Chapter 12

The Structure and Hydrocarbon Traps of Sedimentary Basins

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12.1 Tectonic Regimes and Stress

Old knowledge told the early oil explorationist to focus drilling on structurally high-standing areas because hydrocarbons are trapped in positive structures like folds and rotated fault blocks that, as such, define positive structural features and associated topographic highs. Principally, however, three categories of traps are recognised, namely the structural trap, the stratigraphic trap and the hydrodynamic trap.

Structural geology is significant in all stages of the "value chain" of the exploration for and the exploitation of hydrocarbon resources, including the initial exploration screening phase, the targeted exploration, reservoir description and production (Gabrielsen & Møller-Holst 1997). This involves studies that include the basin-scale tectonic evolution, interaction between structural evolution and sedimentation, maturation, hydrocarbon migration, characteristics of trap type, sealing potential and leakage. But generally, the most common questions to ask a structural geologist in an oil exploration environment concerns the trap itself.

Many types of traps exist, some of which are determined purely by the sedimentary development, whereas the most common are due to structural development alone. Hence the purely stratigraphic trap is due solely to sedimentary processes, such as the primary pinch-out of sedimentary units. Others may be due partly to sedimentological and partly to structural developments (e.g. some types of inconformity traps), or purely structural (e.g. an anticline). To start with the purely structural trap, this depends on the tectonic environment (contraction, extension or strike-slip) in which it was created (see Sect. 12.1.2).

On a worldwide scale, the structural trap is by far the most common hydrocarbon trap type, but there is a tendency for the two other trap types to be more relevant in mature hydrocarbon provinces, because exploration moves into a more detailed approach in such areas. Structural traps are generated by tectonic, diapiric, compactional or gravitational processes. They must contain a reservoir constrained in three dimensions, a cap, and be in communication with a reservoir unit. The structure has a spill-point defining the volume available in the trap, and commonly develops a gas/oil contact at the top and oil/water contact at its base.

Geometrically speaking, the simplest structural trap may be the anticline, or modifications hereof (Fig. 12.1a). Compressive tectonic regimes commonly include a host of contractional folds and thrusts. Good examples of such areas that are also hydrocarbon provinces are the Zagros Mountains of Iran, the Caribbean of South America and the North American Cordillera. To be of interest for the entrapment of hydrocarbons, such structures must have a closure as seen along the fold axis, implying that they should not be strictly cylindrical. This is often the case for natural, contractional folds, particularly in situations where deformation took place above an uneven basement topography and particularly so in areas where double-folding (folding with to sets of cross-cutting fold-axes) and halokinesis have occurred (Fig. 12.1d). The anticline must be thoroughly mapped in three dimensions.

The complexity and communication between contractional structures are, to a great extent, dependent on

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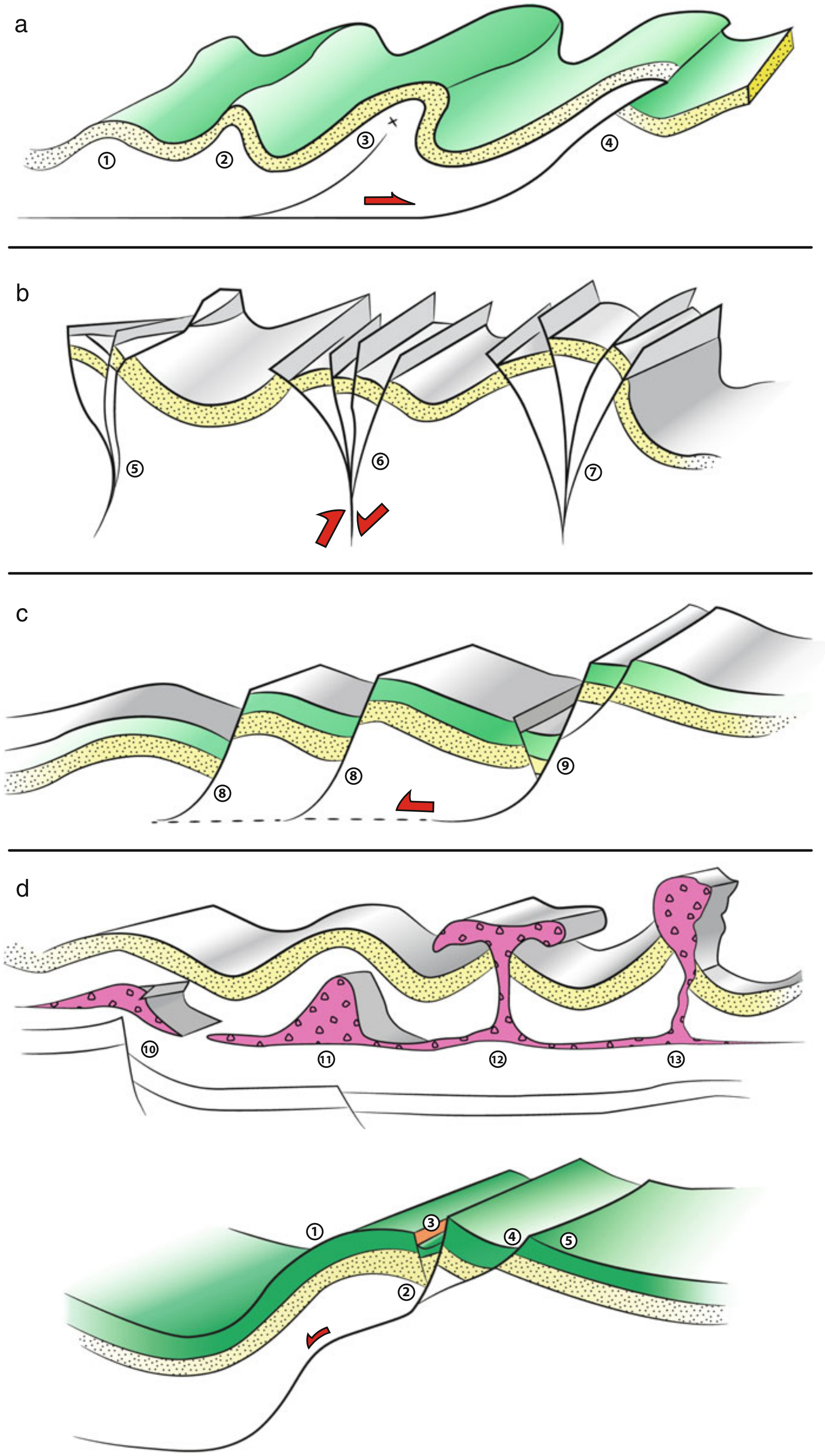

Fig. 12.1 Most common structural trap types. (**a**) Trap types (folds and thrusts) associated with contraction. There can be a close genetic relation between contractional structural traps so that one type (e.g. a symmetrical fold) may be an early stage of

the deformation. Thus, a mildly deformed province may contain sets of symmetrical buckle-folds (Figs. 12.1a and 12.16b), whereas a province of advanced contractional deformation may be dominated by systems of complex, overturned fault-propagation folds and thrusts (Figs. 12.1a and 12.15).

Complex folds are also common in strike-slip regimes, developed either as fold-trains formed orthogonally to the contractional axis of the strike-slip system (Figs. 12.1c and 12.14) in areas of constraining bends (Fig. 12.14c), or in connection with transpressional segments of the strike-slip system. Here, folds may grade into flower structures. (Figs. 12.1a and 12.14c).

Many petroleum provinces occur in extensional basin settings, where the rotated fault block and affiliated accommodation structures provide the most common trap type (Figs. 12.1b, 12.8, 12.10 and 12.11). For example some of the major hydrocarbon fields in the Norwegian continental shelf, like the Statfjord and the Oseberg fields, occur in such settings. The rotated fault block itself becomes tilted according to the amount of extension and the geometry of the principal fault plane (planar or listric). Secondary traps may occur due to elastic uplift of the footwall, roll-overs in the hangingwall due to a strongly listric fault plane, or forced folds associated with ramps in the detachment (Fig. 12.1e).

Halokinesis (the processes associated with the salt movements) occurs because salt has a specific gravity of 2.2 g/cm^3; as opposed to fully consolidated (low porosity) sedimentary rocks that have specific gravities of 2.5–2.7 g/cm^3), creating a buoyancy effect. This is further supported by the fact that on a geological timescale, salt flows plastically. A number of pronounced structural types result, from pillows to diapirs of salt. Strong structuring may also be associated with mud diapirism. In a classical study of mild to extensive salt movements, Halbouty (1979) identified no less than nine different structural trap types (Figs. 12.1 and 12.17). The general experience is that larger structures are affiliated with early-stage salt structures, like pillows, whereas a much more complex structural pattern and dismembering of the structural traps are typical for the more advanced stages of deformation, including diapirism.

In the following sections, trap types and risks associated with the different trap types are investigated.

12.1.1 Principal Stress Regimes and Types of Stress

The concept of plate tectonics offers a useful framework for structural geological analysis on all relevant scales in petroleum geology, from regional in the exploration stage, to local in the reservoir evaluation and production stages. This is natural, because the principal geological stress systems are ruled by processes in the deep Earth like mantle convection and lithosphere subduction, the secondary effects of which are manifested at the base of the lithosphere and along plate margins. Based on these concepts, the basic dynamics of the lithosphere can be quantified, which is a prerequisite for the evaluation and calculation of the state of stress at any point. As seen in the perspective of the petroleum structural geologist, understanding and quantifying the stress situation at the plate margins is a prerequisite for understanding the state of stress in any basin system and in any reservoir.

We term principal stresses originated at plate margins *far-field* or *contemporary* stresses. The stress situation in a basin or a reservoir may be a sum of several far-field stresses combined with a *local* stress, which may be related to burial, erosion, geothermal gradients, topography, basement relief and structural inhomogeneities in the substratum. In other words, the plate tectonic framework provides a basic and general concept on which any structural geological analysis of a sedimentary basin rests, but it must be supplied with information on the local stress system that is superimposed on it. In the context of the far-field plate tectonic stress, we distinguish between the *plate boundary* and the *intra-plate* component.

Fig. 12.1 (continued) the development of fault propagation fold and a fully developed thrust. See Figs. 12.2 (for structural) and 12.15 (for stratigraphic) traps associated with contractional or inverted terranes. (**b**) Typical structural trap type associated with strike-slip (positive and negative flower structures and associated folds. See Fig. 12.13 for possible stratigraphic traps in strike-slip regimes. (**c**) Extensional, rotated fault-blocks above a decollement (floor fault). See Figs. 12.11 and 12.12 for structural, stratigraphic and unconformity traps in such systems. (**d**) Different types of salt structures and associated trap types. (See also Fig. 12.16)

The plate boundary stress is subdivided into three basic plate margin settings, namely *constructive boundaries* where adjoining plates are moving away from each other and new crust is formed by magmatic activity, *destructive boundaries* where lithosphere is consumed by subduction or obduction, and *conservative boundaries* where plates are moving past each other in a strike-slip sense and where lithosphere is neither created nor consumed (Fig. 12.2a). It is important to note that plate boundaries have been generated and destroyed throughout large parts of the history of our planet, so that many may be preserved inside the present plates and hence do not coincide with present continent/ocean margins. The three basic types of plate margin coincide with the three principal stress configurations, namely the tensional, the compressional and the strike-slip regimes. To describe these regimes, we rely on the concept of principal stress configurations and their definition in the context of the principal axes of stress.

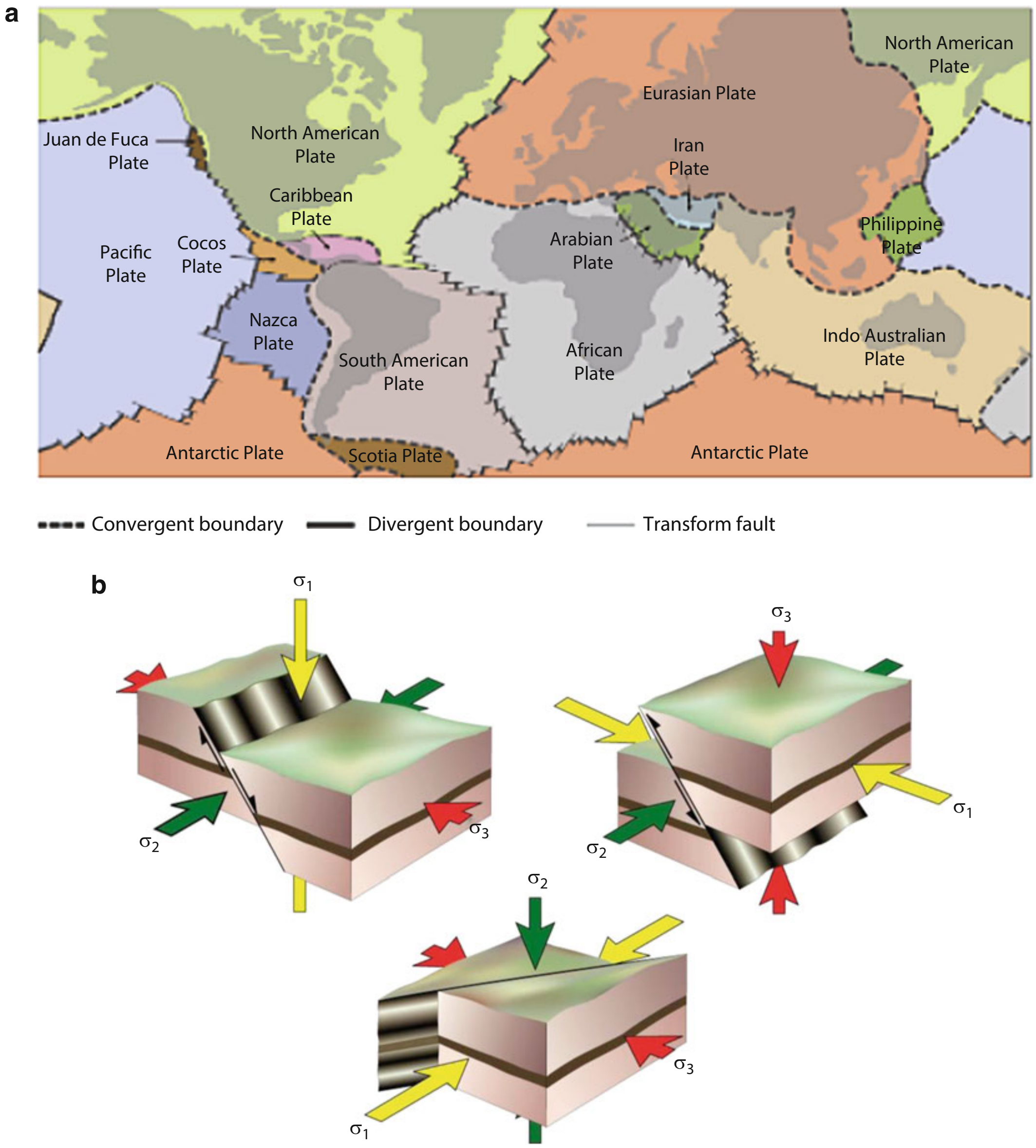

Fig. 12.2 (**a**) The major tectonic plates and their boundaries (from Nystuen in Ramberg et al. 2008). Reprinted with permission from the Norwegian Geological Society and the author. (**b**) Principal stress configurations (from Fossen and Gabrielsen 2005)

12.1.2 Tectonic Stress in the Earth's Lithosphere

As we have seen already, in addition to stresses generated at tectonic plate contacts, several other processes contribute to the generation of stresses. We may, for example, have *residual stress*, which is inherited from previous deformation, where a rock body may have become bent with the elastic stress component remaining unreleased, and *thermal stress* which is related to expansion of rocks during heating or contraction during cooling and stress related to local gravitational gradients. Accordingly, the *total stress* consists of several components:

$$\text{Total stress} = \text{reference stress} + \text{residual stress} + \text{thermal stress} + \text{tectonic stress}$$

where the reference stress refers to the stress inside the plate, devoid of plate tectonic stresses, and thus the

$$\text{Tectonic stress} = \text{contemporary stress} + \text{local stress}.$$

We will not go further into the analysis of residual and thermal stress here, but concentrate on stress generated by primary interactions at plate margins or forces derived thereof.

Taking into consideration the reference stress conditions, one can describe the stress situations for the three principal conditions of deformation, namely extension, contraction and strike-slip. This was done by Anderson in two influential works in 1934 and 1951, in which the framework for all modern tectonic and structural geological analysis is defined. Anderson used the vertical lithostatic stress (σ_v) as a reference:

$$\sigma_v = \rho g z \tag{12.1}$$

(where ρ is the specific weight, g is the constant of gravity and z is the height of the rock column). This applies for the three principal stress conditions because σ_v can be considered similar for one rock type for a constant h, assuming that the burial history has been similar. Thereby the three principal stress systems can be defined, each corresponding to a regime of deformation (Fig. 12.2a, b):

$$\sigma_v > \sigma_H > \sigma_h;\ \text{extension}, \tag{12.2}$$

$$\sigma_H > \sigma_v > \sigma_h;\ \text{strike-slip}, \tag{12.3}$$

$$\sigma_H > \sigma_h > \sigma_v;\ \text{contraction}, \tag{12.4}$$

where σ_H and σ_h are the maximum and minimum horizontal stresses, respectively. By using σ_v as a reference, further calculations become dependent on the reference system applied. In the contractional regime there will be a tectonic component $\left(\sigma_t^*\right)$ in addition to the reference stress, so that the greatest (horizontal) stress is:

$$\sigma_H = \rho g z + \sigma_t^* \tag{12.5}$$

Assuming uniaxial stress, which implies that the reference stress is a function of the elastic properties of the rock, and ignoring the thermal expansion, we can describe the stress as:

$$\sigma_H = \left[\frac{\upsilon}{1-\upsilon}\right]\rho g z + \sigma_t \tag{12.6}$$

where υ is Young's modulus (Chap. 11). Because $(\upsilon/1-\upsilon) < 1$, $\sigma_t > \sigma_t^*$. Accordingly, the reference stress condition selected also influences the calculated total tectonic stress as long as the buried rock is compressive. For a non-compressive rock ($\upsilon = 0.5$), lithostatic and uniaxial reference systems are equal.

12.1.3 Deformation Mechanisms and Analogue Models

Our daily contact with the physical world tells us that materials deform in many ways, depending on the applied stress and type of material and its physical state. Thus, a liquid reacts to outer stress very differently from a piece of rock, and one type of rock like chalk has very different physical properties as compared to another like granite. Furthermore, one material may change its mechanical properties dramatically by change of temperature and pressure. These contrasts are founded on processes occurring on the scale of the grains (of the rock) and on the molecular and atomic scales. We have given these processes and their associated meso- and macroscopic physical expressions names like brittle, elastic, plastic, viscous and ductile, and sometimes combinations like elastico-plastic.

Unfortunately, these terms are not always used in a consequent manner and are therefore liable to cause confusion when taken out of context or not precisely defined. This particularly concerns the term “brittle”, because it is used in a double sense, namely as a deformation mechanism and a deformation style.

When applied in the context of *deformation mechanisms*, *brittle deformation* implies that existing bonds are physically broken between mineral grains, or that fracturing of the individual grains themselves takes place. As a consequence, the rock loses its cohesion and (potentially) physically falls apart. In contrast, the *plastic deformation* mechanism implies that deformation takes place by the transfer of dislocations on the atomic scale. This means that the mineral can change its shape without loss of cohesion. Generally, plastic deformation occurs at higher p,T-conditions than those accompanying brittle deformation.

Concerning *deformation style*, the term *brittle* is used about localised strain, like that associated with jointing and faulting, and particularly in cases when the rock loses its cohesion and where the deformation occurs at lower p,T-conditions (though not necessarily so). The *ductile* deformation style characterises strain also by low stress, which is homogeneously distributed over a wider area, as commonly observed in connection with folding and meso- and mega-scale shear-zones. A ductile style of deformation is predominant at high p,T-conditions, but may also occur under very low p,T-conditions, if it involves weak materials like sand and clay. In such cases, however, displacement takes place along grain boundaries or along borders between rock bodies and not by dislocation creep or other atomic-scale mechanisms that characterise the plastic deformation mechanism.

12.2 Petroleum Systems in Extensional Regimes

Areas of extension are affiliated with horizontal divergent stress and are found in association with constructive or passive plate boundaries and in intra-plate settings. Thus, extensional stress regimes either are associated with subsidence and basin formation (in intra-plate settings) or characterise active break-up of continents (along constructive or passive margins).

Although the conditions for development of petroleum systems in areas of active spreading may be meagre, the remnants of the earlier stages of break-up, now situated in passive margins settings, fulfil all the requirements that characterise productive petroleum provinces. This is because such tectonic regimes have undergone crustal thinning and associated subsidence, which involves all the processes essential for petroleum to be generated, trapped and accumulated in sufficient volumes and concentrations for petroleum fields to be commercially interesting. Accordingly, such settings frequently display an attractive combination and distribution of source, reservoir and cap rocks, structural and stratigraphic traps and the conditions for maturation, expulsion, migration and accumulation of hydrocarbons.

12.2.1 Extensional Basins

The formation of extensional basins may be seen as the first stage of the *Wilson Cycle*, which begins with thinning, stretching and rifting of the continental crust followed by continental break-up and mid-oceanic spreading. The concept of the Wilson cycle predicts that this process becomes reversed, causing closure of the ocean, collision between the adjacent continental plates, and hence the construction of a mountain chain along the zone of collision (Fig. 12.3). The junction between the continental plates defines the *suture* between the two.

If we use the present North Atlantic as one example, the highly hydrocarbon-rich northern North Sea basin system is situated in a passive continental margin configuration, where the extensional basin system developed during continental break-up. In contrast, Iceland, where petroleum resources are less abundant, is situated on the top of the mid-oceanic spreading ridge. However, if one looks more closely at the structural configuration at depth, one finds that the northern North Sea basin system, which includes the Viking Graben that developed in Jurassic-Cretaceous times, is

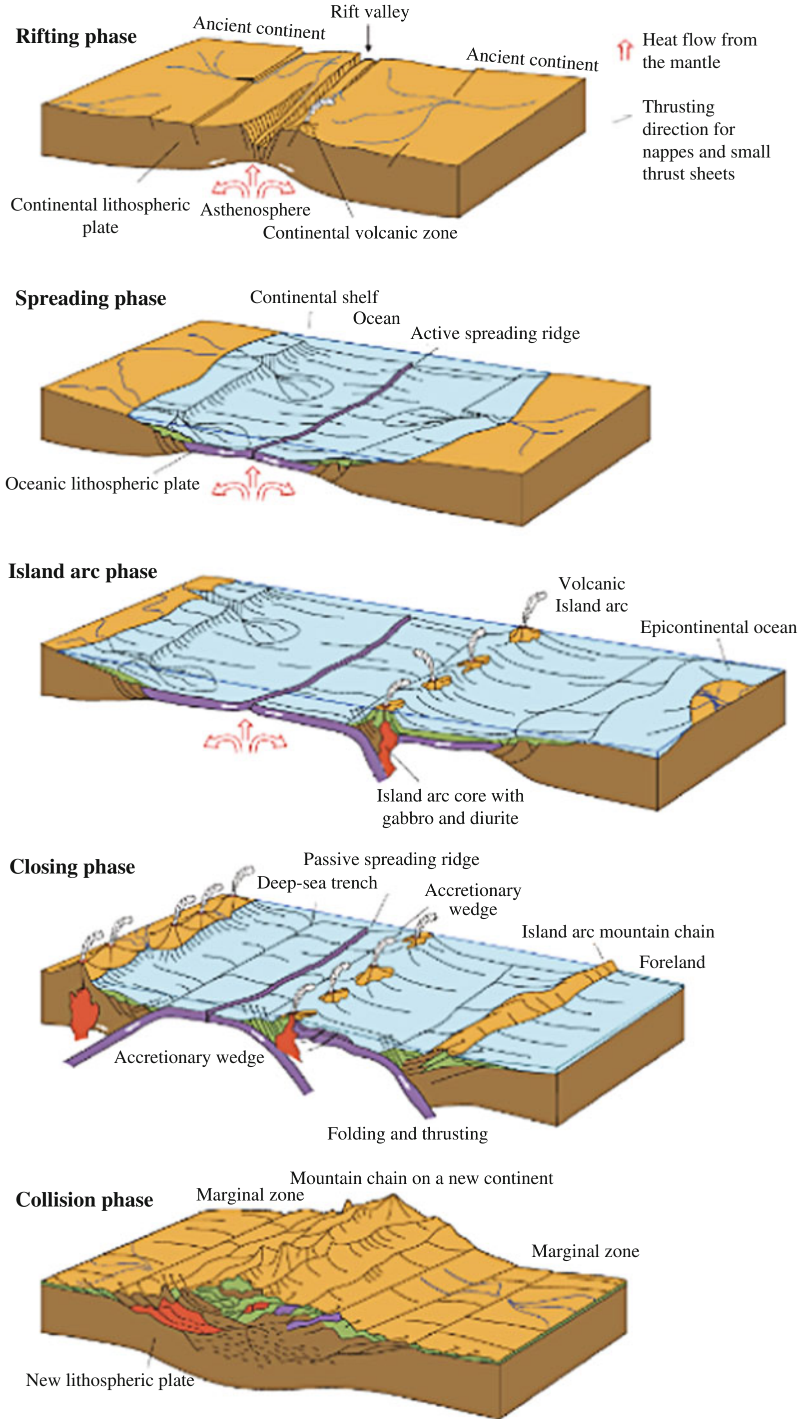

Fig. 12.3 The major stages in the Wilson Cycle (from Nystuen in Ramberg et al. 2008). Reprinted with permission from the Norwegian Geological Society and the author

underlain by an older (Permo-Triassic) basin system (Fig. 12.4). The Permo-Triassic basin system is in turn superimposed on the even older Caledonian orogeny, which was subsequently affected by gravitational collapse in Devonian times, representing the last stage in a previous Wilson Cycle.

There are several models for the lithospheric configurations that accompany extensional crustal thinning, the end members of which are the "pure shear" (symmetrical) and "simple shear" (asymmetrical) models (Fig. 12.5). It should be noticed that these models are not necessarily mutually exclusive; we can find basin systems that display elements from more than one model, such as the "delamination model" (Fig. 12.6).

The "pure-shear model" for extensional crustal thinning was suggested by Dan McKenzie in 1978, in a paper that has become the most frequently cited in geosciences in modern times. This model assumes thinning of the weak lower crust/lower lithosphere by pure shear, and hence is characterised by the development of a symmetrical configuration (Fig. 12.5a). The pure-shear extension of the ductile lower crust is accompanied by thinning of the upper crust by brittle faulting and subsequent development and rotation of fault blocks.

In this context it is possible to separate the active stretching stage, which is associated with fault-controlled thinning of the upper crust, and later subsidence controlled by thermal processes. As a response to extension, the crust and upper mantle lithosphere becomes thinned and, promoted by extensional faulting, the basin floor will subside quickly. This implies that deeply seated warm rocks are transferred upwards in the lithosphere so that the isotherms in the thinned area become elevated and the thermal

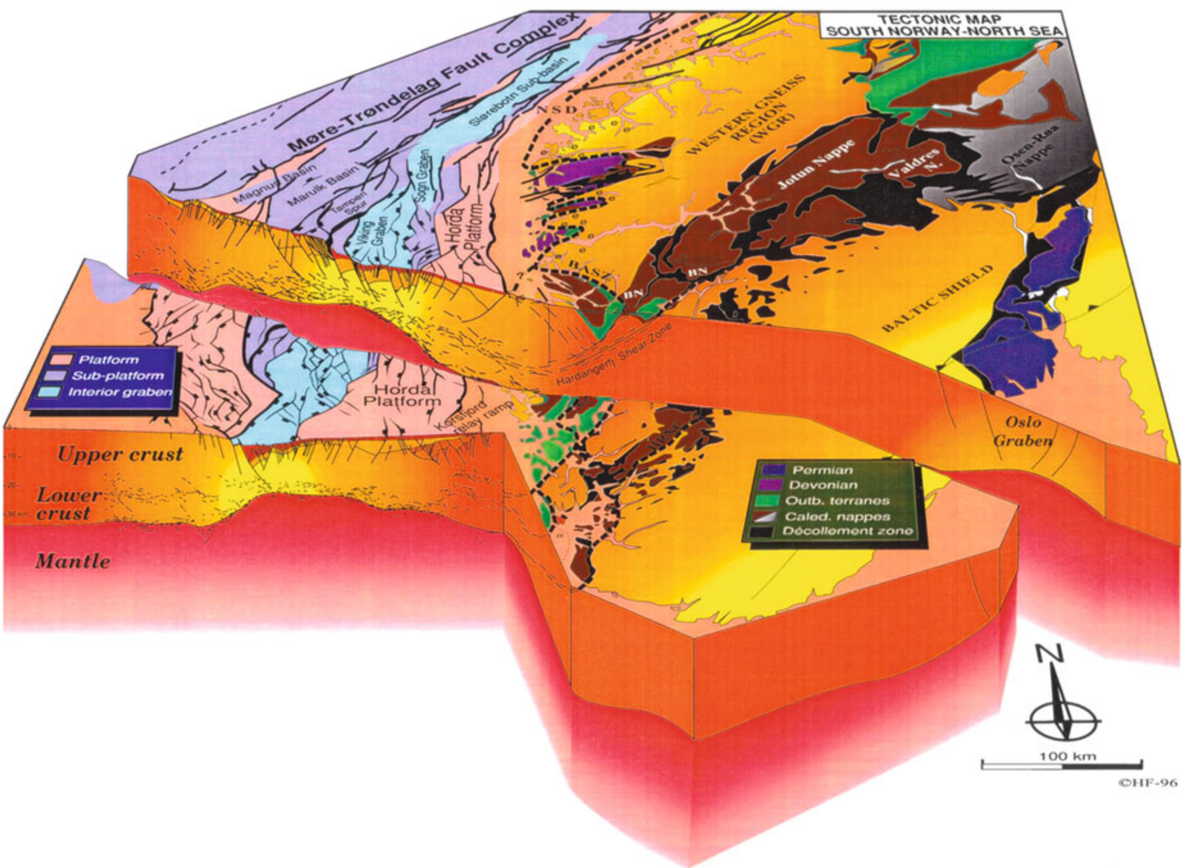

Fig. 12.4 The architecture of the North Sea continental shelf. Note the deep structure, showing a Permo-Triassic rift system buried beneath the younger Jurassic – Cretaceous Viking Graben (from Fossen 2002)

gradients become steepened accordingly. These deep processes influence the relief of the basin floor because heating causes rock volumes to expand and elastic, quasi-plastic and isostatic adjustments to occur simultaneously at lithospheric, basin (e.g. by uplift of the basin margins) and fault block scales. In the next stage of development (the post-rift stage), the basin will continue to subside due to a combination of thermal contraction, sediment compaction and sediment loading. This sounds complex, but luckily these processes are well understood and can be modelled with good accuracy on the basis of the algorithms proposed by McKenzie and supplied with additional modelling tools, developed particularly in the late 1990s.

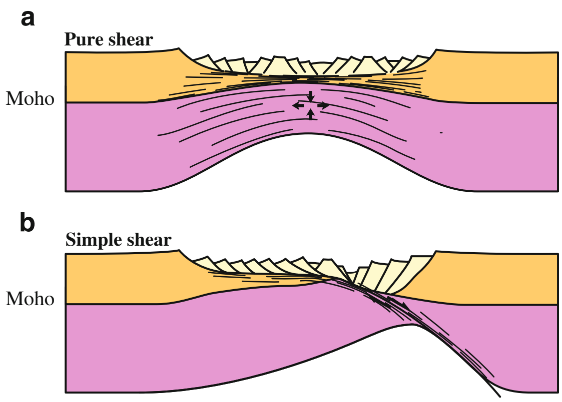

Fig. 12.5 Basic configuration of (**a**) pure shear and (**b**) simple shear extensional basins (modified from Fossen and Gabrielsen 2005)

For modelling purposes and for the analysis of extensional basins with respect to petroleum exploration, three stages of development can be distinguished (Fig. 12.7).

The pre-rift stage is characterised by gentle flexuring and fracturing of the lithosphere. In some rifts we see the development of a gentle bulge, caused by mantle doming and associated warming – and hence expansion – of the lithosphere. In other cases, a gentle subsidence, defining a broad, shallow basin is seen, caused by mild extension of the cold (not-yet-heated) lithosphere. In both cases, the lithosphere is prone to develop steep fractures on a crustal or even lithospheric scale. These fractures have the capacity to accommodate magma, generating dikes. Regarding hydrocarbon reservoir potential characterising the pre-rift stage, sand deposits are likely to be sheet-like and relatively thin, with few structural traps developing at this stage. Sediment transport is mainly transverse to the basin axis, but quite

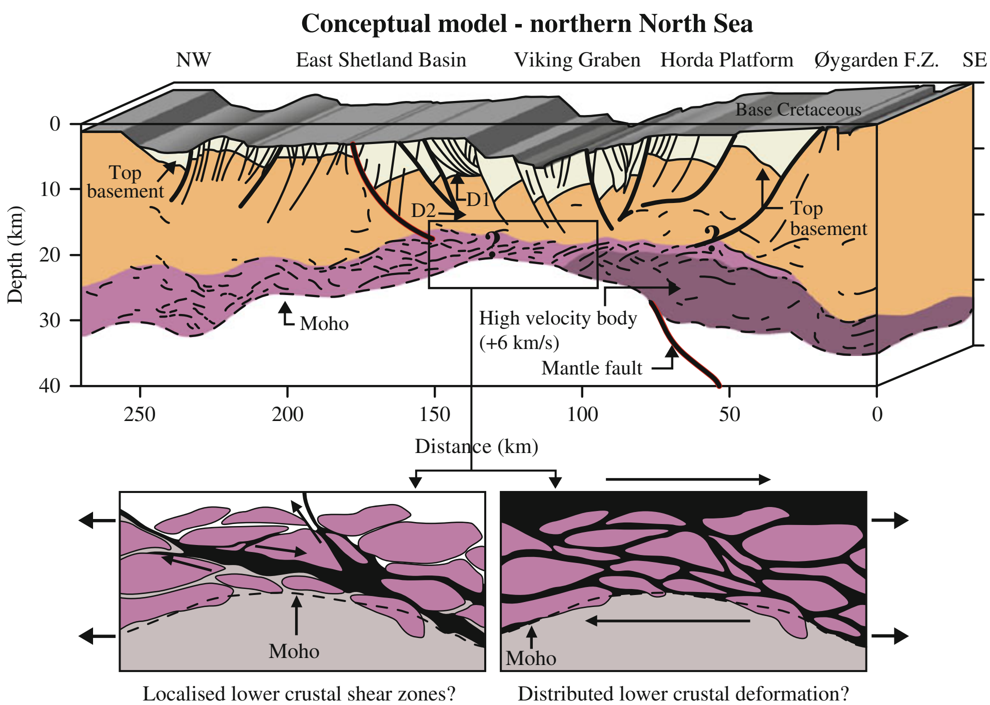

Fig. 12.6 Model of the Viking Graben, displaying elements of pure and simple shear. Modified after Odinsen et al. (2000)

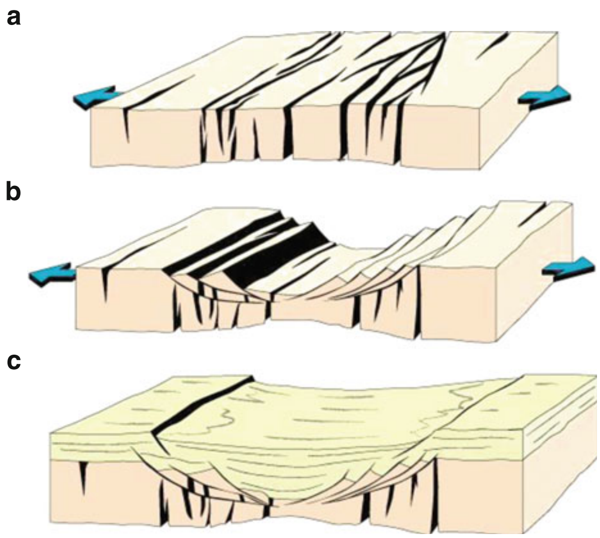

Fig. 12.7 Three major stages in the devlopment of extensional basins. After Gabrielsen (1986)

homogeneous due to lack of pronounced gradients in the basin. The marginal sediment transport system is prone to act in concert with the axial transport system, feeding the latter with sediments. This may consist of braided or meandering river systems, depending on factors like axial basin gradient and climate. Since most rifts are generated by break-up of continents, a terrestrial depositional environment would be most common for the pre-rift stage, so that source rocks and cap rocks, which are mostly of marine depositional origin, may be scarce (Fig. 12.8a). There are, however, numerous examples of both source rocks and cap rocks of terrestrial origin.

In *the active stretching stage* extension, and hence also subsidence, accelerate. Simultaneously, heat input increases due to upheaval of hot layers of the mantle lithosphere. The steep fractures generated in the pre-rift stage will not be able to accommodate the extension and a new set of low-angle planar or listric faults will be activated, separating fault blocks that are detached from the lower crust by a subhorizontal zone of weakness. Gliding on the system of detachments, the fault blocks and their internal beds will rotate away from the basin axis (Figs. 12.7b and 12.9a). From the view of the petroleum explorationist, the active stretching stage deserves particular attention because of the variety of structural and stratigraphic traps that may develop. This stage is also characterised by a complex sediment distribution system that is likely to produce a variety of lithofacies due to the increasing topographic relief associated with high fault activity. The marine transgression that commonly follows the increased subsidence of the basin floor also contributes to this variety in sedimentary facies. Sand that is eroded from the high-standing parts of the basin (e.g. basin shoulders and crests of rotated fault blocks) may be trapped in lows in various structural positions and these units are likely later to be covered by transgressive marine sediment accumulations. The sediment transport system in the active stretching stage is likely to be dominated by complex transverse and locally bi-directional fluvial systems that are strongly influenced by the elongated, rotated fault blocks, generating axis-parallel transport in segments along the basin margin. The central part of the basin may be less complex and axial-parallel sediment transport would prevail there (Fig. 12.8b).

In *the thermal subsidence stage*, thermal contraction of the lithosphere dominates the basin subsidence pattern. Because solids typically contract during cooling, the parts of the basin that have experienced the strongest extension (i.e. those that have been thinned the most and hence heated the most) will contract and subside more than other parts. In a pure-shear configuration this is most likely to be the central segment running along the basin axis. This means that the rotation of strata upwards away from the basin axis becomes reversed so that strata begin to rotate downwards towards the basin axis (Fig. 12.9b). This rotation is strengthened by sediment loading and compaction (thickest sequence in the central part of the basin).

The transverse sediment transport will persist during the thermal subsidence stage, while the basin floor becomes gradually smoothed. An axial transport system may also still be active, but is likely to become less pronounced through this stage of development (Fig. 12.8c). Depending on the balance between subsidence and sediment input, the water depth will vary from one basin to another, but the depositional environment is likely to be marine and the central part of the basin may attain great water depth (thousands of metres). The fault systems that dominated the basin floor geometry during the active stretching stage are now quiescent, and stratigraphic hydrocarbon traps rather than structural ones are likely to be the most common.

Syn-rift to post-rift transition. The *pure-shear model* predicts that a simple geometrical change of

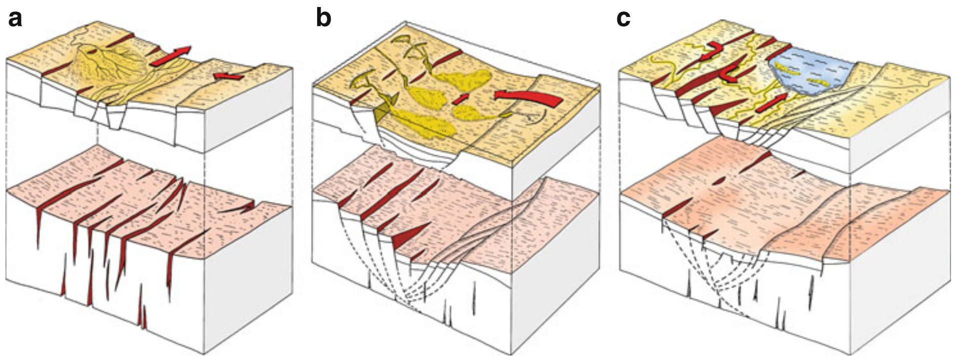

Fig. 12.8 Principal sedimentary transport systems associated with the three mains stages in the development of extensional basins. After Gabrielsen et al. (1995)

the outline of extensional basins will accompany the transition from the syn- to the post-rift stage. In this model the margins of the relatively narrow, steep-walled rift, which traps the syn-rift sediments, become overstepped at the syn- to post-rift transition. This implies that the basin becomes wider and the rate of subsidence decreases asymptotically during the following post-rift stage. Thus, one defines the beginning of the post-rift development as the stage by which the syn-rift faults become inactive and subsidence becomes controlled dominantly by thermal contraction and sediment loading.

In practical terms, the identification of this stage in the basin development is not trivial, because the transition is frequently not synchronous all over the basin, and the criteria for identifying the transition in reflection seismic data are not always well constrained. To overcome this problem, the syn- to post-rift transition should be defined more precisely as the point in time when net heat out of the system is greater than net heat into the system. It is recognised that a lateral heat flow gradient commonly exists perpendicular to the basin axis. This implies that the area closest to the basin axis, which coincides with the area of greatest thinning, is also the part of the basin displaying the highest heat flux at the end of the syn-rift stage. The lithosphere beneath the central part of the basin will accordingly undergo the greatest vertical contraction during the post-rift stage. The enhanced subsidence at the basin axis is further enhanced in cases where the basin is filled by sediments, creating an extra load and also a greater total compaction. Hence, the syn- to post-rift transition coincides with a regional shift in tilt from fault block rotation *away* from the graben axis during the syn-rift stage to tilting directed *towards* the basin axis during the post-rift development (Fig. 12.9). This change is due to a shift from bulk thermal expansion to bulk thermal contraction of the lithosphere and is in most cases clearly distinguishable in reflection seismic data.

It needs to be emphasised that the syn- to post-rift transition is unlikely to occur simultaneously throughout the entire basin. This is due to differences in structural configurations, e.g. the existence of *graben units*, and thermal inhomogeneities associated with variable stretching both along and transverse to the basin axis. For reasons discussed below (Chap. 22), the entire Cretaceous sequence of the northern North Sea is included in the post-rift development *sensu stricto*. Furthermore, analysis of the basin topography permits three sub-stages to be identified within the framework of the post-rift development: the incipient, the middle and the mature post-rift stages. The configuration at the syn-rift/post-rift transition is treated separately in the present analysis (Section 12.3.1).

In the analysis of basin subsidence it is important to remember that in addition to the effects of fault-related subsidence and thermal expansion and contraction, the basin's subsidence is affected by elastic deformation and isostasy, and in many cases also by extra-basinal stress.

The simple-shear model for extensional basins is in considerable geometrical and mechanical contrast to the pure-shear model for extensional basins in that the simple-shear model assumes that extension is concentrated along one or several inclined fault zone (s) affecting the entire crust (Fig. 12.5b). Still, where thermo-tectonic and isostatic responses are concerned

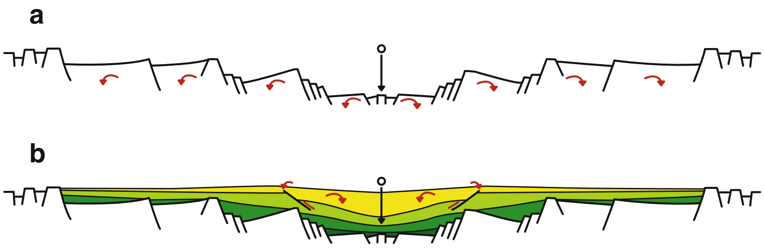

Fig. 12.9 Pattern of rotation of sedimentary units in the (**a**) syn- and (**b**) post-rift stages. After Gabrielsen et al. (1995)

the principles are similar to those of the pure-shear model. The simple-shear model is based on observations in the Basin-and-Range of North America and was formulated by Brian Wernicke in 1981. The Basin-and-Range basin system displays a particular geometry in that the lithosphere is extended to the degree that the lower crust, described as a metamorphic core complex, has become uplifted and exposed in the central part of the basin. The asymmetrical configuration of the basin particularly influences the pattern of isostatic response to extension. An important factor is the relative thickness of the upper mantle/lithosphere. This is because the lower crust commonly is denser than the upper astenosphere, causing large-scale contrasts in differential subsidence and uplift across the basin. Superimposed on this are more local isostatic effects, associated with contrasting thicknesses of layers with different densities and the topography of the basin.

Since the same tectono-thermal principles that apply for the pure-shear basin also are valid for simple-shear basins, the main basin stages and the conditions for hydrocarbon generation and entrapment are also the same. Even though the simple-shear model was inspired by analysis of the Basin-and-Range basin system it has proved relevant for many other basins too, suggesting that simple shear is a common component in the formation of basins.

The delamination model can be seen as a combination of the simple- and pure-shear models. In this case the upper and middle crust extends by simple shear. At depth, the master fault flattens and merges with the lower crust, which becomes thinned by pure shear.

The Viking Graben of the northern North Sea seems to have a configuration that fits the delamination model (Fig. 12.6). Also in this case, the thermo-mechanical pure-shear model applies and, with some modifications, can be used to model the basin development. However, the delamination model makes it necessary to take into account an additional variable parameter, namely that the two parts of the lithosphere situated above and beneath the delamination surface have undergone different amounts of extension.

12.3 The Structural Architecture of Extensional Basins

Comparison of many extensional basins reveals that they have many architectural elements in common. These include the position and geometry of the dominant fault systems, the position of the most prominent terraces or platforms, and the position of structural highs. This does not imply that all basins are similar or that all contain the same types of structures. Nevertheless, a systematisation suggested in the following, is useful in the analysis of extensional basins and their exploration for hydrocarbons. Before looking in detail at the exploration leads that are typical for extensional basins, it is therefore instructive to examine the principal structural building blocks of the extensional basin (Fig. 12.10a, b).

The *extra-marginal fault complex*. The distal part of the basin is separated from the foreland by an enhanced fault frequency (as compared to the foreland of the basin) and a set of planar to listric faults commonly arranged in an *en echelon* geometry and sometimes generating a horst-and-graben-system with a moderate relief. In some cases dike systems affiliated with the initial stage of extension are filling in some of the faults of the extra-marginal fault complex. In many cases, a horst is seen to separate the extra-marginal fault complex from the platform, defining a topographic threshold between the foreland and the basin itself.

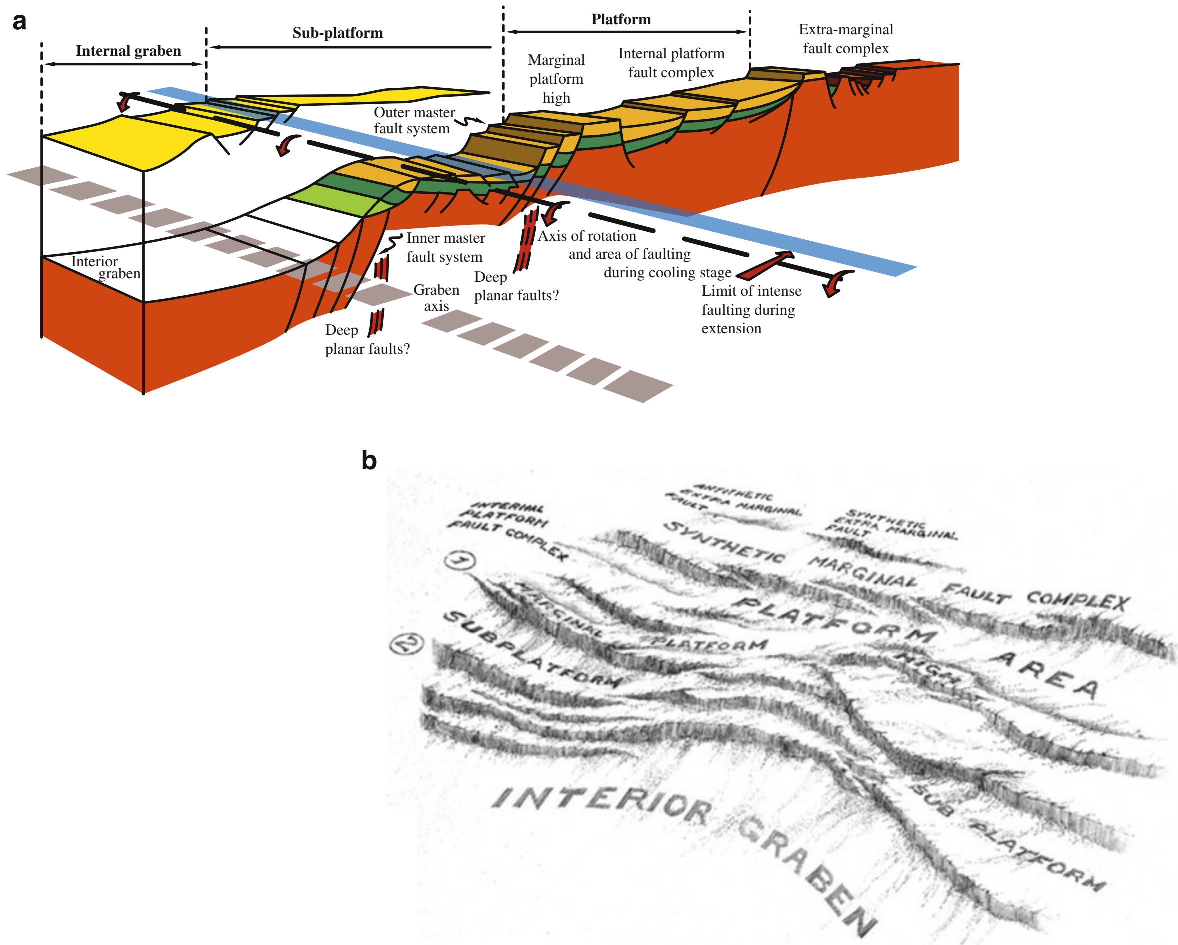

Fig. 12.10 (**a**) Principal sketch of major structural elements in graben systems. (**b**) Graben margin with its main structural elements as seen in three dimensions. Note the change of configuration as seen along the strike of the margin. Redrawn from Gabrielsen (1986)

The *platform* is a relatively flat tectonic unit that is less intensely faulted. The extension may be concentrated on a limited number of faults, but even those display only moderate throws as compared to the major fault complexes that delineate the interior basin. Hence, the platform is characterised by a few broad, and only slightly tilted, structural traps. Where age determinations are possible, the fault systems of the platform area are seen to have been activated at an early stage of the basin development, but the activity slowed down or became arrested when the subsidence accelerated in the inner part of the basin system. The platform may also be delineated on its basinward side by a horst.

A *platform marginal horst* is sometimes developed at the basinward side of the platform, defining the transition from the tectonically quiet platform to the much more heavily faulted sub-platform or the inner marginal fault complex. The platform marginal horst is asymmetrical in outline in that its platform-facing margin is defined by a few faults with moderate throws, whereas its basinward border consists of a complex system of faults with great throws, sometimes in the order of several 100 m. The fact that the marginal platform horst is such a common feature in many mature extensional basins suggests that it has an important mechanical significance, the position of its distal (antithetic) fault perhaps being determined by the mechanical strength of the platform rocks.

The *sub-platform* is delineated by the marginal platform high on its distal side and the interior graben on the other. The *inner marginal fault system* that separates the subplatform from the interior graben, together with the extra-marginal fault complex, are the most profound fault zones of the basin, and the two are likely to be linked along the principal detachment found within the lower crust. The subplatform is heavily faulted and encompasses a number of secondary rotated fault blocks that again may be criss-crossed by a third order of faults. There are examples that the second order faults flatten along local detachments at shallower levels than both the primary master faults of the marginal fault complex and the inner and outer marginal fault systems. The *inner marginal fault system* also coincides with the axis of basinward rotation as activated during the post-rift stage.

The *interior basin* is the unit of the basin which is underlain by the most extensively thinned crust and where the maximum post-rift subsidence occurs. In the case of a symmetrical (pure-shear) basin, it is delineated on both margins by inner margin fault systems, whereas in the case of an asymmetrical basin (simple shear) only one of the margins has this status. Even the deepest part of the basin is underlain by rotated fault blocks, initiated during the active stretching stage of the basin formation.

It should be noted that the structural elements described above are not likely to be present along the entire basin margin. Thus, in some segments, the platform may be in direct contact with the interior basin, whereas in other segments the platform or the platform marginal high may be missing. This inconsistency may reflect the influence of structural or lithological inhomogeneities in the basement, varying strain rates or uneven bulk extension along the basin axis. Indeed, it is common for large rifts that the basin is divided into several sub-basins or basin units, each distinguished by its particular geometry and even polarity.

12.3.1 The Structural Influence on Reservoir and Source Rock Distribution in Extensional Basins

The types of traps related to the different stages in graben formation are illustrated in Fig. 12.11. The numbered trap types mentioned in the following sections refer to this figure.

12.3.1.1 The Initial Stage

According to the model presented above, sedimentation during the initial stage will occur in a broad, shallow basin with moderate surface gradients. Due to the moderate stretching at this stage, sedimentation will generally keep pace with subsidence. Local depocentres would be related to few, steep normal faults with attached accommodation structures and shallow catchments. Because of the moderate surface gradients and only few and minor fault escarpments, it is reasonable to assume that source areas for sand must be sought outside the marginal platform fault system.

Traps generated at this stage can be buried to very great depths (several kilometres) during the total graben subsidence. In addition to potential overmaturation due to deep burial, the scarcity of source rocks may be a general problem for prospectivity of traps related to this stage, because terrestrial conditions are likely to prevail. However, exceptions to this are for example the West African rift basin lacustrine shales.

In cases where later extension has caused significant rotation of the pre-rift succession, the burial problem may be locally avoided, with pre-rift strata riding structurally high, for example along the marginal areas of the Viking Graben tilt blocks. As a consequence of their early establishment in the subsidence history, traps of this type may become faulted and fragmented by later movements. Even though the topography is influenced by active faults to a limited degree, traps related to such structures still may occur.

The general three-stage model suggests that axial transport dominates in the initial graben stage and stratigraphic traps might be generated by the axial fluvial system (Fig. 12.11, trap type 4). The modest tectonic subsidence characteristic of this stage would favour large lateral extension and good continuity of the sand sheet. The stratigraphic traps would be related to meandering or braided river systems and shallow lakes, and the geometry of the system would to a large extent be ruled by the subsidence rates in the incipient central graben. Examples of axially transported sandstones of this type in the primitive Viking Graben include the Statfjord Formation and Lomvi Formation. These sheet sands contrast with the more lenticular sand bodies of the Lunde and Teist formations,

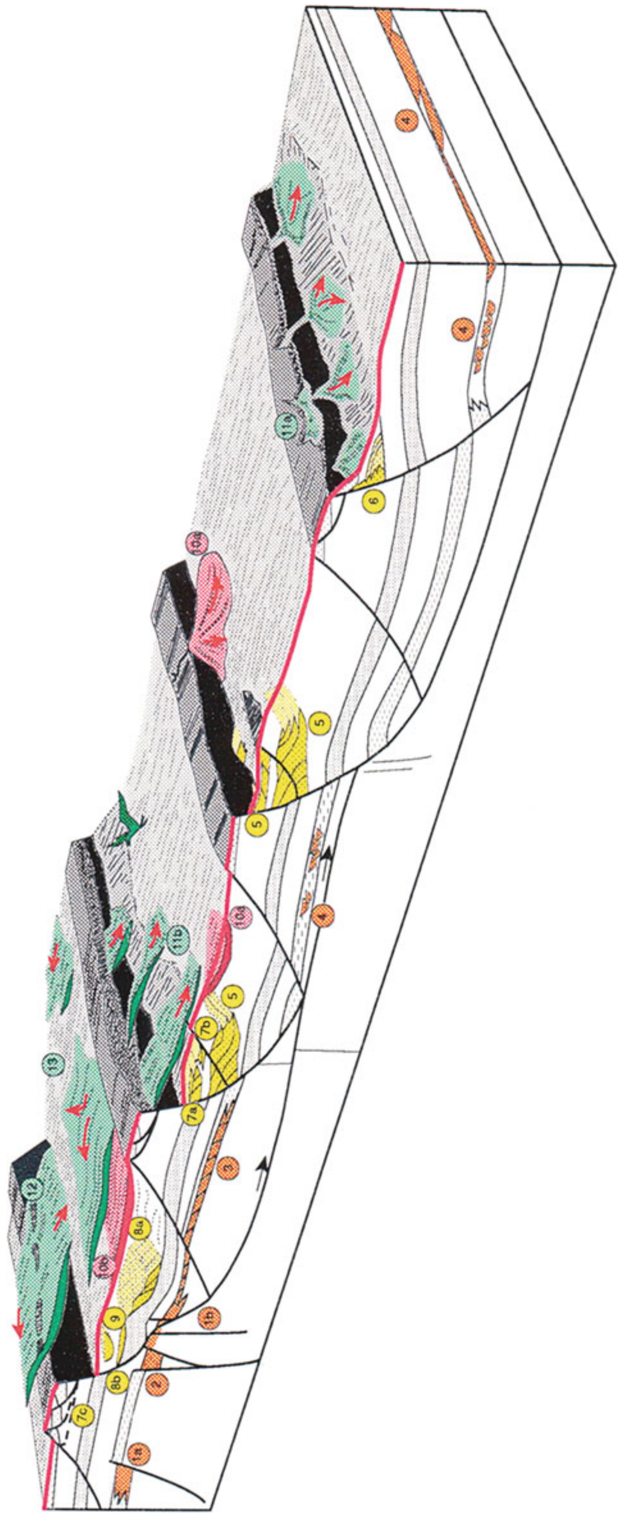

Fig. 12.11 Variations in trap types associated with the pre-rift (*orange*), syn-rift (*yellow*) and post-rift (*green*) stages. The trap types associated with the syn-rift/post-rift transition are marked in *red*. See text for detailed explanation. From Gabrielsen et al. (1995)

deposited under the control of greater subsidence rates.

In the evolving graben which should be characterised by increasing fault activity and eventual magmatism, the continuity of those sand bodies would be later broken.

Units related to transverse sediment transport encompass both pure stratigraphic and mixed traps (Fig. 12.11, trap type 5). Alluvial fan systems, which offer pure stratigraphic traps with potentially great lateral extension and considerable thickness, might represent the most important type. The general model for this stage in the graben development suggests that axial transport systems dominate, but that most of the axial transport comes ultimately from transverse systems upstream.

Clastic fans related to primary synthetic fault-growth represent a mixed trap type, which depends on proximal sealing faults and a distal pinch-out (Fig. 12.11, trap types 1 and 2). Taking the likely moderate relief, low surface gradient and simple fault geometry into consideration, it is probable that the reservoirs in this type of trap are characterised by good lateral continuity in the dip direction. The thickness of the source rock may be considerable in stable basins.

Clastic fans related to accommodation structures along master faults (Fig. 12.11, trap type 1b) classify as mixed palaeomorphic/structural traps. Because of the steep geometries of the master faults in the initial rifting stage, the accommodation structures will be laterally restricted transverse strike, and to constitute a trap of some size accordingly demands good continuity along strike. This is especially the case in late stages of development where channel amalgamation and progradation lead to a sheet-like geometry.

12.3.1.2 The Active Stretching Stage

During this stage the area of subsidence narrows, and the structural elements typical for mature grabens start to appear. The subsidence, and hence the sedimentation pattern, which to a large extent will be influenced by the faulting within the graben area, is characterised by numerous fault-bounded depocentres that are only partly interconnected. The geometry of faults will change from steep planar to low-angle, either by fault-plane rotation or by development of listric faults. Both low-angle rotational and listric faulting will enhance the internal graben relief by rotation and upheaval of fault-block crests and by isostatic and flexural adjustments of the edges of the fault blocks.

The fault pattern, which is initially relatively simple and stable and consists of isolated fault strands, develops into linked structures, and the relief is enhanced. This gives the possibility of accumulation of considerable thicknesses of sediments, and with potentially good strike-continuity in the hangingwall fault blocks of the master faults. Such catchments may be fed with sediments which are transported across the fault scarp from the foot-wall hinterland, and also from the eroded crests of the neighbouring hangingwall dip slope. With continued subsidence, the initiation of accommodation structures (hangingwall anticlines, antithetic faults and forced folds) will restrict both the area of deposition and the continuity of sands. Simultaneously, fault complexity will increase with the development of different types of fault-related transfer zones and steps, which in turn may act as foci for drainage systems.

Finer-grained pelagic and other mature sediments derived from the hinterlands would dominate over the coarser sandy or conglomeratic deposits eroded from local structural highs, although the latter may dominate in isolated basins. Initially, the sediment transport is still essentially axial, but due to enhanced relief in the central parts of the basin will be branched to some extent. A growing local influence from relief caused by rotated fault blocks is expected. This may cause sediment transport parallel to the graben axis also in the more distal parts of the graben system. Locally this pattern will be broken by transverse transport systems cutting across highs related to rotated fault blocks, or in connection with relay ramps and bridges between fault blocks. In broader tilted fault-blocks or in grabens immediately adjacent to the hinterlands, the volume of transversely-transported coarse sediment may be great.

Compared to the initial stage, the final burial depth attained by traps generated in the active stretching stage will of course be shallower. Due to the likely marine influx and sub-basinal restricted conditions, the chances for generation of source rocks will also be higher, and semi-regional or local areas of enhanced subsidence may promote the presence of local pockets of source rocks in addition to those of regional significance.

Fans related primarily to synthetic growth along faults give potentially a mixed trap type, but may also

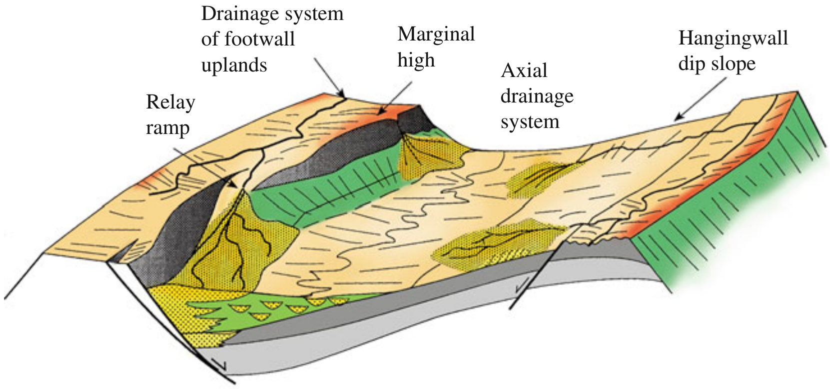

Fig. 12.12 Drainage systems associated with rotated fault block and associated fault scarps. Note that the fan systems associated with the fault scarps may be discontinuous and that sand is also derived from the rotated hanging wall fault block. Modified from Nøttvedt, Gabrielsen and Steel (1995)

include stratigraphic trap elements (Fig. 12.11, trap type 5). The trap depends on the sealing of the fault zone. This trap type is initiated before accommodation structures are activated, and may therefore be characterised by wide extension towards the basin centre. The trap size and complexity depend upon sediment influx relative to basin floor subsidence rate and the stability (potential for reactivation) of the master fault.

In these syn-rift traps the width of the belt of reservoir sand/conglomerate out from the master fault zone is directly related to the subsidence rate and the size of the crestal area being drained. Abundant sediment supply and moderate subsidence rates will cause larger radius fans to develop, whereas high subsidence rates cause all clastic sediment to be trapped in much narrower belts in the immediate vicinity of the fault.

This trap type is well known from the late Jurassic interval in the North Sea, where the most prominent examples are the Brae Field and the Magnus Field. Because the coarse-sediment gravity-flow sands and conglomerates are restricted to narrow fault-parallel belts, they commonly have problematic low continuity in this same direction (Fig. 12.12).

Fans associated with accommodation structures may develop where extended fault activity takes place. Such structures are likely to develop in the hangingwall close to the master fault. The type and geometry of the accommodation structure depend on the geometry of the master fault as well as on the amount of subsidence. Because of development of a fault scarp and the instability associated with such scarps, the foot-wall fault block is likely to be the major source for the hangingwall catchment, even though the hangingwall dip slope may represent the more extensive surface of erosion. Examples of stratigraphic traps in the hangingwall dip-slope position are documented several places in the Norwegian shelf, and are well described from the western margin of the Viking Graben.

The first stage of hangingwall deformation in listric faulting may be normal drag, followed by development of a hangingwall anticline ("roll-over") which mirrors the curvature of the master fault. A structural low is defined on the proximal side of the hangingwall anticline, and opens for entrapment of sediments (Fig. 12.11, trap type 6). Traps in this position are primarily mixed stratigraphic/palaeotopographic and may be related to a local unconformity. Since displacement rates along faults are likely to increase with progressive displacement, it is possible also locally for subsidence to outpace sedimentation with time, to create considerable space for deposition along active faults.

A special type of palaeotopographic trap related to hangingwall anticlines may be expected where forced folds or fault-bend folds above flats in an irregular fault plane have caused surface deformation. As in regular hangingwall anticlines, there is room for sediment entrapment between the master fault and the forced-fold anticline (Fig. 12.11, trap type 9). This trap type is in principle equivalent to that of the conventional roll-over described above.

In addition, sediments may be trapped along the flanks of the forced fold (Fig. 12.11, trap types 8a and

8b). Where fault-bend folds are associated with a large flat with a significant inclination to the regional dip of the master fault, and where sediment supply is good, the trapped volumes might be considerable. In fact, the total volume may further be increased since the depocentre will move as the hangingwall is transported across the flat, causing stacking of the units. The major disadvantage foreseen for this type of stratigraphic trap – especially in view of its distal position – would be the difficulty for sediments to bypass the hangingwall anticline. On the other hand, the model allows the possibility of eroding the top of the anticline, thus utilising a local source area. Large-scale fault-bend folds are not uncommon (e.g. Njord Field), but we are not aware that stratigraphic traps related to such features have been reported in the literature from the North Sea.

By continued subsidence along the master fault, antithetic faults are likely to form on the distal flank of the hangingwall fold, and a graben develops parallel to the master fault. If the sedimentation rate does not keep up with subsidence the graben may influence the drainage pattern, and generate a sediment trap as well. This eventually heralds the shift from an open to a closed system. At starved basin margins the accommodation graben may further restrict development of the primary fan, and in systems where the graben floor is close to the marine level, the top of the accommodation structure may be eroded, and there may be local sediment transport towards the master fault (Fig. 12.11, trap types 7a, 7b).

Examples of this type of trap are found in the "Ula trend", which locally defines the margin of the Central Graben. In block 2/2 a system of antithetic faults, partly triggered by halokinesis, has developed a local high, contributing to the initiation of a graben where the sands of the Ula Formation have been trapped.

Gravity slides across fault escarpments may occur along the crestal areas of the rotated fault blocks (Fig. 12.11, trap type 7c). These are the most unstable areas of the graben system and have their instability enhanced by the flexural cantilever effect, which predicts accentuation of uplift in this area. Eventually, slight inversion may further contribute to the uplift of fault-block crests along basin margins.

In reflection seismic data, erosion of uplifted and rotated fault block crests may be the most easily detectable effect, but deformation of the escarpment by gravity sliding should not be neglected. The basic transport mechanisms would be block sliding, rock fall, or mass flow. In all these cases large volumes of reservoir rocks may become resident in the hangingwall in proximal or distal positions relative to the master fault.

Different types of gravity slides have been reported in this position in front of rotated fault-blocks from the North Sea, but have not so far been deliberately drilled.

12.3.1.3 Unconformity Traps Related to Transition from Active Stretching to Thermal Cooling

The transition from active stretching to thermal cooling is manifested by a change in style of subsidence pattern. In the active stretching stage, fault blocks rotate away from the graben axis, causing similar tilting in the sedimentary cover. The thermal cooling stage, however, causes tilting of strata towards the graben axis, mainly because the most rapid subsidence takes place along the graben centre (Fig. 12.9). In strongly asymmetric grabens, this picture will of course be modified accordingly.

Overall transgression and onlap of the crestal areas may be expected towards the end of the active stretching stage because of the overall subsidence of the rifted area, and because sediment starvation is not uncommon when subsidence outpaces sediment yield. On a regional scale relief may be enhanced by upheaval of graben shoulders due to isostacy and elastic response to faulting. The spatial distribution and the magnitude of elevated (possibly eroded) and subsiding areas is influenced by the geometry and depth of the detachments, and by whether the crustal thinning happens during simple or pure shear. Accordingly, these factors will be of importance to the development, distribution and preservation of stratigraphic traps.

The accelerated axial subsidence, which is likely to occur during the early stages of cooling because of the exponential nature of the thermal decay, will normally be associated with marine transgression, and at the stage where earlier graben walls are onlapped and drowned, a break-up unconformity will develop. Depending on the nature of the subsidence, the unconformity will be diachronously onlapped both along the axis and transversely in the graben system. Depending upon the relation between subsidence and sedimentation rate, isostatic stability of the graben margins and individual fault blocks, the unconformity may be very complex. In this situation a number of unconformity traps may develop.

Truncational unconformity traps are formed by erosion of units deposited during the active stretching stage. Fault block rotation and incipient compaction of the sediment package are likely to cause slightly tilted units which may be eroded and sealed by the thermal subsidence shales (Fig. 12.11, trap type 10a). In principle, this trap type may form in all parts of the graben system, but it is most likely to develop along the graben shoulders.

On-lapping unconformity traps are found above the unconformity, and are dependent upon whether erosion on nearby highs has taken place or not (Fig. 12.11, trap type 10b). The relief across master faults may be considerable at this stage, and it is likely that there is significant erosion and reworking of sands from the marginal highs at this stage. Since the fault activity is retarded, the graben relief will diminish during this process, and the traps may be of considerable lateral extent and contain large volumes of sand.

12.3.1.4 Stratigraphic Traps Related to Thermal Subsidence and Sediment Loading

At this stage sediment transport may be both axial and transverse within the graben system, but it is likely that the transverse systems will dominate in the basin margin areas close to major hinterland relief. Minor continued fault activity should still be expected along the master faults of the graben margins due to isostatic adjustments. These areas will also act as pivots during the shift in subsidence pattern. Altogether this implies relatively smooth graben slopes, with an increased possibilty to develop thick and extensive sandsheets with axes oriented transversely to the graben axis.

Because of the high rate of subsidence of the graben floor at this stage, deposition is likely to take place in a marine environment and the axial basin may be starved of sediment. As the thermal gradients of the system approach equilibrium, the basin will fill in and finally level out the relief completely.

Basin-margin fans (Fig. 12.11, trap type 11a) represent a well-described trap type. These are true stratigraphic or palaeotopographic in type, and will normally have great lateral extent. Their thickness will depend upon the degree to which the graben relief was levelled out when deposition took place.

As the graben is expected to be filled by water, transport agents may be gravity mass flows and turbidity currents (Fig. 12.11, trap type 11b). Large transport distances are therefore possible and the submarine fans may be completely separated from the delta systems along basin margins, particularly in periods with low-stand of sea level (Fig. 12.11, trap type 12). Examples here include some of the main reservoirs in the North Sea, like the Frigg, Forties and Bruce fields.

The *platform-vergent fans* are closely related to the graben-vergent fans, but occur between crests of rotated fault-blocks (Fig. 12.11, trap type 13). During infilling of the graben, sedimentary packages in areas with the thicker sedimentary fill will tend to suffer a stronger compaction than areas with thin packages. This results in development of a hangingwall compaction syncline, which, if it has surface expression, may act as a local sediment trap.

The platform-vergent fans will, however, be more restricted than the basin-vergent ones, and are more dependent on a local sediment source. These circumstances make this trap type less attractive due to small potential sediment volumes.

12.4 Strike-Slip Systems

The strike-slip structural regime is characterised by horizontal orientations of σ_1 and σ_3, whereas σ_2 is vertical (Fig. 12.2). Hence, the orientation of the plane of τ_{max} is vertical, which is also the orientation of the master faults. The displacement along the master fault will be in the horizontal plane (parallel to strike) and the faulting includes initiation of a complex system of secondary fractures.

The general development of shear systems can conveniently be analysed by the use of analogue mechanical experiments. Strike-slip systems are highly dynamic, and the geometry of the initial stages is very different from that of the mature stages of development. Figure 12.13a shows the relation between the structural elements at the initial stage of strain for a right-lateral (dextral) shear system. To analyse the shear system, one decomposes the shear-forces into compressional and tensional vectors by constructing a vector parallelogram. The dominant features define a system of conjugate fractures (Riedel- and Riedel'-shears) that are related to the compressive component of the shear. Of these, the

Riedel-shears are synthetic to the major shear (meaning that they are sub-parallel and have similar shear-sense), whereas the Riedel'-shears are oriented at a large angle to the regional orientation of the fault zone and also display an opposite (antithetic) shear-sense. In addition to the Riedel- and Riedel'-shears, contractional structures (reverse fault, thrusts and folds) with their axes oriented 90° to the compressional stress component may develop. Due to their favourable orientation and synthetic sense of shear, with continued movement the Riedel-shears tend to become dominant at the expense of the antithetic Riedel'-shears. In accordance with the orientation of the tensional vectors, a set of tensional fractures (T-fractures) may be initiated with orientation parallel to the compressional component of the system.

By continued displacement there will be a tendency for both Riedel- and Riedel'-shears to rotate and for the tips of Riedel-shears to become joined to constitute a system of linked fractures striking parallel to the main shear trend. These Y-shears require that some displacement has taken place and are accordingly not present at the initial state of shear. Another set of shear fractures, P-shears, also occurs at a more advanced stage of development. These are oriented at an angle of 60° to the contractional component (Fig. 12.13a) and are accordingly symmetrical with the Riedel-shears, with the Y-shear direction as the plane of symmetry. Dynamically, the P-shears nucleate at the profound, principal fault trace and develop up-section to create an array of fractures with an *en echelon* geometry.

In total, the interaction between the sets of secondary fractures (R, R', T, P and Y) contributes to the complex geometry of the strike-slip fault and generates an uneven and step-like morphology. The arrangements of the steps in left-stepping and right-stepping arrays generate contrasting stress-configurations along the strike of the strike-slip fault, depending on the relative shear-sense (Fig. 12.13b). Thus a right-lateral (*dextral*) shear that affects a right-stepping system of strike-slip fault branches causes extension in the overlap-zones (ramps or bridges) between the individual fault branches, whereas a left-stepping system generates overlap zones of contraction for the dextral system. For a left-lateral (*sinistral*) shear-sense, the relations are opposite.

This implies that a variety of structures, and hence a variety of hydrocarbon trap types, are likely to develop along a strike-slip fault. In the ideal case, where the fault trace is planar and the movements of the opposing fault blocks are absolutely parallel, the trace would be one vertical plane. But since this is the case only for very restricted segments of strike-slip faults, there will be segments where material is squeezed up and out of the fault zone, and cases where slivers of the footwall and hangingwall fall into the fault zone. In both cases the faults are likely to have a steeply dipping root, creating diagnostic geometries for strike-slip faults called (positive and negative) flower-structures (Fig. 12.13c). In cases where distinct fault segments overlap, but are not in direct contact, zones with pull-apart basins or turtle-back structures will occur, whereas extensional and contractional duplexes will develop where the fault-segments are in contact in zones of releasing or restraining bends.

Movements in shear-zones are in many cases not entirely parallel, so that a contractional or extensional component adds to the shear. These situations are called transpressional and transtensional, respectively, and contribute to exaggerating the morphological expressions of the structures described above. In such cases there is a tendency for forces to decompose along weak beds in the deforming units so that strain is taken up in separate systems. Thus a transpressional stress can be decomposed into a pure contractional regime and a pure strike-slip regime. The process is called *strain partitioning* and is well known for e.g. the transform delineating the western Barents shelf, where a dextral transpression is decomposed into collision in the West Spitsbergen Fold- and Thrust Belt and shear along the Hornsund Fault Zone.

12.4.1 Hydrocarbon Prospectivity in Strike-Slip Regimes

Large-scale strike-slip systems may be highly dynamic depositional systems for sediments and also offer a great variety of structural and sedimentary traps. However, compared to extensional basins, there are two significant differences. Firstly, the thermal development is different in that the steep dips and deep roots of the master faults are likely to cause very significant and fast thinning. This involves the substratum of the basin down to the level where the faults detach, which may be top of the lowermost crust or even the Moho. This implies that the thermal gradient

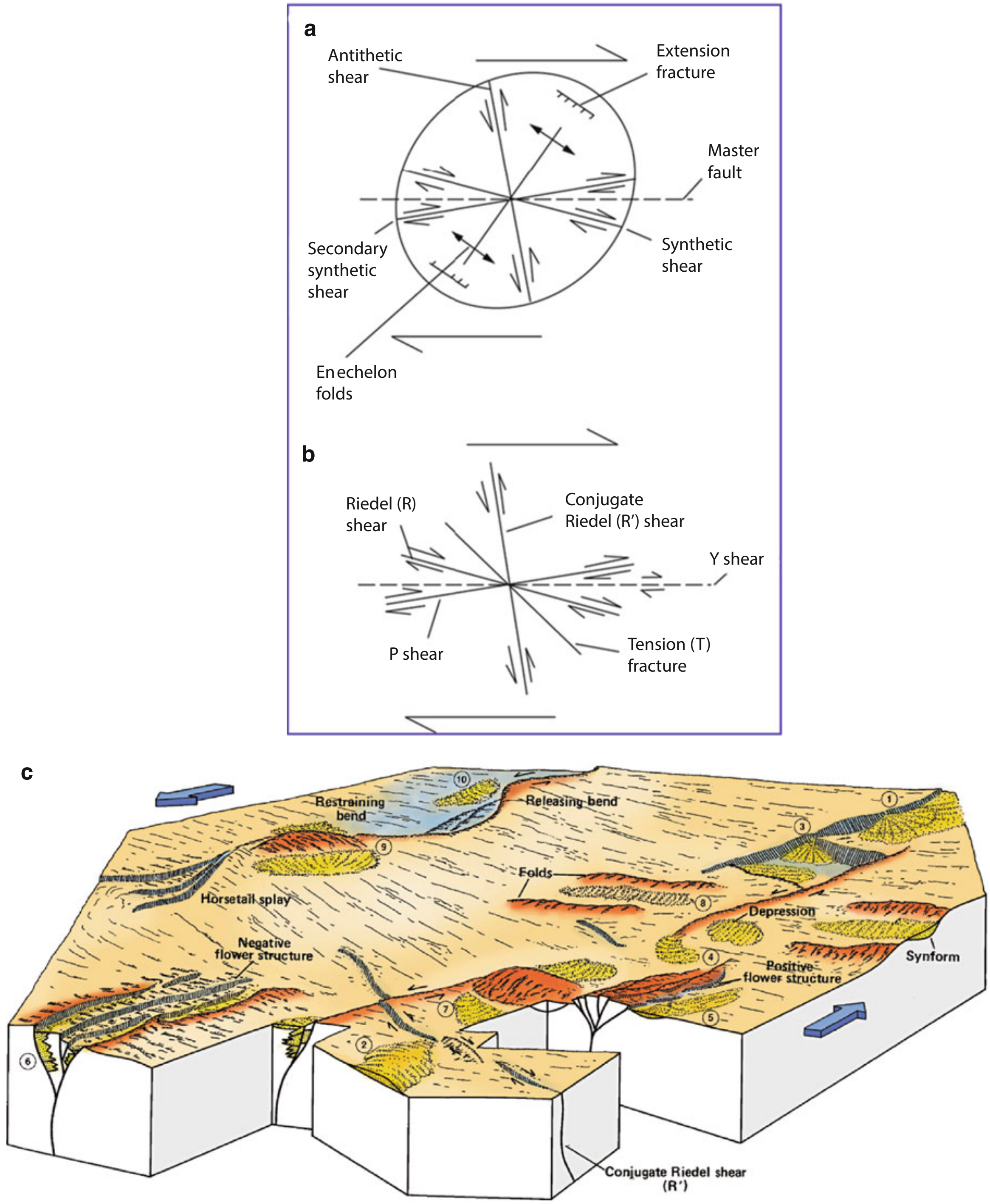

Fig. 12.13 (**a–b**) Fracture types typical for a dextral strike-slip system. By advanced stages of strain, synthetic shears (Riedel shears) and fractures associated with the master fault (Y-shears) tend to dominate the geometry of the fault zone. (**c**) Most common hydrocarbon traps in strike-slip systems. Types of structures are indicated in the figure. (**a**) and (**b**) are redrafted from Crowell (1974)

may very rapidly steepen during early stages of the basin formation and that significant leakage of heat may start before the syn-rift stage is passed. This may be accompanied by fast burial and maturation (and perhaps over-maturation) of the source rock, and also make basin modelling difficult. Secondly, the geometry and tectonic position of strike-slip systems are such that the likelihood for accumulation of marine source rocks is less than for extensional basin systems now situated in passive margin settings.

On the other hand, the structural complexity and variability may develop structural and stratigraphic trap types that are not found in extensional basin systems. Examples are flower-structures and arrays of anticlines that may be found along the strike-slip fault at regular intervals.

12.5 Contractional Regimes

When reading the following, it is important to remember that *compression* characterises the stress system, whereas *contraction* describes the physical process of shortening. Contractional regimes are associated with large-scale orogenic processes (formation of mountain chains), but areas of shortening also may occur on local scales within regional extensional or strike-slip realms. Compression is characterised by the principal stresses being oriented so that $\sigma_1 > \sigma_2 > \sigma_3 = \sigma_{\mathrm{h\,max}} > \sigma_{\mathrm{h\,min}} > \rho gz$. In other words, the smallest stress acts in the vertical plane so that the most energy-efficient way of shortening is by transferring excess mass towards the surface. Also, due to the orientation of the principal stresses, the dip of the plane of maximum shear is 30°, promoting the development of thrust faults (Fig. 12.2).

Orogens are extremely mobile tectonic zones and such systems may accommodate displacements that are several orders of magnitude greater than that which is typical for extensional basins. This implies that the pattern of deformation may be very complex and involve a number of stages or "phases", each phase representing a unique set of stress conditions and p,T-relations. Still, even the most intricate pattern can be analysed by the utilisation of relatively simple geometrical methods and modelling techniques.

The major strain in a large-scale contractional system like an orogen is generally associated with plate margins. In the deep and central parts of the oregen, deformation takes place under high to extremely high p,T-conditions. Such settings are not optimal for the generation and accumulation of petroleum resources and will therefore not be considered further here. However, in the upper (shallow) parts of an orogen as well as along its frontal parts, sedimentation and structuring take place during p,T-conditions that are compatible with the generation and accumulation of hydrocarbons. Here basins related to the interior development and collapse of the central part of the orogen will occur, particularly in the later stages of the mountain building, whereas foreland basins may be active throughout the entire life of the orogen.

12.5.1 The Architecture of Thrust Systems

The most important building blocks of orogens are folds and contractional faults. Both these types of structures affect rocks volumes, so that they may be constrained from their surroundings by a certain geometry and intrinsic style of deformation. Thrust faults tend to climb up-section, because this is the direction of σ_3. Still, contractional faults are characterised by shallow dip and a tendency to flatten over greater distances, particularly where they follow beds of low mechanical strength. They are also frequently seen to merge with other faults along horizontal fault strands and surfaces between mechanically weak beds, and they regularly intimately affiliate with folds. This is so because the folds and faults frequently are seen to develop in concert. One example can be initial buckling of a bed followed by a fault breakthrough along the fold hinge. In other cases folds can develop in front of an advancing fault (fault-propagation folding). In both these examples, the folds would be asymmetrical with the longer fold limb dipping at a shallow angle away from the transport direction, and the shorter fold limb at a steeper angle. The fold axes would be oriented transverse to the direction of transport and may constitute structural traps of considerable magnitude. By continued shortening the faults may link up, trapping isolated, lens-shaped rock bodies, which commonly incorporate folded beds. The lenses are referred to as *horses* and where they are stacked between a horizontal floor fault and a roof fault, they constitute a *duplex* (Fig. 12.14a).

In cases where the faults propagate systematically in the direction of the front of the contractional

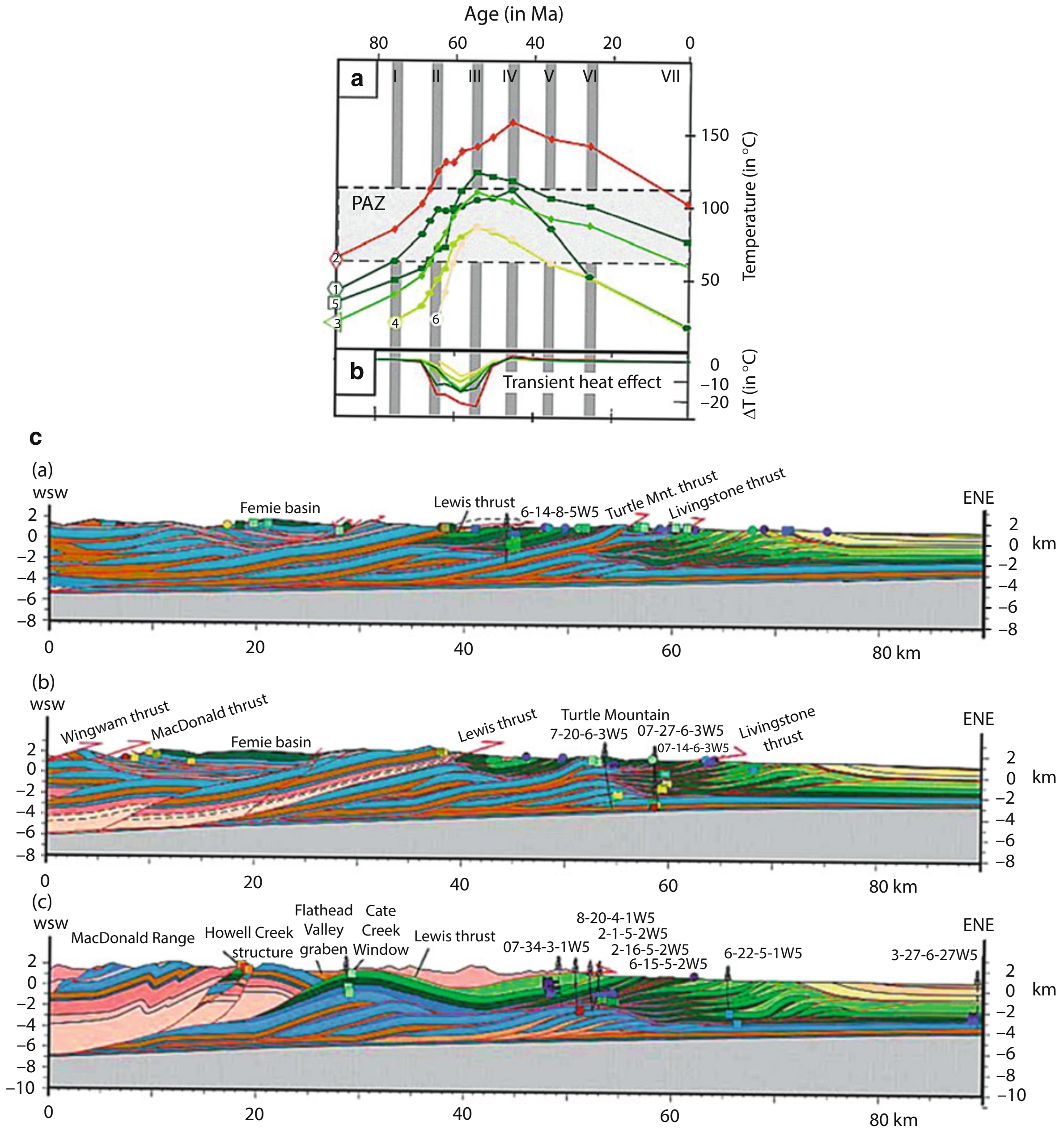

Fig. 12.14 (**a–b**) Temperature-history curves for different parts of the Canadian Cordillera at different positions relative to the apatite annealing zone (PAZ). (**c**) Structural cross-sections with corresponding organic maturity indications (*blue*: low maturation, *dark red*: over-maturation). *Light blue, green* and *orange* colours indicate oil – gas maturation for the same profile. From Hardebol et al. (2009). Reproduced by permission of the American Geophysical Union

system, detachments become linked by climbing fault branches. Together these faults may generate duplexes developed by foreland-directed in-sequence thrusting (Fig. 12.14b). In cases where the system halts e.g. due to increasing friction, the younger horses may pile up on top of the older ones, and an antiformal stack is generated. Changes in overburden and friction may also cause fault activity to switch from one place to another in an unsystematic way. This is termed out-of-sequence thrusting (Fig. 12.14b). Thus, it is not uncommon that shortening is accommodated by hinterland-directed thrusting. In cases where the single faults do not become joined along a roof fault, a system of “blind” faults may develop, terminating at a tip-line. Alternatively, the faults may break the surface. In rarer cases the faults climb down-section in the direction of transport. This is an indication that the strain rate is greater along the roof-fault than it is along the floor fault.

The regions where contractional faults climb up-section are termed ramps and the total geometry of the fault is that of a ramp-flat-ramp. In such cases, the pure geometry of the fault planes forces the strata inside the horses to become folded. The folds reflect the geometry and steepness of the fault plane because the front of the horse depends on the cut-out angle of the original ramp. Ramps parallel to the transport direction may also influence the development and the geometry of the thrust system. Such features may potentially separate subunits of contrasting deformational style. In cases where strain rates are not similar across the ramp, shear and strong rotation occur, and when the structures propagate to affect the surface topography they may strongly influence the depositional systems associated with the mountain chain.

It is obvious that the tectonic processes described above produce a variety of structural traps, among which anticlines with along-strike closure and horses delineated by sealing faults may be the most obvious. Because the subsurface structuring also per definition affects the topography during mountain building, different types of subtle and stratigraphic traps are also likely to be generated (Fig. 12.14a).

12.5.2 Hydrocarbon Prospectivity of Contractional Regimes

The very dynamic character of contractional systems obviously produces a variety of sedimentary systems and structural and stratigraphic traps. The relief associated with orogens normally is measured in kilometres. Strong erosional forces, gravitational instability and climatic influences are important parameters in the development of mountain chains. The system is flooded with a variety of clastic erosional products, the mineralogical composition of which reflects the types of rocks that are involved in the orogen in the first place.

This implies that reservoir rocks and hydrocarbon traps of all kinds are abundant. Because the mountain chain necessarily is uplifted, however, organic-rich marine deposits of the kind that would produce the source rock, would be rare. An exception to this would be cases where a source rock deposited before the contraction started becomes involved in the orogen. In such cases, the critical factors would be the depth of tectonic burial of the source rock, the geothermal gradient of the greater orogeny and the positioning of the source rock relative to the reservoirs. In the dynamic environment of a nappe pile, it must be taken into consideration that units now separated by tens of kilometres may have been juxtaposed at the time of maturation and migration.

Three principally different basin settings can be distinguished. *Intramontane basins* are collapse structures or structural lows generated by folding and thrusting inside the realm of the mountain chain. Those active at the peak tectonic activity are likely to trap large amounts of coarse clastics over a short period of time. They are in most cases of restricted size and source rocks are rarely associated with them. *Foreland basins* are far more interesting from a hydrocarbon exploration point of view. These are stabilised as accommodation areas for sediments due to the gravity load of the progressing orogen, and may trap the bulk of the sediments eroded from the rising mountains and transported towards the orogenic front. The central parts of foreland basins may reach thousands of metres in depth and constitute deep marine depositional systems. Thus, the foreland basin may offer a whole range of sedimentary environments from fluvial, via shallow marine to deep marine. The structuring in the foreland basin position is moderate, but increasing during the progressive development of the orogen. Thus, the basin will eventually become overrun by the advancing deformation front and cannibalised. In this process, a variety of structures are developed from gravitational extensional to contractional complexes of folds, thrust sheets and duplexes. In *basins related to subduction zones*, different types of accretionary prisms may provide source rocks and reservoir rocks, as well as traps, both stratigraphic and structutral. However, these are very dynamic systems, sometimes too dynamic to provide low-risk exploration targets. Also, the sediments in such systems are likely to be too fine-grained to provide good reservoirs.

Although perhaps more rare than in extensional regimes, source rocks may be involved in orogens, such as in the Canadian Cordillera. Detailed studies here performed by Hardebol and his co-workers reveal a complex maturation history, where both tectonic and sedimentary burial have to be taken into account. Although some mountain chains may reveal surprisingly homogeneous bulk geothermal gradient patterns, the individual tectonic units may have undergone contrasting histories of burial and uplift, making a full tectonic restoration necessary in the evaluation of the hydrocarbon maturation of the system (Fig. 12.14).

12.6 Structural Inversion

By the term "structural inversion", or simply "inversion", we generally mean a system of extensional structures that has subsequently undergone contraction. This implies that the principal axes of stress have been changed from being orientated such that

$$\sigma_1 > \sigma_2 > \sigma_3$$
$$=$$
$$\sigma_v > \sigma_{h\,max} > \sigma_{h\,min}$$
$$=$$
$$\rho g z > \sigma_{h\,max} > \sigma_{h\,min}$$

switches to

$$\sigma_{h\,max} > \rho g z > \sigma_{h\,min}$$

This implies that the dip of the plane of maximum shear will switch from 60° to 30° (Fig. 12.15). Thus, although an established zone of weakness will represent a potential zone of reactivation when structural inversion occurs, it is unlikely that the already established faults will be able to accommodate much strain, meaning that new faults with lower angles of inclination will be initiated. The most common characteristics of an inverted system are:

- Reverse reactivation of (extensional) faults
- Generation of new, low-angle fault traces
- Development of secondary contractional structures (folds, reverse faults, thrusts)
- Uplift of basin margins
- Uplift of central parts of basins.

The most common configuration at an early stage of inversion of a fault is shown in Fig. 12.15a. In this case, the accommodation space generated in the hangingwall during extension is completely filled by sediments. During inversion of the master fault, these sediments will be squeezed out of their position in the hangingwall and onto the footwall, accompanied by uplift and folding. For inversion without any oblique component, the fold axes will be oriented parallel to the strike of the extensional fault and, accordingly, orthogonal to the new $\sigma_{h\,max}$. By continued deformation, the pre-existing fault may be squeezed against the hangingwall and become steepened as a consequence, whereas new, low-angle faults generated in the footwall may create local thrusts (Fig. 12.15b).

If one looks at the entire basin, the response on inversion will depend on the geometry of the basin and the mechanical properties of the crust and lithosphere. In the case of a basin that has already been affected by thinning and thermal weakening, the central part may become overdeepened and the basin shoulders uplifted. In contrast, in the case of a mechanically strong basin fill, the central basin may be uplifted, forming an inverted eye-shaped basin geometry. Alternatively, the basin fill may be folded and squeezed out of the basin, as described above for faults.

12.6.1 Hydrocarbon Prospectivity in Basins with Structural Inversion

From a petroleum exploration point of view, structural inversion is an effect that comes on top of and subsequent to the development of a regular extensional basin, and particularly affects the basin margins. Inversion structures may provide additional structural traps as very well exemplified in the mid-Norwegian margin by the Ormen Lange and Helland Hansen structures. On the other hand, inversion invokes an additional risk for breaking of the seal and leakage though reactivated faults. Finally,

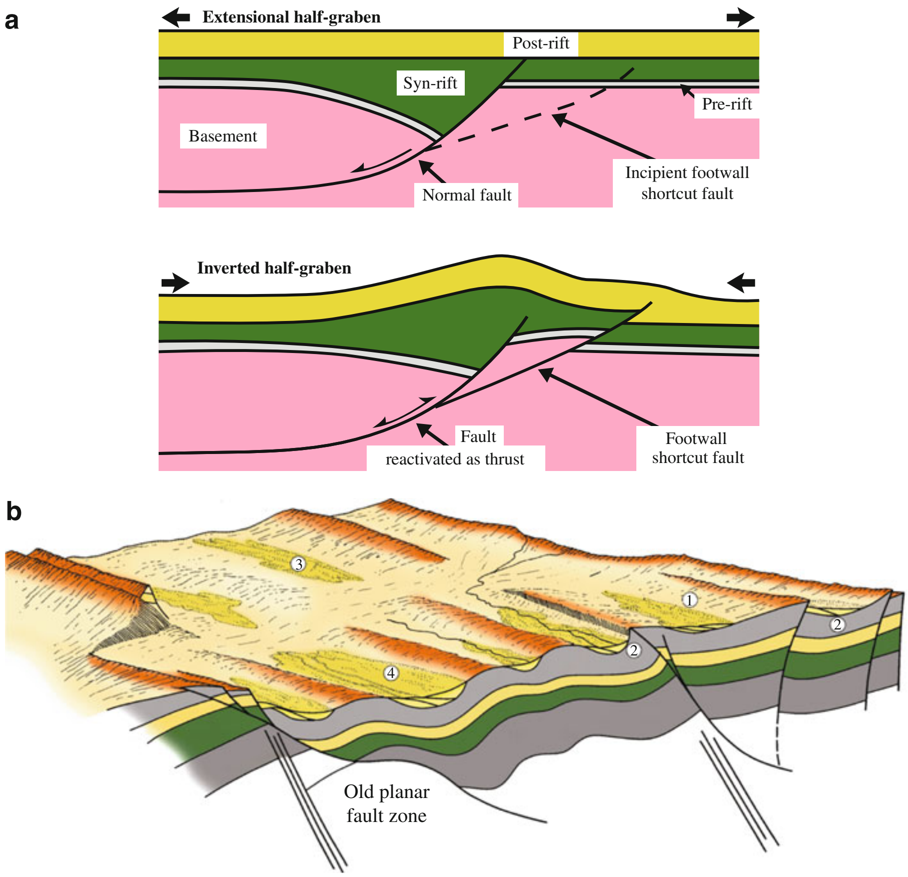

Fig. 12.15 (**a**) Typical geometry of an inverted extensional fault. Note the low-angle footwall shortcut fault (modified from Cooper et al. 1989). (**b**) Surface expressions of deformation in an inverted terrane. Elevated areas, prone to yield erosional products, are marked in *red*, sand accumulations *yellow*. 1: Topographic low inside a pop-up. 2: Topographic low in the hangingwall of an inverted fault. 3: Incipient syncline between a train of emergent anticlines. 4: Syncline between two emergent and eroded anticlines

inversion is commonly associated with uplift and erosion, that in most cases add to the complication of the geological history and reservoir pressure.

12.7 Basins with Evaporites

Rock salt is strictly speaking a crystalline aggregate of the mineral halite (NaCl), which is one of more than 20 evaporite minerals formed by precipitaion from saturated brines, most commonly from solar evaporation. Although salt deposits may contain large portions of other evaporite minerals like anhydrite and gypsum, most studies of the mechanical properties, and hence the dynamics, of evaporites, consider rock salt as the dominant mineral. Although the very strong influence of water on the mechanical strength and flow properties of halite is well established (the deformation mechanism changes from dislocation creep to diffusion creep at a water content as low as 0.05%), there is still not enough data available on the rheological properties of evaporites to predict the detailed strain path and geometry that large evaporite bodies develop when exposed to gravity and loading from a clastic sedimentary overburden. In addition, many evaporite sequences contain a high proportion of clastic material that may be involved in the deformation, and the rheological effect of these clastic "contaminations" is not easily predictable.

Although one tends to associate the problem of seismic imaging in areas with extensive salt deposits to be affiliated with complex salt structures, one should not overlook that many basins contain evaporites that are tabular, flat-lying and stable, and where the problem of seismic imaging is restricted to seeing through the salt as such. It is still fair to say that the structural geology of evaporites has attracted much attention, partly because the high mobility of such deposits poses intriguing structural geological problems, and even more so because salt structures are associated with a variety of structural and stratigraphic traps of significance for the petroleum industry. The structural geology of salt very much reflects the local tectonic environment, be it extensional, contractional or strike-slip. This implies that salt bodies come in an almost infinite variety of shapes, some of which even challenge the limits of the imagination. This variety of shapes, combined with the acoustic properties of salt, poses great challenges for reflection seismic imaging of salt bodies and the strata beneath and close to them.

Gravity-driven deformation at continental margins is generated by the regional gradient of the margin itself and is characterised by upslope extension and downslope contraction. Salt and its overlying sedimentary pile spread in a seaward direction due to regional tilting in response to lithosphere cooling, whereas synkinematic sedimentation induces loading instabilities. At basin scale, thin-skinned deformation may induce extreme upslope salt thinning, leading to the formation of salt welds, as well as massive downslope salt thickening. Most often, the extensional domain can be divided into three sub-domains, which are, in a seaward direction, the sealed tilted block, growth fault/rollover, and diapir domains (Fig. 12.16a). The upper domain is characterised by tilted blocks that are sealed early by synkinematic sedimentation. The rollover domain displays a large amount of extension, whereas the domain of diapirs is generally considered as gently translating, accommodating small amounts of extension. Diapirs correspond to weak zones and are easily and often squeezed. Downslope of the margin, contractional structures balance the amount of upslope stretching.

The domain of shortening is also divided into three sub-domains. In a seaward direction they are composed of diapirs squeezed at late stage, polyharmonic folds and thrust faults developed at early stage, and folds and thrusts developed at late stage. Contractional structures are initiated in a domain located at a distance from the initial salt edge. Compression remains localised in this domain during the initial stages of evolution by continued deformation. The upslope migration of contraction can then reach the extensional domain and squeeze the diapirs.

Analogue experiments show that the overall structural zoning is mainly controlled by the initial condition (salt basin) and the basal slope angle, whereas the type of structures in the structural domains strongly depends on sedimentation rate (Fig. 12.16b).

An early attempt to classify salt structures systematically and to set this into a dynamic context was made by Trusheim in 1960. He suggested that salt impiercements grow from elongated low-profile ridges (anticlines and rollers) triggered by gravitational contrasts, developing into rows of pillows, diapirs and eventually into walls and sheets of salt. The diapirs come in a variety of shapes from regular massive stocks, via irregular masses to elegant mushrooms. This geometric classification is undoubtedly valid for a tectonically stable, evenly subsiding basin. But even this relatively predictable kinematic growth pattern of salt structures causes great problems in seismic imaging due to the complex pattern of internal flow in the salt structure itself, including horizontal displacement, affiliated with the development of overhanging or even horizontal walls.

12.7.1 Hydrocarbon Prospectivity in Salt-filled Basins

Deep parts of basins where salt structures tend to be situated may be excellent sediment traps. The growth of diapirs contributes to the development of local depocentres and the areas around salt diapirs may accumulate large volumes of reservoir rocks of good quality and be associated with excellent structural and stratigraphic traps. However, due to the capacity of salt to flow horizontally at shallow levels and develop overhanging bulges and sheets, and even to become detached from its deeper sources, the detailed geometric configuration around the stem of the salt structure, including its diameter, is commonly disguised and the diameter of the stem itself may be impossible to determine from reflection seismic data.

In addition to the bulge above the salt empiercement itself, which will reflect the geometry of the upper layers of the empiercement, the main types of features that may constitute structural hydrocarbon traps adjacent to salt diapirs are the rim syncline system (sometimes several generations),

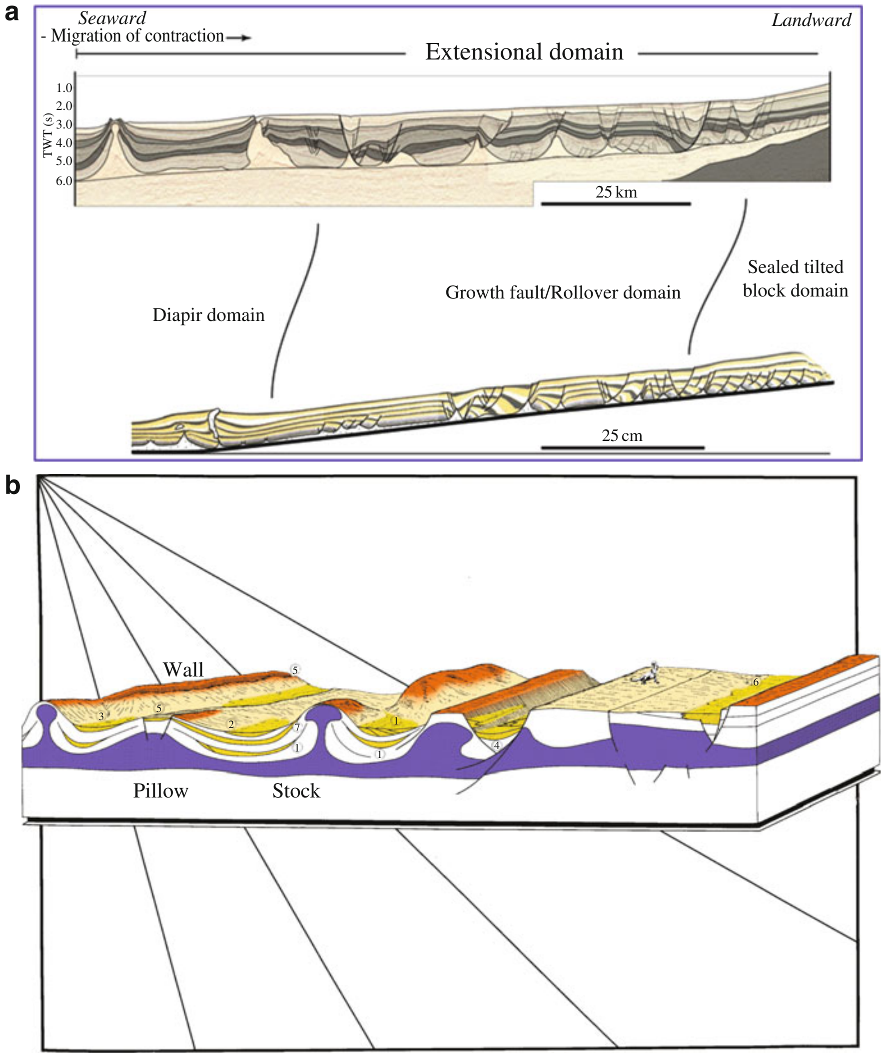

Fig. 12.16 (**a**) Analogue experiment showing salt structures and related structuring of sediments between salt ridges in an extensional, inclined slope (from Brun and Fort 2008). (**b**) Surface expressions of deformation in a terrane affected by halokinesis. Elevated areas, prone to yield erosional products, are marked in *red*, sand accumulations given in *yellow* for different structural positions. 1: Rim syncline. 2: Rim syncline between a stock and a pillow. 3: Rim syncline along a salt wall. 4: Salt-induced graben. 5: Graben on *top* of collapsed salt pillow. 6: Salt-induced rotated fault block. 7: Stratigraphic trap covered by overhang in salt pillow

anticlines associated with the rim syncline system, faults generated due to volume reduction during vertical transport of salt, and drag structures close to the stem of the diapir (Fig. 12.16b). Due to the circular nature of the diapir, all these structures are likely to be closed when seen in three dimensions. In addition, numerous types of stratigraphic trap may be related to any of these structural features. For salt anticlines and simple walls, which have not developed overhangs, seismic imaging of the structures is usually relatively straightforward. For mushroom-shaped diapirs, however, this is much more challenging and several parameters have to be taken into consideration in the structural analysis: the distance of the rim syncline system from the centre of the diapir and its amplitude and wavelength depend on the thickness of the original salt sequence, on the diameter of the diapir, and the salt flow rate relative to sedimentation rate. In many cases however, this structure is situated sufficiently far away from the diapir for seismic imaging to be straightforward. When it comes to the fault systems and the drag structures, these occur close to the stem of the diapir and are likely to be covered and completely obscured by the overhanging diapir bulb.

One particular difficulty in seismic interpretation may occur in cases where dynamic salt interacts with faulting. At the crests of salt diapirs and stocks a combination of ring-shaped and radial fault systems is commonly found, but these are unproblematic in seismic imaging and interpretation. In addition, numerous examples exist of fault activity promoted by the underlying active salt, providing a substratum for detachment. Again, such structural relations are also clearly displayed in reflection seismic data. Finally, though, due to transfer of larger volumes of salt towards the basin axis, smaller pillow-shaped volumes of salt are frequently left and trapped along basin margins, where they interact with the basin margin fault system. In such cases different configurations are sometimes developed in the hangingwall and the footwall of the salt-involved fault, causing complex and contrasting sedimentary conditions across the fault so that significant problems arise in sequence correlation. Where salt has intruded along the fault-plane, interpretation of seismic data may be hampered by reduction of the general data quality.

In more complex tectonic environments (strike-slip and contraction), the complexity and variety of salt body configuration is commonly much greater, because the final geometry of the salt body will be determined by directed flow reflecting varying differential stress and strain. For example, the importance of evaporite sequences in the development of many thrust belts like the Pyrenees and the West Spitsbergen thrust-and-fold belt is well documented. In such settings seismic imaging may be complicated by salt being involved as an extensive, continuous or disrupted unit during the thrusting, and also because it may have accumulated unevenly and become integrated in contraction structures like fold cores and duplexes, and as intrusions along fault planes.

The quality of seismic imaging performance in areas of salt has been greatly improved in recent years. Still, the days of surprises are not yet over when results from drilling become available. The effort in improving seismic imaging techniques must therefore continue. And it should go hand in hand with field study and analogue modelling.

12.8 Faults and Fault Architecture

Fractures (faults and joints) are found in practically all hydrocarbon reservoirs and are crucial elements because they both influence the migration of hydrocarbons within the reservoir and contribute to the entrapment of fluids. Due to great variation in fault rocks and fracture types and their distribution, the influence of fractures on reservoir communication is not easily predictable. It is therefore natural that the analysis of single faults and fracture systems is receiving increasing attention.

For the assessment of the architecture of faults and fracture systems one can, in principle, chose between stochastic and deterministic or a combination of the two. Given the complexity and generalised architecture of larger faults, however, stochastic methods are less favourable in analysis of such features. This also seems to be the case for fracture systems generated in stress situations where σ_1 is distinctly different from σ_3. To provide input to a deterministic reservoir model that aims at taking the complexity of larger faults into consideration, we have performed field studies in order to constrain realistic characteristics for the units that commonly can be defined within the realm of a fault zone.

Since mesoscopic and macroscopic faults affect volumes of rock, and accordingly should be described as composite rock bodies that include a complex system of structures (fault rock, folds and fractures), the term “fault zone” is applied here. In the following we use “fracture” as a general term that includes faults, deformation bands and joints, and we distinguish between shear fractures (microscale) and faults (meso- and megascale). Also, we use the term “high-strain zone” for the parts of the fault core where shear is concentrated.

12.8.1 *The Structural Elements of the Fault Zone*

It is well established that faults are commonly zoned and composed of several units with distinct deformation styles (Fig. 12.17). These include the fault core, where most of the displacement is accommodated, and its associated damage zones that are geometrically and mechanically related to the development of the fault.

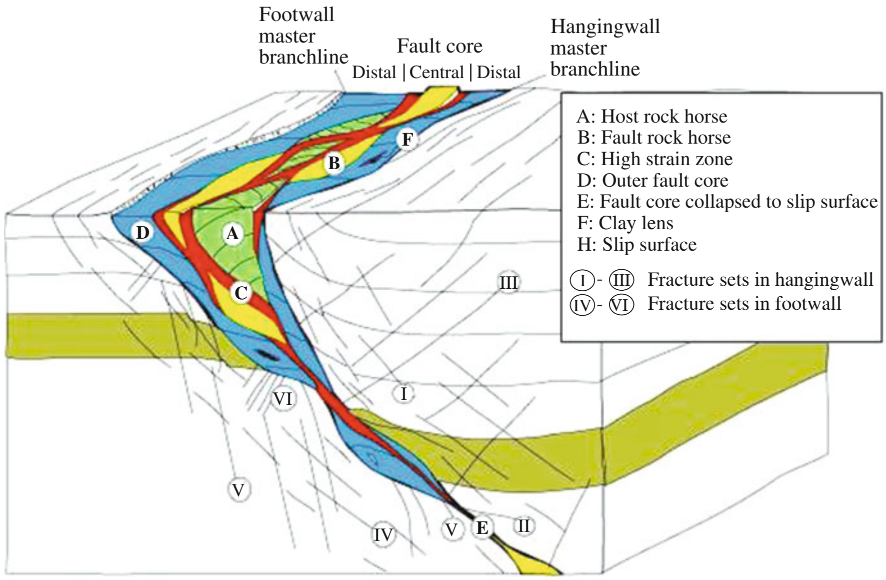

Fig. 12.17 The structural elements of an extensional fault

The fault core is in general separated from the footwall and hangingwall damage zones by distinct fault-branches. Lozenge-shaped rock bodies frequently dominate the cores of extensional faults. These are commonly referred to as fault lenses or horses, which may occur in isolation, as *en echelon* trains, or be stacked to constitute duplexes. The fault-rock lenses may consist of relatively undeformed country rock derived from the footwall or the hangingwall of the fault core. In faults with greater displacement, the fault lenses may represent lithologies exotic to that of the observable footwall and hangingwall or be completely reworked fault rocks like cataclasites and breccias. The geometry of the lenses, their relative arrangement and their relation to intervening high-strain zones are important for the fluid communication along and across faults in cases where contacts between units of high or low permeability control the fluid flow.

Field study of the shape of fault core lenses suggests that such features have relatively regular shapes and that the a:c-ratio (relation between length measured in the dip-direction and maximum thickness) in extensional faults is in the order of 10:1 and that the b:c-ratio (relation between length measured in the strike-direction and maximum thickness) is somewhat less than this, perhaps 9:1 or 8:1. It is also suggested that the lenses are close to symmetrical with reference to both the central a- and b-axes. The available data also suggest that these relations are roughly valid also for the higher-order lenses (2nd, 3rd, 4th order etc.), although there is a tendency for the a:c- and b:c-values to become slightly reduced for the higher-order lenses.

The high-strain zones separating individual or groups of fault lenses may include deformed units that can be recognised as country rock in the footwall and hangingwall, as well as several types of fault rocks, the host rock of which cannot be determined. For example the most intensely deformed zone of the fault core is easily distinguishable and this zone may represent the latest area of deformation, indicating that strain softening has occurred.

The damage zones define a halo of fractures on both sides of the fault core. The fractures of the damage zones are associated with the dynamic development of the fault and may encompass remnants of the propagation of the incipient fracture, commonly termed the process zone. The strain intensity in the damage zones is generally modest compared to that of the fault core, and in sedimentary rocks bedding and other primary features can commonly be recognised. The fracture distribution, frequency and orientation in the hangingwall and the footwall are generally different, and it is therefore convenient to distinguish the footwall damage zone from that of the hangingwall.

Field investigations show that damage zones are not symmetrically distributed around the fault core. Accordingly, it is commonly observed that the fracture frequency curves build up towards their maximum values more steeply in the footwall than in the hangingwall. Also, there are differences where orientation of the major fracture populations is concerned. The footwall damage zone is characterised by a set of fractures oriented subparallel to the footwall master fault-branch. This fracture population may interfere with fracture sets that dip both more and less steeply than the master fault-branch. The hangingwall damage zone is influenced to a greater extent by antithetic fractures to the master fault and a wider area is affected by these fractures. The fractures of the damage zones may be of different kinds, depending on lithology, depth of burial at the time of deformation, and strain intensity.

Faults that are in their early stage of development do not possess the complexity described above. In sandstones they develop from single or arrays of deformation bands that with increasing strain coalesce into one zone of focused shear. This is generally also the case for incipient faults in carbonates.

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