperature drops of 10–20 °C, maintenance of chemical equilibrium enables hydrocarbon generation to continue.

Natural fracture systems can have a dramatic impact on reservoir performance chiefly because they act as highly permeable fluid conduits. Compared to rift-related faulting, we consider that sedimentary basin inversion promotes maximal partitioning of strain between major faults and their wall rocks. This reflects the association of inversion with more rapid strain rates, non-coaxial fault kinematics, and greater brittleness of lithologies due to cooling and smaller confining pressures (attributable to the lithostatic pressure reduction that accompanies exhumation). Consequently, predictive modelling of naturally fractured reservoirs (e.g., Segall and Pollard, 1980; Pollard and Segall, 1987; Bourne and Willemse, 2001; Bourne et al., 2001) finds particular relevance in inverted basins.

Most models of fault–fracture interaction represent faults as frictionless surfaces within an isotropic and homogeneous linear elastic medium. Based on imposing an external stress field on the known structural geometry of the reservoir, it is possible to map the distribution of the elastic stress field in wall rocks in the vicinity of interacting fault surfaces. Segall and Pollard (1980) apply this methodology to en-echelon fault arrays with the objective of investigating the nature and extent of fracturing at the steps between contiguous segments of discontinuous fault systems. Their analysis demonstrates that, in comparison to transtensile steps where the zone of brittle fracturing is relatively small, fracture zones at transpressional steps are much larger. Whilst the size of the region of tensile failure is only arbitrarily larger in transpressional settings, the contour enclosing the zone of shear failure extends far beyond transpressional steps (Fig. 13). In the context of sedimentary basin inversion, these models indicate that in the wall rocks adjacent to faults undergoing transpressional reactivation—the kinematic norm in most inverted basins—pervasive brittle fracturing has the potential to enhance reservoir permeability considerably.

A global review of inverted petroliferous basins shows that exploration success rate is dependent on the severity and geographic extent of the inversion (Macgregor, 1995). Thus, locally inverted rifts like the Central Sumatra and Malay basins display comparable success rates to uninverted rifts, with slightly smaller maximum field sizes. This largely reflects the creation of simple anticlinal traps during mild inversion episodes, but without concomitant seal breach and resultant loss of structural integrity. By contrast, success rates and maximum field sizes in regionally inverted rifts like the West Netherlands (De Jager et al., 1993), Sole Pit (Glennie and Boegner, 1981) and Wessex (Macgregor, 1995) basins are much lower. The largest fields in these basins characteristically dominate their total reserves, an extreme example of which is provided by the Wytch Farm field containing some 98% of Wessex basin reserves (Macgregor, 1995). Severely

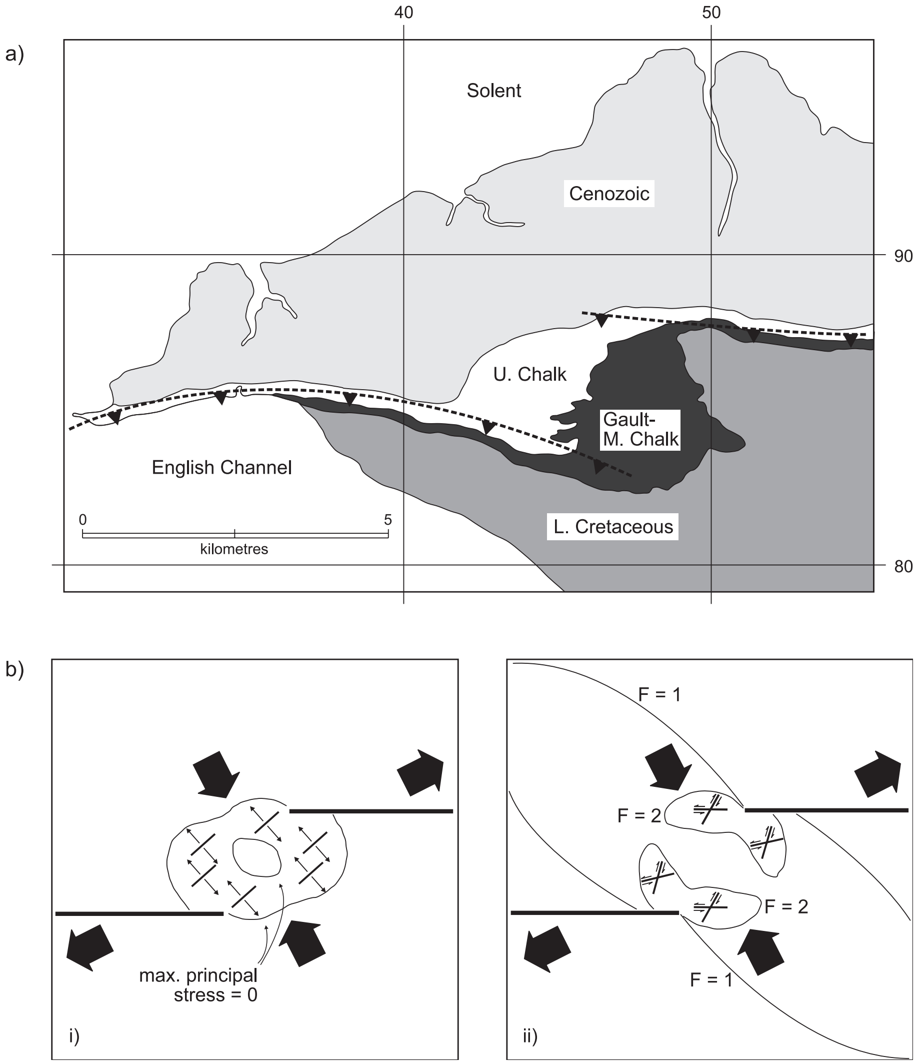

Fig. 13. (a) Geological map of a relay ramp at the overlap between en-echelon normal faults, reactivated during Cenozoic inversion of the Wessex basin, Isle of Wight, southern England (cf. Underhill and Paterson, 1998). The width of outcrop of Middle and Upper Cretaceous rocks gives an indication of their inclination, with the steepest tilts above the subsurface traces of the reactivated faults. Dashed lines mark the inferred traces of subsurface thrust faults, shown with barbs in their hanging walls. Grid line labels from the UK National Grid. Map from Geological Survey of England and Wales (1935). (b) Model of secondary fracturing near left-stepping en-echelon discontinuities subjected to right-lateral shear, comparable to the Isle of Wight example shown in (a). The model incorporates the effect of elastic interaction between cracks, with contours enclosing regions of potential tensile fracturing (i) and shear fracturing (ii) for which the shear failure condition, *F*, must be >1. The results of models like these are consistent with observations of fault-related fracturing over a wide range of scales and they serve also to predict the distribution of earthquake swarms and aftershocks at steps along en-echelon segments of active faults. After Segall and Pollard (1980).

inverted basins suffer variously from the detrimental impacts on prospectivity listed above. In particular, the 'switching off' of source rock kitchens, and the creation of tertiary migration pathways due to seal breach and reactivation of formerly sealing faults are profound.

An inevitable implication of the highly non-Gaussian distribution of field sizes in inverted basins is that most of their hydrocarbon resource has been lost through seepage and tertiary migration. A review of petroleum seeps across the UK (Selley, 1992) reveals that many of them are associated with inverted basins. Indeed, the Wessex basin has become well known for the occurrence within its Mesozoic stratigraphy of modern and ancient seeps in the hanging wall of the severely reactivated Abbotsbury–Purbeck fault zone (e.g., Parfitt and Farrimond, 1998). Furthermore, fossil hydrocarbon contacts in what are today water-saturated structures in inverted basins (e.g., East Irish Sea basin: Francis et al., 1997) testify to former petroleum accumulations that leaked away during sedimentary basin inversion.

9. Geodynamics of sedimentary basin inversion

The observations that: (1) intra-plate stress is uniform over wide areas (Zoback, 1992; see also present stress maps of Mueller et al., 2000); (2) the interval between rifting and inversion often exceeds the c. 60 Ma thermal time constant of the lithosphere (see Ziegler et al., 1995: Table 1); and (3) the repeated localization of sedimentary basin inversion (e.g., van Wees and Beekman, 2000) implies that extensional sedimentary basins comprise zones of long-term weakness where intra-plate stress becomes focused. As such, they are prone to reactivation following extensive periods of quiescence (e.g., 300 Ma: Ziegler, 1989), and often at considerable distances from plate margins (e.g., Alpine compression affected basins up to 1600 km from the orogen: Ziegler et al., 1995).

The idea that sedimentary basins are inherently weak confounds intuitive notions of the strain-hardening effect of lithospheric extension and has stimulated numerous investigations of the interplay between rheology, heatflow, stress distribution and zones of mechanical weakness during sedimentary basin evolution (e.g., Sandiford, 1999; Nielsen and Hansen, 2000). Kusznir and Park (1987) showed that lithospheric strength is strongly dependent on geothermal gradient, the length of time since the application of tectonic forces and, critically, the thickness of relatively weak quartzo-feldspathic crust. Whether the lithosphere undergoes strain hardening or softening during extension is largely a trade-off between two factors: (1) the strengthening it experiences due to replacement of weak crust by strong mantle lithosphere versus (2) the weakening effect of increasing the geothermal gradient. The rate of stretching appears to be critical. During slow stretching ($< 5 \times 10^{-15}$ s^{-1}), heat loss to the surface will keep pace with extension thereby promoting replacement of crust by mantle and consequent strain hardening. Conversely, rapid extension is dominated by increased geothermal gradient (>65 mW m^{-2}), localizing deformation into narrow, highly stretched basins underlain by strain softened lithosphere.

Sandiford (1999) also concludes that, in certain settings, significant long-term lithospheric weakening may accompany extensional basin formation. His modelling generates up to 5% weakening per kilometre of sedimentary basin infill where the prerift lower crust was initially strong (i.e., long time since last rifting episode), and where the basin fill is characterized by appreciable heat production (i.e., dominated by shale) and low thermal conductivity.

Of particular significance for understanding the mechanics of basin inversion, intra-lithosphere detachments are shown to be a natural consequence of the marked strength discontinuities that develop at major compositional boundaries such as the Moho during rapid extension. Such detachments serve chiefly to partition strain vertically such that the locus of deep lithospheric strain is physically offset from shallow strain by as much as several hundred kilometres (e.g., Castro, 1987).

Investigations of the consequences of depth-dependent strain models have tended to focus on extensional scenarios, where the implications of the so called simple shear model of extensional basin formation (Wernicke, 1981) remain controversial. Of particular interest in such settings is that depth-dependent strain reduces the buffering effect of deep lithospheric thinning on extensional basin subsidence. Potentially, therefore, synrift subsidence during depth-dependent stretching is rapid and deep. Hillis (1992) applies these ideas to the inverted basins of the western UK. By direct analogy, preferential thickening of the upper lithosphere during depth-dependent shortening may result in large and rapid uplift due to the displacement of the locus of deeper lithospheric thickening which, in uniformly shortened compressional settings, buffers surface uplift. However, despite the ability of models of lithospheric extension to account for the existence of intra-lithosphere detachments and weak zones, there is no indication that such detachments are ubiquitous beneath inverted basins.

The enigma of sedimentary basin inversion is emphasized by the modelling of four European inverted basins by van Wees and Beekman (2000). Their results show lithospheric strength increases of up to 20% during the extension and subsequent inversion of sedimentary basins. Yet, despite the fact that flanking structural highs are underlain by deeper, warmer and therefore weaker mantle lithosphere, it is in the central parts of basins where inversion is localized repeatedly. The key to understanding this paradox is the existence of long-term weak zones that deviate from standard rheological assumptions. Mechanisms of weakening of the upper mantle are presently unclear but may include shear zones (e.g., Beach, 1976), relict subduction zones along which lower crustal material was inserted into the mantle (Ziegler et al., 1995), and/or the occurrence of rheologically weak material (e.g., volatile-enriched zones: Menzies and Bodinier, 1993) as indicated by upper mantle seismic reflectors (e.g., Thomas and Deeks, 1994).

At upper crustal levels, weakening is attributable to pre-existing faults marked by reduced friction angles. In other words, Byerlee friction exerts the primary control over upper crustal deformation. In the model of Nielsen and Hansen (2000), localized fault reactivation triggers crustal thickening and the generation of short-wavelength topography. The topography of the inverted basin is compensated regionally by downward flexure of the relatively strong upper mantle resulting in the formation of marginal troughs synchronous with inversion. Subsequent cessation of inversion leads to isostatic uplift of the inverted basins and marginal troughs creating an exhumed inversion zone considerably wider than the initial rift basin.

To summarize, the precise mechanism by which stress is transferred the often considerable distances from collisional plate boundaries to synchronously inverting sedimentary basins is an outstanding problem. Nonetheless, the integrated approach to modelling sedimentary basin inversion described above, combining information from rheology, heatflow, strain rate and tectonic structure, is advancing our understanding considerably. The picture now emerging exemplified by the NW European foreland and its adjacent craton is that of a heterogeneous mosaic of mainly Mesozoic extensional basins with strongly contrasting rheological characteristics. Subjected to a fairly uniform, NW-oriented compressional stress field, the inversion is focused on discrete elements of this mosaic of former rift basins according to (1) the orientation and frictional properties of their basin-bounding fault systems and (2) their integrated lithospheric strength profiles.

In collisional mountain belts, synorogenic basin inversion has provided some of the most dramatic examples of inversion structures, often accessible to direct examination in the field. In such settings, inversion is promoted by:

1. high-magnitude horizontal compression,
2. low effective stress/high fluid overpressure due largely to the metamorphism and progressive dehydration of buried sedimentary rock (Fyfe et al., 1978) and
3. gravitational potential of the thrust wedge.

Thrust systems evolve in response to a negative feedback between the gravitational potential of the thickened wedge of shortened crust, promoting collapse of the thrust wedge, and the strength of the wedge and its basal detachment, serving to support the wedge. In the early stages of continental collision, a thrust wedge will build up to the maximum sustainable angle of taper by surface thrusting, duplex development and pure shear. Conversely, those segments of a thrust wedge exceeding their critical angle of taper are characterized by normal faulting and/or thin-skin topographic collapse.

Compared to the original numerical models of critically tapered wedges (Davis et al., 1983), which assumed uniform basal shear strength and an isotropic thrust wedge at Coulomb failure throughout, the situation in real orogenic forelands is considerably more complex. Specifically, orogenic forelands comprise previously stretched and rifted ‘passive’ continental margins and they are characteristically highly anisotropic. The Alpine forelands, for example, display considerable local variation in elastic properties (Sinclair, 1997), rift basin geometry and sedimentary fill, and mechanical weakness due to the presence of pre-existing fault zones. Nonetheless, the ideas of critical taper provide a ready explanation for the cyclic nature of sedimentary basin inversion in synorogenic settings, without the need to invoke significant changes in the regional stress field.

A good example of synorogenic sedimentary basin inversion is the Pyrenees, the crustal-scale structure and synorogenic evolution of which are today interpreted in terms of an overall inversion model. The Pyrenean thrust belt evolved in response to thin-skin shortening above a mainly Triassic basal detachment. However, recognition of a thick-skin Mesozoic normal fault system rooted beneath the main thrust belt (Camara and Klimowitz, 1985; Turner and Hancock, 1990; Turner, 1996), and information on the deep structure from the ECORS reflection seismic profile (Choukroune et al., 1988), reveals that most Pyrenean crustal thickening was accommodated by inversion of a Cretaceous rift basin system.

In the western Alps, Gillcrist et al. (1987) document a wide range of structural geometry accommodating the most efficient incorporation into Alpine shortening of deep (>5 km) Mesozoic half-graben containing highly thickened synrift successions. The footwalls to the half-graben comprise prominent blocks of crystalline basement rocks uplifted on low-angle Mesozoic normal faults reactivated during Alpine shortening (e.g., Pelvoux, Belledonne, Mont Blanc, Aiguilles Rouges massifs). The shallower, steeper segments of these faults generally acted as buttresses against which the relatively plastic synrift successions underwent intense upright folding, backthrusting and local cleavage development. Thus, the steeper segments of the Mesozoic normal faults experienced little or no reverse reactivation (cf. Fig. 9). Coward et al. (1991) show how the arrangement of the Alpine basement massifs exerted a major control over the pattern of Alpine thrusting which tends to flow around the buttressing effect of the Mesozoic normal faults.

In post-orogenic settings, where the contribution of continental collision to horizontal compression is relatively small, body forces originating from surface and/or subsurface topography can generate large deviatoric stresses sufficient to drive sedimentary basin inversion. Thus, Bott (1990) used finite element analysis to follow the consequences of thickening lithosphere made up of a 20-km-thick elastic upper crust above a viscoelastic layer replicating the rheological behaviour of the lower crust and mantle lithosphere. Where a large lower lithosphere root underlies the mountain belt, he showed that the compressive effect of the root is buffered by the tensile effect of thickening the upper crust, resulting in complex distribution of upper crustal deviatoric stress in the aftermath of orogenesis and lithospheric thickening. Whilst his models do not address sedimentary basin inversion directly, they are important for inversion studies because they demonstrate that, irrespective of far-field 'tectonic' stress, mass excess/deficiency in the lithosphere can result in large (up to 112 MPa) deviatoric compressional stress in the upper crust.

An example of the role of topography-induced gravitational stress in basin inversion is provided by Bada et al. (2001) from the Pannonian basin. Here, numerical stress models indicate that high levels of compressional stress (40–60 MPa) concentrated in the thinned Pannonian lithosphere originate from gravitational forces generated by the elevated topography of the surrounding Carpathian mountains. In particular, the kinematics of the inversion of the western Pannonian basin are consistent with topography-induced gravitational stress which locally exceeds the magnitude of the far-field stress.

Despite discouraging the use of "negative inversion" to describe normal displacement on formerly reverse-sense fault systems, it is undoubtedly an important process in orogenic and postorogenic settings. Syn- to post-orogenic extension has been documented in several regions such as the Basin and Range of the western US (e.g., Wernicke, 1981), the Aegean Sea (e.g., Gautier and Brun, 1994), the northern Apennine–Tyrrhenian Sea region (e.g., Carmignaani and Kligfield, 1990; Jolivet et al., 1998), the Massif Central (e.g., Malavielle, 1993) and the British and Norwegian Caledonides (e.g., McClay et al., 1986; Coward et al., 1989; Seranne and Seguret, 1987). In all these examples, the extent to which extant thrusts and reverse faults are reused during extension is variable. Probably, the clearest examples are those described from the British Caledonides where reflection seismic data image listric normal faults bounding Devonian half-graben sited above the traces of basement thrusts (e.g., Coward et al., 1989).

10. Conclusions

1. Sedimentary basin inversion is a global process that describes the compressional or transpressional reactivation and shortening of formerly extensional basins. It effects all the principal types of sedimentary basin.
2. The diagnostic criterion for recognizing sedimentary basin inversion is identification of the null point (null line in 3D), the point along reactivated faults at which net displacement changes from normal in the lower reaches of the fault, to reverse in its upper reaches.
3. Fluid overpressure is fundamental during basin inversion. As the primary mechanism for reduction of effective stress across existing faults and fractures, overpressure facilitates reactivation of faults which, theoretically, are unfavourably oriented with respect to maximum principal stress.
4. Sudden breaches of permeability seals during fault reactivation increments are responsible for transitory episodes of vertical fluid flow. This results in the transfer of potentially very large volumes of hydrothermal fluid to the relatively shallow subsurface, with attendant implications for petroleum reservoir geology and mineralization.
5. Measurement of inversion-related shortening strain in exhumed basins is problematic because (1) fault heaves in inverted basins are apparent and, therefore, not a good record of horizontal shortening and (2) strain partitioning during basin inversion means that significant shortening will be accommodated by deformation in the wall rocks of major faults. Classic inversion structures exemplified by recent seismic examples may record extreme inversion of more than 200% initial extensional strain.
6. Alongside epeirogeny, basin inversion is one of the two main geodynamic processes driving sedimentary basin exhumation.
7. Modern methods of thermal history reconstruction permit the cooling history and exhumation magnitude of deeply eroded basins to be quantified readily, but identifying the large-scale processes responsible for their exhumation is an outstanding problem. The problem is particularly acute in regions like the UK where epeirogeny and inversion overprint one another.
8. Despite the potentially detrimental effects of inversion, many prospective sedimentary basins have undergone severe inversion. Whether basin inversion enhances or suppresses petroleum prospectivity is not clear-cut and depends largely on the post-inversion hydrocarbon budget available through secondary and tertiary migration.
9. Pervasive brittle fracturing in the wall rocks of faults undergoing transpressional reactivation has the potential to enhance reservoir permeability considerably.
10. The idea that sedimentary basins are inherently weak confounds intuitive (and investigative) notions of the strain-hardening effect of lithospheric extension. The picture emerging from the NW European Alpine foreland is that of a mosaic of former rift basins with strongly contrasting rheological characteristics. Subjected to a fairly uniform compressional stress field, inversion was focused on discrete elements of this mosaic of basins according to (1) the orientation and frictional properties of their basin-bounding fault systems and (2) their integrated lithospheric strength profiles.

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