Comparison of modern fluid distribution, pressure and flow in sediments associated with anticlines growing in deepwater (Brunei) and continental environments (Iran)

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ABSTRACT

Differences in fluids origin, creation of overpressure and migration are compared for end member Neogene fold and thrust environments: the deepwater region offshore Brunei (shale detachment), and the onshore, arid Central Basin of Iran (salt detachment). Variations in overpressure mechanism arise from a) the availability of water trapped in pore-space during early burial (deepwater marine environment vs arid, continental environment), and b) the depth/temperature at which mechanical compaction becomes a secondary effect and chemical processes start to dominate overpressure development. Chemical reactions associated with smectite rich mud rocks in Iran occur shallow (\(\sim 1900\) m, smectite to illite transformation) causing load-transfer related (moderate) overpressures, whereas mechanical compaction and inflationary overpressures dominate smectite poor mud rocks offshore Brunei. The basal detachment in deepwater Brunei generally lies below temperatures of about 150 \(^\circ\)C, where chemical processes and metagenesis are inferred to drive overpressure development. Overall the deepwater Brunei system is very water rich, and multiple opportunities for overpressure generation and fluid leakage have occurred throughout the growth of the anticlines. The result is a wide variety of fluid migration pathways and structures from deep to shallow levels (particularly mud dykes, sills, laccoliths, volcanoes and pipes, fluid escape pipes, crestal normal faults, thrust faults) and widespread inflationary-type overpressure. In the Central Basin the near surface environment is water limited. Mechanical and chemical compaction led to moderate overpressure development above the Upper Red Formation evaporites. Only below thick Early Miocene evaporites have near lithostatic overpressures developed in carbonates and marls affected by a wide range of overpressure mechanisms. Fluid leakage episodes across the evaporites have either been very few or absent in most areas. Locations where leakage can episodically occur (e.g. detaching thrusts, deep normal faults, salt welds) are sparse. However, in both Iran and Brunei crestal normal faults play an important role in the transmission of fluids in the upper regions of folds.

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1. Introduction

In the early days of hydrocarbon exploration the identification of surface anticlines associated with oil seeps, contains the implicit understanding that anticlines are sites of focused fluid flow, where some fluids are at least temporarily trapped, and others escape to the surface. As reviewed by Evans and Fischer (2012), today we understand that a very complex interplay exists in folds between fluids, stratigraphy, stages of fold development and fold style. Fluids comprise three key properties: temperature, pressure and chemical composition, each of which is controlled by a number of processes. These properties can vary considerably even around a single fold (Evans and Fischer, 2012). Fluids exert an effect on rock strength, (by permitting pressure solution and other chemical reactions to occur, stiffening the rock by depositing cements, and reducing rock strength by overpressure) and impact structural style. Conversely the structural and stratigraphic permeability architecture influences the location of fluids and fluid migration pathways (e.g.}
Sibson, 1996, 2005). These fluid pathways typically involved downward percolation of meteoric waters through fractures, and upwards movement of warm fluids (particularly along thrusts, e.g. Barker et al., 2000; Beaudoin et al., 2011). Fluid filled zones, predominantly in fractures, are commonly transient features and evidence for their existence is likely to be lost if cements are not deposited. So, while understanding the development of folds by studying veins is very important and allows the sequential development of fold-fluid relationships through time to be determined (e.g. Barbier et al., 2012; Beaudoin et al., 2012; Evans and Fischer, 2012), it is also important to study active folds, and to understand how fluids are distributed at a snapshot in time of fold and thrust development. For example the modern relationships between folds, depth to overpressure, and structural evolution led Cobbold et al. (2009) to suggest that the deepwater thrust front of the Niger Delta tracks the onset of the hydrocarbon maturation window, and that hydrocarbon generation and chemical compaction, not burial-disequilibrium compaction, facilitates the detachment development.

This paper compares the modern distribution of fluids, including hydrocarbons associated with two young (Late Miocene—Recent) regions of detachment and fault propagation fold-type
anticlines and considers the fluid development in terms of the types and timing of overpressures developed with respect to burial during sedimentation and fold development. The depositional environments are very different; one is a shaly-prone deepwater fold and thrust belt, offshore Brunei; the other is a shallow marine to semi-arid continental sequence of deposits (evaporites, clastics, and carbonates) in Central Iran (Fig. 1) (see Hall, 2013 and Moutherau et al., 2012 for reviews of the tectonic settings of these areas). These represent end members of the types of depositional environment in which folds grow and hence permit comparison of a broad spectrum of the processes that influence fluid flow, and fluid migration pathways as fluids are driven from regions of relatively high pressure to lower pressure. Understanding the mechanisms creating overpressure, and the timing of these mechanisms relative to anticline growth are very important to understanding the origin of the fluids and their availability/ability to influence structural development during deformation.

2. Development of overpressures in basins

Zones of overpressured fluids in sedimentary basins are generally transient features that exist until some environmental change causes fracturing of the seals that initiates migration. The fluids towards regions of lower pressure (e.g. Morley et al., 2008). As reviewed by Swarbrick et al. (2002) overpressures in basins are caused by: 1) an increase in applied load (disequilibrium compaction), 2) changes in the volume of the pore fluid within a confined volume of sedimentary rock (fluid expansion), 3) fluid movement or buoyancy related to density differences, and 4) redistribution of overpressured fluids from one site in a basin to a different area (fluid transfer or infiltrationary overpressures).

Disequilibrium compaction encompasses a range of mechanisms where overpressures arise from the failure of fluids to be expelled rapidly enough in response to an applied load including: increasing vertical stress during burial, lateral compression due to tectonic stress, and load transfer due to diagenetic changes. These and other mechanisms are discussed below.

Burial of mud produces a number of important changes to its constituent minerals, which affect how overpressures are generated with increasing temperature and depth. Initially pressure is the most important control on compaction (mechanical compaction). In basins where smectite is the dominant early-stage clay mineral, such as the Gulf of Mexico and the North Sea, the 60–90 °C temperature range marks the zone where the effects of mechanical compaction become less important and temperature-controlled chemical changes become significant or even dominant (see review in Nadeau, 2011). However, other important diagenetic changes can occur shallower. For example in biogenic, silica-rich sediments shallow (typically about 300–600 m below the sea floor) diagenetic transformation of Opal A to Opal C/T can reduce porosity in the order of 15–20% (Negeau et al., 2010), and trigger fluid loss and volume reduction mechanisms that lead to the creation of polygonal faults (e.g. Cartwright, 2011). Likewise, in a megasulphate evaporite basin, the loss of structural water as a body of thick primary gypsum (CaSO₄·2H₂O) converts to anhydrite (CaSO₄), via loss of structural water, can lead to overpressure at depths of less than a kilometre (Jowett et al., 1993).

Chemical processes changes mineralogy, volume, increase shale density, cause dissolution of load-bearing grains, convert bound water to mobile water, and increases in the preferred orientation of clay mineral grains (Ho et al., 1999; Aplin et al., 2006; Day-Stirrat et al., 2008; Lahann and Swarbrick, 2011). Some chemical changes may strengthen the rock framework. For example the smectite-illite transition releases fine-grained quartz that helps to stiffen mudstone. 1–3 μm spherical, discrete grains or short chains and clusters of quartz form a pervasive network when burial reaches temperatures around 80°–85° (Thyberg et al., 2010). This mechanism will retard further mechanical, but not chemical compaction.

Diagenetically and catagenically induced changes can also weaken parts of the rock framework, which responds by compacting (e.g. Meissner, 1981; Osborne and Swarbrick, 1997; Swarbrick et al., 2002). Key changes include: solid kerogen hydrocarbon conversion to liquids, gases and residuals; and grain-size changes associated with the smectite to illite and kaolinite to illite transformations. If fluid cannot escape, and the rock cannot continue to compact, then the load is partially transferred to the fluids (‘load transfer’ type disequilibrium compaction, Swarbrick et al., 2002). This process for the smectite-illite transition occurs at temperatures starting between 60° and 80 °C depending upon the carbonate phases present, and ends around 110 °C (Hower et al., 1976; Buller et al., 2005). Key permeability-reducing processes include the dissolution of smectite and re-precipitation as fibrous illite that intensely subdivides existing pore space (Nadeau et al., 2002), and precipitation of microquartz (Thyberg and Jahren, 2011). Kaolinite undergoes dissolution and precipitation as illite at temperatures around 130–140 °C, but an extra source of potassium (such as feldspar) is needed for the reaction (Bjorlykke, 1998). ‘Load transfer’ can result in fluid pressures that are 1500–3000 psi higher than pressures interpreted from shallow compaction trends (Lahann and Swarbrick, 2011). Goulty et al. (2013) have described high-temperature overpressures from the Kutai Basin, eastern Borneo that fit with chemically-induced overpressures.

It should be noted that the load transfer mechanism in the smectite-illite transition is a different overpressuring mechanism from that originally proposed for this transition, which is a volume expansion mechanism. In the volume expansion mechanism interlayer water in the clay crystal is released during diageneis as free water, which causes expansion of the volume of pore water (Powers, 1967; Burst, 1969; Bruce, 1984). This mechanism is thought to be a relatively minor factor in development of overpressure (Osborne and Swarbrick, 1997). The conversion of kerogen to oil and/or gas, and particularly cracking of oil to gas is regarded as the most significant volume expansion mechanisms occurring in the subsurface (e.g. Osborne and Swarbrick, 1997; Tingay et al., 2013).

3. Overpressure and fluid flow in folds and thrusts offshore Brunei

The latest Early Miocene- Holocene deltaic deposits offshore Brunei formed a thick (>10 km) depocentre that lies both onshore and offshore, and extends into the deepwater area of the NW Borneo Trough. Uplift and rapid erosion of the interior of Borneo provided the sediment source area (Hall and Nichols, 2002; Morley and Back, 2008). Deformation and uplift commenced as a consequence of the Dangerous Grounds continental crust entering and jamming the Palaeogene to Early Miocene subduction zone of the Proto-South China Sea (see Hall, 2013 for a review of the tectonic development). The post-collisional Miocene sand-prone shallow marine deposits are called the Belait Formation, while the Pliocene-Recent equivalent deposits are called the Liang Formation. Shale-prone sequences deposited more seaward on the shelf, or in deepwater are known as the Setap Shale Formation (e.g. Sandal, 1996). The basin that received these sediments along the NW margin of Borneo is known as the Baram-Balabac Basin (Cullen, 2010).

Deformation of the Baram-Balabac Basin comprises in part classic gravity-driven delta-style growth faults on the shelf and toe folds in the deepwater area (Morley et al., 2008, Figs. 1C, 2). There is
also a lithospheric stress component to the compressional deformation, which has resulted in: a) inversion of some of the growth-fault depocentres particularly onland and along the inner shelf, b) propagation of folds and thrusts under the deltaic deposits, and c) caused an additional component of shortening to the deepwater fold and thrust belt (Morley et al., 2003, 2008, 2011; King et al., 2009b; Sapin et al., 2013). The amount of extension in the shelf region is less than the shortening in the deepwater area (Hesse et al., 2008; King et al., 2009b). Consequently deformation in the deepwater fold and thrust belt is thought to be driven both by lithospheric stresses and by gravity (Morley et al., 2008, 2011). The exact cause of the lithospheric stress component is uncertain, one possibility is crustal-scale gravity collapse (Hall, 2011). This mixture of gravity and lithospheric stress-driven deformation is seen in the modern stress field where the maximum horizontal stress is sub-orthogonal to the shoreline (appropriate for compression) along the inner shelf and deepwater area, and sub-parallel to the shelf (appropriate for extension) in the outer shelf-upper slope region (Tingay et al., 2009b; King et al., 2009a).

3.1. General model for fluid migration in Brunei

In deepwater basins the shallow sediments are very under-compacted, with up to 60–80% of the sediment volume comprising seawater trapped in pore spaces (see review by Day-Stirrat et al., 2010). From the deposition of turbidites and mass transport complexes, to burial and deformation, structures associated with dewatering (e.g. flame and dish structures, fluid escape pipes, sand injectites, polygonal faults) occur at a variety of scales and are important features of the deepwater environment (e.g. Hurst et al., 2005; Cartwright et al., 2007). In addition to the vertical escape of fluids, a component maybe expelled oceanwards, laterally from the thicker section of the deltaic shelf (Van Rensbergen and Morley, 2000). An additional source of fluids may come from the under-thrust Dangerous Grounds block. Although the setting is not as dynamic as a classic accretionary prism setting Zielinski et al. (2007) noted that the mean high heat flow ($83 \pm 66.5$ mW/m$^2$) in the landward half of the deepwater fold and thrust belt compared with the seaward half ($59 \pm 22.6$ mW/m$^2$; Fig. 3) followed a similar pattern to accretionary prisms, where fluid is expelled laterally (e.g. Barnes et al., 2010).

In Brunei, overpressures arising from burial-induced disequilibrium compaction cause relatively low density and low sonic velocity responses on well logs in the zone of overpressure (Tingay et al., 2003, 2007, 2009a). The overpressure zones plot on the burial loading curve for normally pressured intervals on the porosity-effective stress cross-plot (Tingay et al., 2009a). These characteristics reflect the halting of mechanical compaction at some point along the burial curve, and the preservation of anomalously high porosity, for a particular depth of burial. The occurrence of burial-induced disequilibrium compaction has been extensively documented for the shelf area of offshore Brunei (Tingay et al., 2009a). However, some overpressures in the shelf area, particularly shallow overpressures, are interpreted as caused by fluid expansion (inflation) because they have no large associated porosity anomaly, and plot above the burial loading curve on sonic velocity-effective stress cross plots (Fig. 4; Tingay et al., 2007). The overpressures migrate from regions of primary overpressure development via carrier beds, faults, fractures or fluid pipes and are forced into a confined pore volume laterally or higher in the section.

In the preceding “Development of Overpressures” section several key chemical reactions are related to the transformation of smectite to illite. The chemical composition of shales in NW Borneo is not well documented, but the data that exist indicate smectite is a minor proportion of the clay minerals. Hence the smectite-illite
transition is not regarded as a significant factor in overpressure generation in the basin. In surface outcrops of the Jerudong area, Middle Miocene Setap Shale samples from both the country rock, and from shale dykes (e.g. Morley et al., 1998; Morley, 2003) are composed of about 40–45% kaolinite, 35–50% illite and 10–20% mixed layer smectite-illite (Fig. 5). Recently deposited clay minerals sampled from the Malay Peninsula, North Borneo and South Borneo are predominantly kaolinite and illite + chlorite, with less than 14% smectite (Wang et al., 2011; Liu et al., 2012). These data suggest the low percentage of smectite in the Jerudong samples is a characteristic feature of weathering in the region and not because smectite has been converted to illite. In the absence of a smectite-rich mud rocks, the onset of significant overpressures due to chemical effects is unlikely to occur in the 60–90 °C range discussed by Nadeau (2011) and shown for illustrative purposes in Figure 6. Kaolinite to illite transformation can trigger load transfer overpressure, and will also release bound water (about 8% of the rock volume for 40% kaolinite). However, this reaction is very slow below temperatures in the range of 120–140 °C (Bjorlykke, 1998).

Figure 3. Results of bottom sampling of the deepwater area of Brunei. A) Heat flow, B) Hydrocarbon content from piston cores. Data from Zielinski et al. (2007), modified from Morley et al. (2011).

Figure 4. Sonic velocity-effective stress plot for 1400 repeat formation tests from 31 fields across Brunei. Grey dots and normally pressured points that define the loading curve. Black triangles are overpressured points (>11.5 MPa/km). Overpressured points lie both on and off the loading curve. Points that plot on the loading curve likely represent disequilibrium compaction, while points that lie off the loading curve are likely to be the result of fluid expansion or vertical transfer. Modified from Morley et al. (2008).

Figure 5. Clay composition of outcrop samples of Setap Shale, Brunei Darussalam. The samples are from a mud diapir, intrusive shales, and country rock, in the Jerudong area (described by Morley et al., 1998; Morley, 2003).

Figure 6 shows a plot of the depth and temperature at top of overpressure in Brunei. The top of overpressure for many wells occurs at relatively low temperatures (40°–60°), at depths between 1000 m and 2300 m. In some cases, for example in the Mergani-1, Laksamana-1 and Merpati-1 wells on the outer shelf (Tingay et al., 2009a), the top of overpressure is related to burial-induced disequilibrium compaction. Conversely, the top of overpressure in wells on the inner shelf is related to inflationary overpressure (e.g. Peragam-1, and Bugan-1 location a, Fig. 6), and the Bugan-1 well also shows a deeper onset of disequilibrium compaction overpressures (Fig. 6, location b).

The top of 6 overpressure points in the temperature field (>60 °C) where mixed chemical and mechanical compaction operate (if the smectite content of the mud rocks is high; Fig. 6 A). However, there is no strong support in Figure 6A for temperature control of overpressure development following the predictions of Nadeau (2010). A comprehensive analysis of wells throughout the
region clearly indicates that the depth of onset of overpressure is dependent on a range of variables, not just temperature and pressure. For example the structural/stratigraphic location of a well is important, particularly in growth faults where drilling the depo-centre (e.g. Maharaja Leyla-1, Fig. 6A) demonstrates the occurrence of a very thick, well-drained sequence of alternating sands and shales. Consequently the maximum depth to the top of overpressure exceeds 4 km. In contrast on the outer shelf, overpressure in shale prone-sequences occurs much shallower in the section (1e1.5 km; Figs. 2 and 6A,B). The top of overpressure on the inner shelf is commonly associated with an inflationary mechanism, while the top of overpressure on the outer shelf is related to disequilibrium compaction (Tingay et al., 2007, Fig. 6B). This boundary coincides with the occurrence of Miocene–Pliocene inversion structures on the inner shelf, and the absence of inversion on the outer shelf (Tingay et al., 2009a,b). Hence, there is the strong indication that inversion is promoting the migration of fluids upwards in the section. Those fluids include hydrocarbons; the largest fields in Brunei are associated with inversion anticlines along the shoreline and inner shelf. The high overpressures encountered in the 1953 Seria, 1974, 1979 Champion Field internal blowouts (Tingay et al., 2005) are associated with inversion anticlines.

Figure 7 schematically shows how three main pulses of fluid migration may have developed in the Miocene–Pliocene deep-water shales of Brunei. Initially expulsion of fluids under generally hydrostatic conditions occurs during compaction. But, even in the shallow environment, local overpressures may develop for a variety of reasons including: a) local development of a low porosity zone (perhaps by shallow cement development, or the presence of a gas hydrate seal), b) rapid loading by mass transport deposits, and c) inflationary overpressures transported along a pipe, fault, or fracture. A very large loss of porosity occurs at shallow depths, and by about 1.5 km depth the initial porosity (initially about 60–70% for deepwater shales) is reduced by half (e.g. Dawson and Almon, 2006). However, for overpressured shales the porosity loss will be arrested at some point in the burial history and fluid expulsion will slow, or cease (fluid isolation depth), and overpressures begin to build.

Nadeau (2010) suggested that thermo-chemical reactions could explain the overpressure origins in the Baram-Balabac Basin. However, in Brunei the temperatures at the top of disequilibrium compaction overpressure can be as low as 40 °C, as well as greater than 80 °C (Fig. 6). The low-temperatures to the top of overpressure and smectite-poor mud-rocks clearly indicate that mechanical, not thermo-chemical processes are driving overpressures in the basin.

In an extensional stress environment hydraulic fracturing and a pulse of fluid expulsion will occur when the pore fluid pressure exceeds the minimum horizontal stress plus the tensile strength of the rock. Repeated pulses of burial, pressure build up, and fracturing will occur until the pore fluids are depleted. For offshore Brunei further burial causes increased overpressures and triggers another cycle of fluid expulsion through chemical diagenesis and catagenesis (hydrocarbons). Chemical diagenesis includes kaolinite to illite and a relatively minor smectite to illite transformation. In the Golden Zone model the upper temperatures of this deeper zone tends to be around 120 °C (e.g. Nadeau, 2010). However, for offshore Brunei the low proportion of smectite, and high kaolinite content of the shales suggests that this zone begins and ends at higher temperatures than those typically used in the Golden Zone model. Additionally load transfer-type disequilibrium compaction mechanisms are less important than volume expansion mechanisms related to hydrocarbons.
Another likely contributor to overpressures in the deepwater fold and thrust belt is tectonic (horizontal)-related disequilibrium compaction, which can produce a significant amount (up to 26%) of early ‘ductile’ shortening (e.g. Butler and Paton, 2010). Quantifying the contribution of this mechanism in mud rocks is difficult since compaction is not controlled only by the vertical effective stress, and typically tectonically-induced overpressures have no measurable impact on compaction (e.g. Yassir and Addis, 2002). Tectonic loading can generate near lithostatic overpressure since in the general case the minimum principal stress is vertical. Qualification of this statement is necessary because the local stress state around folds can vary significantly from extensional through strike-slip to compressional. Overpressures in shales are developed (e.g. Cobbold et al., 2009), since widespread overpressure is the main facilitator of deformation. Hence extensional growth faults, and down-dip folds and thrusts will accelerate fluid depletion from the zone of disequilibrium compaction overpressures, into adjacent formations, and ultimately to the surface (e.g. Morley et al., 2008).

3.2. Fluid migration along the basal detachment

Figure 2 shows a cross-section passing from the deepwater fold and thrust belt to the shelf counter-regional extensional fault systems. The approximate locations of the 60 °C, 100 °C and 150 °C geotherms have been plotted. As discussed in Section 2, the interval between the 60 °C and 100 °C geotherms typically marks the transition where mechanical compaction starts to be replaced by chemical processes (although this is to a lesser degree than smectite-rich provinces). The detachment to the outer-most folds of the deepwater fold and thrust belt lies within the zone where overpressures are related to mechanical compaction-induced disequilibrium compaction, whereas the majority of the folds and growth faults detach within a zone with temperatures above 150 °C that is now well within the zone of load transfer-induced disequilibrium compaction and volume-expansion mechanisms. Although deltaic detachments are commonly said to be associated with mobile shales, the modern temperatures appear to be far to high for the shales to be mobile in response to loads in the same
way that salt is mobile. The shales are likely to be well compacted, but weak due to lithology, high overpressures, and the development of low-friction polished (graphitic) slip surfaces (for the potential importance of the latter see Rutter et al., 2013). The low critical taper of the fold and thrust belt wedge indicates the presence of near-lithostatic pore fluid pressures along the basal detachment (Morley, 2007a).

As is well documented for accretionary prisms, fluids are likely to be pumped from the inner part of the wedge along the basal detachment to the more external parts of the fold and thrust belt (e.g. Fisher and Hounslow, 1990; Moore et al., 1995; Ruppel and Kinoshita, 2000). An example of this plumbing is the megaseep (Zielinski et al., 2007) that affects anticline VI (Fig. 3; anticline labelling from Morley, 2009). The megaseep has a maximum heat flow of 604 mWm², which Zielinski et al. (2007) calculated requires that the fluids originated from around 6 km depth. If this source were located vertically beneath the anticline it would lie below the detachment. However, detachments are thought to act as barriers to vertical fluid migration (e.g. Wawrzyniec et al., 2003). Consequently it is more likely that the source is from within or above the detachment zone, about 20 km to the SE of the surface flow (Fig. 2).

The distribution of hydrocarbons in the deepwater area falls into an oceanward zone where hydrocarbon seeps are very rare (2% of all sites) and only one site showed a heat flow greater than

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**Figure 9.** Seismic examples of folds, deepwater Brunei. A) Line drawing, and B) seismic line showing the difference in development between a BSR (bottom simulating reflection) developed on a simple anticline, and one affected by a mud pipe and degradation complex. The BSR over the younger, simpler anticline to the east is deeper than the BSR over the complex anticline. 1 – Top (?) of gas hydrates appears to act as a shallow detachment for shallow minor normal faults. 2 – BSR crossing mud pipe is closer to the surface than at 1, suggesting the mud pipe is episodically still active and warmer than the surrounding sediments. HA, HB, HC are locations of regionally mapped horizons (Morley, 2009). TWTT = two way travel time. C) and D) details of BSR in anticline affected by crestal normal faults. C) Typical amplitude display. D) Spectral colour palette, which highlights amplitude variations along the BSR and underlying bright spots. Double BSR is evident as an alignment of bright spots parallelising and underlying the BSR.
99 mWm², and a more landward zone where 16% of sites showed abundant oil-associated hydrocarbons and 20% of sites exhibited high heat flows (Zielinski et al., 2007). Thermogenic gas is only present in the landward zone, but sites with mixed biogenic and thermogenic gases are present in both zones (Warren et al., 2011). Warm active chimneys of fluids associated with thermogenic organics are also present, (Warren et al., 2011). The high-displacement thrust associated with the Anticline VI and the megaseep appears to have diverted the migration of hot fluids along the detachment so that the outer fold zone did not receive these fluids in abundance (Fig. 2). Temperatures beneath the outer folds probably permitted some generation of hydrocarbons, but volumes appear to have been relatively low.

3.3. Fluid migration within anticlines

The crests of anticlines are the confluence of fluid migration from depths of the basal detachment at around 6 km to more local

![Stratigraphic column for the Central Basin, Iran (Morley et al., 2013).](image-url)
fluid migration in the upper 1 km. Fluids migrating from depth along faults commonly depart from the fault trace and rise as vertical pipes through the crests of anticlines (Figs. 8 and 9), and may form a chain of mud volcanoes at the sea floor.

As anticlines grow on the sea floor they are modified by gravity, material may slump from the forelimbs and crests of anticlines when they become too steep, and/or gravity-driven crestal normal faults develop (Morley, 2007b; 2009). The faults can be local pathways for fluid migration in about the upper 1 km of section (Figs. 2 and 8), and since they dip away from the fault crest, can help feed gas towards the anticline crests to help form gas hydrate layers. In the shallow deepwater subsurface carbonate cements are commonly developed in response to microbial degradation of methane, and gas hydrates can be associated with authigenic carbonate cements (Dela Pierre et al., 2010; Warren et al., 2011).

Bottom simulating reflections (BSR) related to gas hydrates are common features of anticlines in the deepwater of Brunei (Laird and Morley, 2011, Fig. 9). Bright spots beneath the BSR show a positive AVO response and are related to free gas trapped by the gas hydrate saturated layer (Laird and Morley, 2011, Fig. 9C). These gas accumulations occur mostly on the backlimbs of fault propagation folds, reflecting the effects fold asymmetry that focus fluid migration onto the back limb. Differing depths to the BSR reflect variations in shallow geothermal gradient, which for some anticlines in the deepwater of Brunei range from 4.8 °C to 6.4 °C/100 m (Laird and Morley, 2011). These unusually high gradients are interpreted to reflect the migration of hot fluids from depth into the anticlines (Laird and Morley, 2011). Detailed studies of shallow fluid flow around gas-hydrates in deepwater folds shows fluid migration occurs through existing conduits such as faults, and feeder beds until they reach a depth where fluids are able to fracture their way vertically to the surface (e.g. Morley, 2009; Sultan et al., 2011, Fig. 8).

4. Folds in central Iran

The Central Basin of Iran lies in the upper plate of the Zagros Mountains Collision Orogen, which formed during the Late Palaeogene-Neogene as a consequence Arabia-Eurasia collision (e.g. Morley et al., 2009; Mouthereau et al., 2012). The Central Basin of Iran displays many folds developed within the plateau region of an orogenic belt, in a large basin influenced by normal faults that was inverted in the Late Miocene to Pliocene (Morley et al., 2009). The basin stratigraphy comprises a basal evaporite (halite-dominant) and red bed unit (Lower Red Bed Formation) of Oligocene age, which overlies Eocene volcanics (Fig. 10; Furrer and Sonder, 1955).

The Lower Red Bed Formation is succeeded by a 1–2 km thick, Late

Figure 11. Location map for the Central Basin in the Qom area, Iran. A6, A5, A14 − Alborz wells, S1 − Sarajeh-1.
Oligocene to Early Miocene marine sequence of limestones, sandy limestones, marls with occasional black shale and anhydrite layers, called the Qom Formation. The sequence is capped by the Miocene Upper Red Formation, the lowest section comprises halite-dominated evaporites a few hundred metres thick, capped by red bed clastics. The thickness of the formation varies greatly, but can be up to 6 km (Morley et al., 2009).

In the Central Basin the source of the fluids is either meteoric, or formation waters. The Miocene to Present Day climate is semi-arid, and consequently the continental sediments do not contain a high percentage of low-salinity meteoric water in the shallow subsurface after deposition. This has resulted in the Miocene-Upper Red Formation being a compacted, dense formation even very shallow in the section. In the Silak-1 well from near surface to 500 m depth, bulk density increases from 2.1 g/cc to 2.4 g/cc; the average density increases to 2.48 g/cc below 1500 m, and increases slowly to over 2.5 g/cc until about 4000 m.

The composition of the Upper Red Formation is dominantly volcanic lithics, with significant contributions from carbonate lithics, opaque minerals, and feldspar. Cements are dominated by zeolites (including analcime), calcite and iron. With burial augite, biotite and chlorite are replaced by hematite, and the Fe cement

![Seismic line across the Silak Anticline](image)

**Figure 12.** Seismic line across the Silak Anticline (see Fig. 11 for location). A) Uninterpreted line, B) interpreted line. Hydrocarbons and overpressured fluids leak through weld in evaporites (I) and up a normal fault (II).
content increases. This reflects a sediment source predominantly from the Eocene volcanic arc sequence (including interbedded carbonates, clastics and evaporites), which forms an extensive belt in central Iran.

A study on the diagenesis of two clay-rich sections in the Upper Red Formation in central Iran was conducted by Amini (2001), the following description is based on their work. Early weathering reactions involve aluminosilicate glass + water transforming to smectite + alkaline solutions. The alkaline solutions can develop in three ways: a) (evaporative concentration) saline fluids + zeolites, b) evaporative concentrations + CO$_2$ = a) above + calcite, and c) (drainage) ~ which leaves a smectite-dominated assemblage. In terrestrial environments with good drainage or relatively high precipitation zeolites would be lost, while in arid conditions pore water saturation and evaporite-derived solutions favour zeolite growth. Amini (2001) found that the URF contained stable smectite-dominated assemblages between depths of 400 and 1900 m. In parts of the basin analcime and other zeolites helped buffer high pH pore fluids and stabilize K-smectite. In one thick section in the northern margin of the URF basin, starting around 1900 m depth, smectite was replaced by discrete illite and chlorite and by 4900 m smectite was virtually eliminated.

Four anticlines Aran, Silak, Sarajeh and Alborz (Fig.11) have been drilled during hydrocarbon exploration. Three of them (Silak, Sarajeh, Alborz) are capped by Upper Red Formation halite seals, while halite is absent in Aran anticline and only thin metre-scale anhydrites are present. The folds are detachment anticlines, which have developed on thick Lower Red Formation evaporites (Morley et al., 2009, 2013, Fig. 12). The Silak Anticline was drilled by Silak-1 during 2008 and 2009, and is the latest exploration well in the basin. The well encountered drilling problems related to flowing salt, overpressures, and highly fractured zones within the Qom Formation.

Figure 12 is a dip 2D seismic line across the Silak Anticline that shows the fold detaches in the Lower Red Formation, above a thick, relatively strong Eocene volcanic-dominated sequence. The anticline was probably localized by the presence of a normal fault. Displacement on the normal fault is relatively low in the seismic section and increases in displacement passing eastwards. The fold was probably caused by the buttressing effect of the normal fault, but instead of inversion occurring along the normal fault, the detachment fold style developed in the hanging wall (further east inversion does occur along the normal fault). The URF shows a local synclinal depocentre is developed in the lower part of the formation on what is now the northern limb of the fold. This depocentre is interpreted to have resulted from local loading of the URF evaporites and downbuilding during localized salt withdrawal. Subsequently the anticline began growing and the upper part of the URF shows thinning towards the fold crest. A number of normal faults, with convergent dips affect the crest of the anticline.

4.1. Overpressure development

The pressure gradients in wells down to the top of the Upper Red evaporites ranges from hydrostatic to moderately overpressured (Fig. 13). The evaporites are mostly pure halite, but interbeds of anhydrite and clastics, mostly claystones, also occur. A 10 m thick, overpressured, red, sticky clay zone within the evaporites associated with a thrust presented a particular problem to drilling the Alborz Anticline (Abai et al., 1963). The presence of poorly lithified clays indicates that very locally disequilibrium compaction was operating due to the sealing potential of halite.

4.1.1. Upper Red Formation overpressures

The Upper Red Formation above the evaporite unit can be normally pressured (Alborz wells) or exhibit some degree of overpressuring (Aran-1, Silak-1). The distribution and connectivity of sands within the Upper Red Formation is highly variable laterally and vertically, which affects drainage and overpressure development. The absence of associated porosity anomalies or density reversals in overpressure zones indicates burial-type disequilibrium compaction is an insignificant contributor to overpressure development. As discussed above, chemical reactions are very important within the formation and smectite to illite transformation is a very significant factor starting at around 1900 m depth (Amini et al., 2011). Geothermal gradients are moderate 2.5–3°C/100 m indicating temperatures at this depth are around 70–80°C, assuming a 20°C surface temperature. In Aran-1 the onset of mild overpressure around 2000 m indicates load transfer disequilibrium compaction arising from the smectite-illite transition would be a viable mechanism.

In the Alborz Anticline a thrust in the Upper Red Formation detaches within the evaporites and has transmitted overpressured fluids, including oil along the fault zone, both within the evaporites, and higher in the Upper Red Formation (Abai et al., 1963). The only source rock intervals lie within the Qom Formation, hence despite the evaporite seal, the presence of the thrust and/or hydraulic fracturing has enabled fluids to leak from the Qom Formation. Not surprisingly the anticline is under-filled with hydrocarbons.
In Silak-1 there is no obvious Alborz Anticline type-thrust on seismic reflection data, but gas and oil have still managed to leak into the Upper Red Formation, where they were encountered in convergent, conjugate extensional faults at the crest of the structure (Fig. 14). The crestal normal fault zone coincides with flows of overpressured fluids into the wellbore between depths of 1450—2050 m depth (Fig. 14). The fluid is rich in calcium (~1000 ppm) and chloride (~1000 ppm) and also sulphate, which caused considerable expense since a large quantity of soda ash had to be added to the drilling mud to neutralize the calcium; drilling rates slowed. The origin of the calcium could be from diagenetic processes within the Upper Red Formation. Alternatively, the fluids are deeper CaCl2-basinal brines derived from within the lower part of the thick evaporites and associated hydrothermal/volcanic waters near the base of the sequence. According to Abaie et al. (1963) typical Qom Formation water samples are highly saline, with a specific gravity at 20 °C of 1.21 (similar to the Dead Sea brines), and are rich in calcium (25,000 ppm), sulphate (900 ppm) and chloride (207,000 ppm). If the fluid in the crestal normal fault is partly derived from the Qom Formation then it has been considerably modified during migration by either chemical reactions and/or mixing with fresher water (perhaps meteoric water migrating down the crestal normal faults).

In Silak-1 overpressure begins shallow (~1600 m) and is associated with crestal normal faults (Fig. 14). There is no density reversal on well logs, and the overpressures are probably at least partially inflationary as indicated by the presence of oil, which must have originated from the Qom Formation. Around 3300—3600 m the overpressure magnitude decreases in a more sand-prone, and hence better-drained interval. However, a shallower decrease in overpressure from about 1.8 specific gravity (SG) to 1.45 SG around 2800—3000 m occurs within an interval of uniformly high density. Possibly the presence of faults is responsible for this decline in overpressure (Fig. 14B).

4.1.2. Overpressures in the Qom Formation

Despite potential breaches to the Upper Red evaporite seal, very high overpressures occur within the Qom Formation in the Alborz, Sarajeh and Silak anticlines (Figs. 13 and 14). These overpressures are separated from lower-magnitude overpressures in the Upper Red Formation by the intervening evaporites, which can reach thicknesses up to about 400 m. The most spectacular demonstration of the high overpressures is the 1956 Alborz-5 well, which exited the Upper Red evaporites by only a few centimetres and blew out at a depth of around 2677 m. The mud column of 2.07 g/cc (19 ppg) density was ejected and the well then initially ‘produced’ about 80,000 barrels of oil a day, plus an unknown quantity of water and gas for 25 days, the well was brought under partial control and finally self sealed 80 days after the blow out (Abaie et al., 1963). In total an estimated 5 million barrels of oil was expelled during the blow out. The formation pressure at the top Qom Formation is estimated to be > 8000 psi (Gretener, 1982). Three other wells (Alborz-9, 10 and 11) each produced about 30,000 barrels per day under less spectacular circumstances. Alborz-6 was drilled 2.6 km away from Alborz-5, below the oil/water contact. The well encountered high pressures (~2.18 g/cc) at the top Qom Formation, indicating that there was no pressure relief from the Alborz-5 well, which blew out one year earlier.

An intense blow-out also occurred when the first well (Sarajeh-1) was drilled into the Sarajeh Anticline in 1958. The well only had surface casing at the time of the blow-out. Fractures and gas seeps occurred at the surface up to 3 km westwards, 1.5 km eastwards, 7.5 km northwards and 0.5 southwards of the well. The subsurface conditions at the time of the Sarajeh-1 blow-out are unknown, and subsequent wells drilled into the anticline have not encountered high overpressures, indicating the blow-out depleted the pressures over a wider area than the Alborz Anticline.

Figure 14. A) Formation pressure—depth plot for Silak-1. B) Cross-section through the Silak-1 drilling location based on 2D seismic data (see Fig. 12 for example), showing the geological context of plot A. See Fig. 11 for location. Subtle bend in hydrostatic pressure gradient is due to much higher salinity and density formation waters associated with the Upper Red Formation evaporites.
The Qom Formation is a poor conventional reservoir, the upper part is a marl, the lower section limestones and sandy limestones interbedded with marls (Fig. 10), and effective porosity rarely exceeds 6–8%. The limestone intervals are crossed extensively by bed-parallel stylolites, and also tectonic stylolites that lie at a high-angle to bedding (Fig. 15A). Hence a dense network of fractures is needed to connect any remaining porosity. The delivery of high volumes of hydrocarbons along fractures is amply demonstrated by Alborz-5. The types of fracture networks required to produce the performance seen at Alborz-5 are present.

Figure 15. Examples of fracture networks at different scales within the Qom Formation. A) Satellite image of the Mil-Zangar Anticline in the Qom Formation, thrust over Upper Red Formation. B) Detail of image A) (see A for location) showing conjugate fault network. C) Detail of image B (see B for location) showing very closely spaced network of faults spaced in the order of 10’s metres. D) Limestone breccia within fault zone, E) Main fracture types affecting the Qom Formation on sub-vertical bedding surface, conjugate fractures, tectonic stylolites (bedding parallel stylolites are also present) and fault zone (D is a detailed picture of breccia in the fault zone). F) Widespread, intense Type 1 fracturing on a sub-vertical bedding surface.
Formation section was buried under the 400 m thickness of halite-dominant evaporites. During this period, fluids within the formation would have been trapped and isolated from those above. About 50% of the volume of the formation would have followed the carbonate diagenetic pathway of porosity reduction (e.g. precipitation of carbonates in pore spaces, recrystallization, compaction by pressure solution), whereas about 50% of the formation volume would have followed the clastic pathway, with mechanical compaction followed by later diagenetic porosity loss. Today the marls are highly indurated, and are stiff enough to contain the fracture network that fed the Alborz-5 blow out.

The Upper Red evaporite seal is very important to developing and preserving overpressures. The Aran anticline is missing the Upper Red halite-dominant evaporite seal, and the Aran-1 well encountered moderate overpressures in the Qom Formation, but much less than the Alborz wells (Fig. 13). The other three drilled anticlines in the area (Alborz, Sarajeh, Silak) are all sealed by thick sections of the Upper Red halite-dominant evaporites and exhibit high overpressures (Figs. 13 and 14). There are two likely causes of the overpressure operating at present: 1) hydrocarbon generation, and cracking of oil to gas (see Tingay et al., 2013 for an example of this mechanism), and 2) tectonic-disequilibrium compaction arising from shortening and compression of the rock during Late Miocene-Pliocene folding and thrusting. The abundant orthogonal/ high-angle to bedding stylolites present in the Qom Formation are evidence for this process (Fig. 15E).

Figure 16 summarizes the likely development of overpressures in the Qom Formation. Prior to deposition of the Upper Red Evaporites the <1 km thick Qom Formation would have undergone fluid loss due to compaction, but following evaporite deposition the fluids in the Qom Formation were sealed off and probably could not escape. Overpressures, particularly in the smectite-rich marls would have been developed due to burial-disequilibrium compaction and later due to load-transfer. The underlying Lower Red Formation underwent a similar diagenetic development to the Upper Red Formation, and during the smectite-illite transition probably became over-pressured, lost fluid into the Qom Formation and caused some vertical transfer of overpressure. During the Late Miocene folding, source intervals within the Qom Formation approached maximum burial and oil and gas generation and gas cracking would have also contributed to the overpressures (Morley et al., 2013).

It is uncertain whether hydraulic fractures were able to temporarily breach the salt seal during periods of extreme over-pressure. For example, one explanation why the Sarajeh and Alborz Anticlines are underfilled with hydrocarbons is that a phase of hydraulic fracturing naturally depleted the reservoirs, and the salt seal subsequent annealed itself. Such a system has operated in the Ara Salt in the South Oman Salt Basin, where breached halokinetic salt is indicated by bitumen-defined polyhedral edges to biaxially-flattened halite crystals and is locally known as “black salt” (Kukla et al., 2011).

What is clear in central Iran is that at some stage in the burial history oil has leaked in places into the Upper Red Formation, in the case of the Alborz anticline leakage was via a thrust (Fig. 17), while in Sialk-1 it was probably via a salt weld in the local area where a downbuilding syncline is developed in the lower Upper Red Formation (Fig. 12). The final stage in the overpressure history is pore space reduction due to tectonic stylolite development during Late Miocene-Pliocene folding and thrusting.

4.2. Vergence and timing of fluid migration in the Alborz and Sarajeh anticlines

The Alborz and Sarajeh anticlines, lie en echelon to each other and the Qom Formation is present at similar depths in both
Superficially there would be every reason to expect similar fluid histories for each anticline, yet the Alborz anticline has trapped oil, while the Sarajeh anticline has trapped predominantly gas and condensate (Abaie et al., 1963). Hence the Sarajeh anticline trapped fluids formed at higher temperatures, which migrated later than those in the Alborz anticline. Both anticlines are underfilled by hydrocarbons, which indicates that fluid leak off is probably part of the history. Morley et al. (2013) presented a structural and geochemical study that concluded the SW dip of the detachment which was also a key to understand the trapping behaviour. The Sarajeh anticline is an SW verging structure while the Alborz anticline a NE verging structure (Morley et al., 2013). This difference in vergence was sufficient for the Alborz anticline to close relatively early and trap oil, while the Sarajeh anticline was closer to the main source basin depocentre, attained closure later than the Alborz anticline, and so trapped gas. Greater downbuilding of the salt depocentre south of the Sarajeh Anticline enabled the Qom Formation source rock to be buried into the gas window, and also influenced the SW vergence direction of the Sarajeh Anticline. This example highlights how sometimes minor changes in fold characteristics can exert an important control on fluid distribution. While vergence is argued to be a significant factor, loss of hydrocarbons by leakage along thrusts, normal faults, salt welds, and/or by hydraulic fracturing is clearly another important mechanism operating in the basin and is likely to have contributed to the underfilled nature of some traps.

5. Comparison of crestal normal faults in different fold settings

According to Stearns and Friedman (1972) conjugate and extensional fracture patterns form in three main orientations controlled by local stress orientations within folds (Type 1 = maximum horizontal stress sub-orthogonal to the fold axis; Types 2 (extensional) and 3 (compressional) = maximum horizontal stress sub-parallel to the fold axis). These patterns have been widely recognized in folds, although additional complexities can be identified (e.g. Stephenson et al., 2007), and fracture patterns tend to evolve through time and become superimposed, or reactivated with different sense of motion (e.g. Brita et al., 2006; Ahmadhadi et al., 2008). These fractures are important for fluid flow, and particularly in carbonates become partially or completely cemented with time.

Crestal normal faults are higher displacement features that tend to correspond with Type 2 conjugate fractures both in orientation and an outer arc location. However, crestal normal faults have more diverse origins than Type 2 fractures. Some, such as the convergent crestal normal faults of the Silak Anticline (Fig. 12) can be explained as arising from stresses related to folding (e.g. outer arc bending stresses, or a component of strike-slip motion; e.g. Ramsay, 1967; Strayer et al., 2004)). Other arrays of crestal normal faults are caused by collapse related to a salt or shale diapir, or, in the case of deepwater Brunei, as gravity-driven features related to sediment instability on the flanks of growing anticlines (see review in Morley, 2007a,b).

In one SE Asia deepwater fold and thrust belt crestal normal fault are modified polygonal faults. The polygonal geometry represents deformation in an isotropic stress field, probably in response to diagenetically-driven shear (Shin et al., 2008). Figure 18 shows that approaching the deformation front of the fold and thrust belt the polygonal faults respond to the stress field creating the folds and thrusts, and cease to be polygonal, grading into faults sub-orthogonal to the maximum horizontal stress direction. As folds propagate into this stress front they incorporate the strongly oriented normal faults of polygonal fault origin (Fig. 18). These faults lie at a high angle to the fold axis and impose permeability on the folds of very different orientation from typical normal faults that lie sub-parallel to the fold axis. They are also a novel stress orientation indicator.

![Cross-sections based on 2D seismic lines. A) Alborz Anticline. B) Detailed section of the Alborz Anticline showing wells and location of leaking hydrocarbons. C) Sarajeh Anticline. (Modified from Morley et al., 2013). See Fig. 11 for location.](image-url)
Crestal normal faults in Brunei and Iran have played an important role in the transmission of fluids. In Iran the crestal normal faults have a convergent conjugate geometry (i.e. the faults dip towards each other) and fault sets extend to 3 km deep (Fig. 12), while in Brunei the faults are divergent and tend to be confined to the upper 1 km. Consequently the potential for crestal faults in Iran to transport deeper fluids is greater than those in Brunei. The presence of oil, gas, Ca, Cl—rich overpressured fluids along the crestal faults of the Silak Anticline, is probably related to expulsion of basinal fluids from >4 km depth. These fluids could have leaked across breaches in the Upper Red evaporite seal either at welds associated with downbuilding structures or at deep normal faults (Fig. 12).

6. Discussion and conclusions

The general similarities and differences in structural development, overpressure generation and fluid type/migration between
the Brunei and Iran examples are summarized in Table 1. There are many different factors in the two areas that contribute to fluid distribution, type, overpressure distribution and interaction with structures, these include: smectite rich vs smectite poor clay layers (affecting depth of onset of temperature-controlled overpressure processes), water-rich vs water poor surface environment (affecting volume of trapped pore waters present, and available for expulsion during burial), multiple clay/shale seals vs single evaporite seal (affecting trapping of overpressure and overpressure magnitude), overpressure development in carbonates vs clastics (type of over-pressure generating mechanism), type of crestal normal fault (fluid pathways), detachment type, and structural style at depth.

In Brunei classic burial (mechanical-compaction) disequilibrium compaction and inflationary overpressures are important causes of overpressure in the smectite-poor upper 3–4 km of section. The basal detachment lies below about 150 °C, where chemical processes and metagenesis are inferred to drive overpressure development.

In the Central Basin of Iran the dense Upper Red Formation is a smectite-rich sequence. Below about 1900 m the smectite to illite transformation begins, causing moderate overpressures related to load transfer-disequilibrium compaction, plus local occurrences of inflationary overpressures. Near-lithostatic overpressures only occur below an evaporite seal in the carbonate-rich Qom Formation, and are probably related to a succession of overpressure processes during burial and different structural stages, the latest inferred to be tectonic-disequilibrium compaction and hydrocarbon generation/cracking volume-expansion.

Predominantly deepwater, clastic settings like offshore Brunei are very fluid rich resulting in numerous shallow water escape features during burial that include mud dykes, sills, laccoliths, volcanoes and pipes, and fluid escape pipes (polygonal faults and sand injectites occur in other deepwater settings). Deeper over-pressured fluids migrate along thrust faults until stress conditions trigger vertical pipe formation, typically in anticline crestal areas. This focus of fluids including thermogenic and biogenic gases commonly leads to the development of a shallow gas hydrate layer, which episodically traps gas.

The Central Basin of Iran is a water-poor environment and lacks well-developed fluid escape features. A comparison of the vertical stress gradients, for Brunei and for the Silak-1 well, Iran, show considerably higher, and more rapidly increasing vertical stress gradients in Iran (reflecting a much higher degree of shallower chemical diagenesis-related compaction in Iran compared with Brunei). In both Brunei and Iran, crestal normal faults have played an important role in the transmission of fluids towards the surface. The crestal normal faults in the anticlines of Central Iran penetrate deeper in the section (to about 3 km in the case of the Silak Anticline) than deepwater Brunei (generally present in the upper 1 km only). Consequently their potential for transporting deep fluids is greater. The presence of oil, gas, Ca- and Cl –rich overpressured fluids along the crestal faults of the Silak Anticline, is probably in part related to expulsion of basal fluids from >4 km depth.

The presence of thick evaporites at two levels in the Central Basin has exerted a considerable influence on fluid type and distribution in the basin by: 1) Providing a seal to highly over-pressured fluids in the Qom Formation, 2) influencing the pore water chemistry, 3) providing areas of fluid leakage where thrusts detach within the Upper Red Formation evaporites, and where downbuilding has occurred to create a weld, 4) downbuilding in the Lower Red Formation evaporites to create local depocentres where the Qom Formation is mature for gas, and 5) influencing the verge direction of folds (due to downbuilding), which in turn caused differences in timing of closure between the Alborz and Sarajeh Anticlines, which resulted in the former being oil filled, and the latter gas filled.

Overall the deepwater Brunei system is very water rich, and multiple opportunities for overpressure generation and fluid leakage have occurred throughout the growth of the anticlines. The result is a wide variety of fluid migration pathways and structures from deep to shallow levels, and widespread inflationary-type overpressure. In the Central Basin the near surface environment is water limited. Mechanical and chemical compaction led to moderate overpressure development above the Upper Red Formation evaporites, it is the trapping of fluids below the evaporites where the large overpressures have developed. Fluid leakage episodes across the evaporites have either been very few or absent in most areas. Locations where episodes of leakage can occur (e.g. detaching thrusts, deep normal faults, salt welds) are sparse.


