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REVIEW

Origin mechanism of overpressure in saline lacustrine formation of the Paleogene and Neogene in the Western Qaidam Basin, NW China

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Abstract
Previous research shows that the origin and distribution of formation overpressure are governed by several dominant factors including disequilibrium compaction, hydrocarbon generation, tectonic compression and diagenesis, influenced by salt components and their concentration, but it is unclear how salts affect formation overpressure in saline lacustrine basins. This paper investigated the effect of salts on formation overpressure based on organic geochemistry, rock mineralogy, logging curve comparison and wave velocity-density cross-plot by combining the sedimentary and structural background. In the Western Qaidam Basin, the proportion of abnormally high pressures rises from the Neogene to the Paleogene. The top surface of the overpressure is between 2300 and 2500 m deep. As the subsidence and sedimentary centers of the basin moved eastward, the centers of the overpressure migrated from west to east. In the Upper Oligocene, the overpressure is developed in the deep and semi-deep lacustrine facies, and the pressure coefficient is 1.8–2.0. Disequilibrium compaction is the primary control factor with a contribution rate of more than 60% in the intersalt and subsalt strata, followed by tectonic compression with a contribution rate of 20–30%. Fracture reducing by salt filling and fluid volume expanding by gypsum dehydration increase the fluid volume in the formation, which promotes formation overpressure. The gypsum salt rocks also have strong plasticity and sealing effect, thus providing a closed environment for the formation overpressure. Through providing the primary migration driven force and sealing conditions for oil and gas, the overpressure is meaningful to petroleum accumulation and preservation in saline lacustrine formation of the Paleogene and Neogene in the Western Qaidam Basin, NW China.

Keywords Saline lacustrine formation · Overpressure · Qaidam Basin · Paleogene · Origin mechanism

Introduction

Formation pressure refers to the pressure of the pore fluid in the formation, which is equal to the pressure generated by the hydrostatic column. The pressure caused by the weight of the matrix and the pore fluid of overlying sediments or strata is called the static rock pressure. Generally, the formation pressure coefficient or the formation pressure gradient is used to describe the formation pressure state of sedimentary basins. The pressure coefficient is the ratio of formation pressure to hydrostatic pressure at a certain depth underground. Vertically, formation pressure can be divided into negative pressure (pressure coefficient less than 0.9), normal pressure (pressure coefficient being 1–1.3), overpressure (pressure coefficient being 1.3–1.73) and strong overpressure (pressure coefficient more than 1.73).The origin and distribution of formation overpressure are determined by multiple factors, such as disequilibrium compaction, organic hydrocarbon generation, tectonic compression and diagenesis (Dickinson 1953; Lewis and Rose 1970; Barker 1972; Luo and Vasseur 1992; Hunt et al. 1994; Yu and Lerche 1996; Osborne and Swarbrick 1997; Goldsmith 1998; Swarbrick and Schneider 1999; Zhang and Dong 2000; Gao et al. 2005; Zhang et al. 2013; Jiang et al. 2016; Zhao et al. 2017).

Among over 180 overpressured basins in the globe, more than 160 are abundant in hydrocarbons (Hunt 1990; Du et al. 1995; Hao and Dong 2001). Chlorides and their

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concentrations affect organic matter hydrocarbon generation, sediment diagenesis, and caprock sealing, which are related to the origin of formation overpressure (Melvin 1991; Qi and Huang 1992; Chen et al. 2003; Warren 2006, 2016; Jin et al. 2008; Liu et al. 2013, 2014, 2016, 2019). It is inferred that there is a certain internal relationship between salts and formation overpressure, which in turn affects the origin mechanism and spatial distribution of formation overpressure.

Salts have some impact on the formation disequilibrium compaction. Compared with others, the strata with higher salinity are often denser, and fluids cannot be discharged in time under the action of the overburden, thus cause stronger disequilibrium compaction, and ultimately lead to the formation overpressure to greater degree (Osborne and Swarbrick 1997; Chen et al. 2003; Liu et al. 2019).

Salts also have a significant impact on diagenesis. High salinity environments are rich in carbonate, sulfate or chloride minerals. The increase of salt content leads to the strengthening of rock cementation, thus blocking the pores of the reservoir rock to ultimately form an overpressure sealing layer. The dehydration temperature of calcification promotes the dehydration and conversion process of montmorillonite, which increases the volume of fluid in the formation and saline lacustrine formation overpressure (Wang and Ouyang 1993; Liu et al. 2019).

A high plastic salt rock has special rock mechanical properties (Jiang et al. 2012). During the deformation process, it mainly absorbs stress in the form of folds, while fractures are less developed. After being subjected to extremely strong structural compression, the sealing of the caprock can still be maintained (Hao and Dong 2001; Zhuo et al. 2014), which is conducive to the formation overpressure. In addition, under extreme stress, a salt rock is also prone to diapir puncture, which alters the local geothermal gradient, organic matter pyrolysis rate, and hydrocarbon generation amount, all of which have impacts on the formation pressure state (Tang et al. 2007; Luo et al. 2004; Liu et al. 2019; Wang et al. 2023).

The objectives of this study are to identify the regional distributions of formation pressure, explore the correlation between salinity and pressure in various strata; probe the effect mechanism of salts on formation overpressure, including fluid volume change during the process of hydrocarbon generation and diagenesis influenced by salts, compression stress change during disequilibrium compaction and tectonic compression by salts; establish quantitative or semi-quantitative models to assess the genetic contribution of overpressure by salts and their concentration in the Western Qaidam Basin.

Geologic setting

The Qaidam Basin is an intermountain basin situated in the northeastern Qinghai–Tibet Plateau with a total area of 12×10^4 km^2, and is one of the seven inland lake basins in China. The East Kunlun orogenic belt borders the basin in the southwest, the northeast is the Qilian orogenic belt, and the northwest is separated from the Tarim Basin by the Altun mountain. The Qaidam Basin is composed of three geological units: West Depression, North Margin Depression and the Sanhu Depression. The West Depression consists of four secondary structural units: Kunbei fault stage, Mangya Depression, Dafengshan Uplift, and Yiliping Depression (Fig. 1).

The Qaidam Basin has gone through various periods of tectonic upheaval during the Cenozoic Era. The compression and collision brought on by the ongoing migration of the Indian Plate toward the Eurasian Plate have impacts on the tectonic deformation of the basin, closely related to the activities of the surrounding mountains and regional faults. The sedimentary environments were saline lakes and the paleoclimate turned drier and colder from the Paleogene to Neogene in the Western Qaidam Basin. The sedimentary center gradually migrates eastward from the Lulehe Formation of Paleocene–Paleocene (E_{1+2}), Lower Member of Lower Ganchaigou Formation of Oligocene (E_3^1), Upper Member of Lower Ganchaigou Formation of Oligocene (E_3^2), Upper Ganchaigou Formation of Miocene (N_1), Lower Youshashan Formation of Pliocene (N_2^1), Upper Youshashan Formation of Pliocene (N_2^2) to Shizigou Formation of Pliocene (N_2^3). Distinct from the inland freshwater lakes, high salinity, frequent mobility of the salinization centers, and the presence of different salts are characteristics of the Western Qaidam Basin, where the saline lakes experienced three evolution stages of origin, development, and extinction during the Paleogene and Neogene and the main salt intervals are in the E_3^2 and N_1 (Liu et al. 2013, 2014, 2016, 2019).

Samples and analytical methods

55 drilled core samples including mudstones, calcareous mudstones, and silty mudstones were collected from the E_{1+2} to N_2^3 in the Western Qaidam Basin. TOC content, extractable organic matter content, rock–eval pyrolysis and chloride ion (Cl^-) content were conducted.

Dry combustion was adopted for testing TOC contents. After crushed and washed by the hydrochloric acid solution to eliminate the interference of carbonates, the samples were heated to 900 ℃ with an oxygen-rich carrier

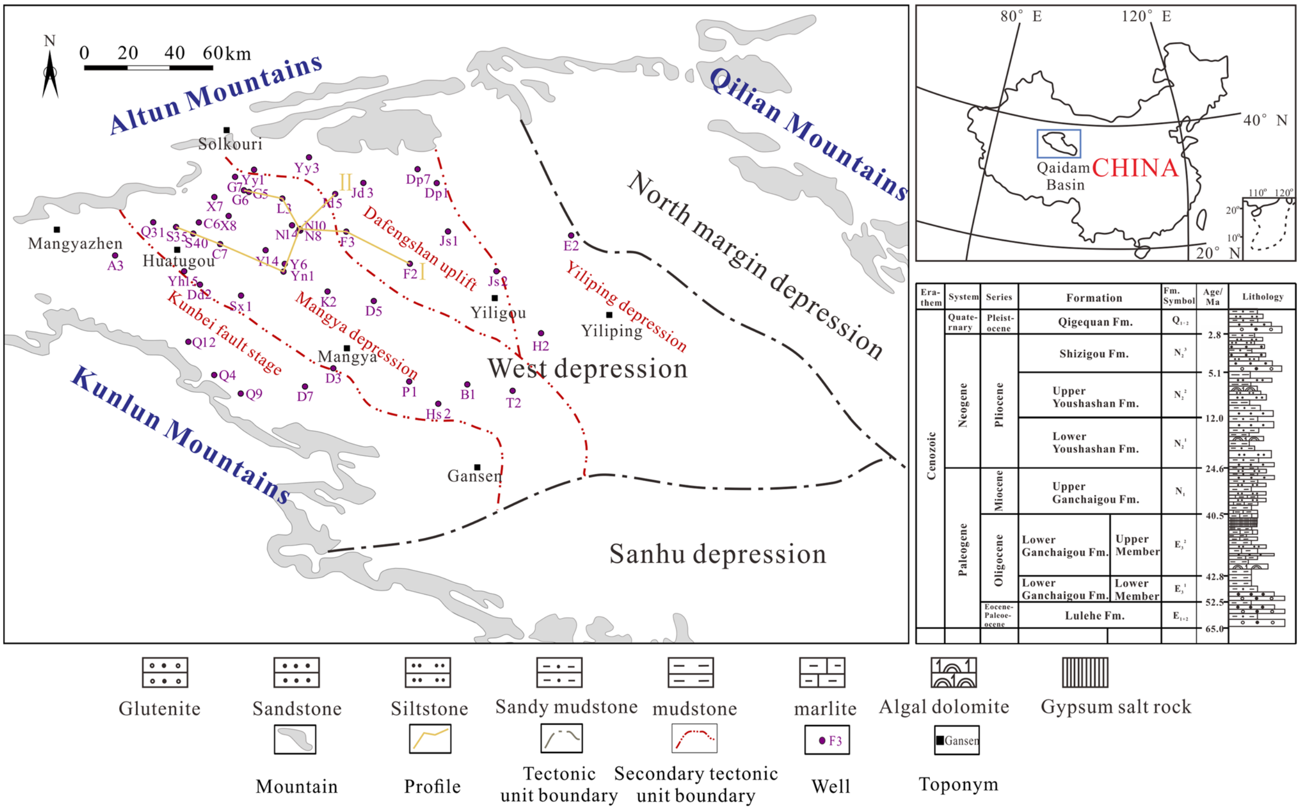

Fig. 1 Structural unit divisions, formation divisions, and lithology in the Western Qaidam Basin

gas, which fully oxidized the samples' organic carbon into carbon dioxide. TOC contents were calculated based on the volume of generated carbon dioxide. These tests followed the State Standard of the People's Republic of China, Determination of Total Organic Carbon Contents in Sedimentary Rock (GB/T 19145–2003).

Extractable organic matter content was tested by the Soxhlet extraction method. Ground source rock samples were extracted by chloroform in a Soxhlet extractor and kept in a water bath (80 ℃) for 48 h. Extractable organic matter was then obtained after rotary evaporation and nitrogen gas drying. It was then further divided into four fractions including saturates, aromatics, non-hydrocarbons and asphaltenes by eluting n-hexane, dichloromethane, and chloroform-ethanol. The tests followed the State Industry Standard of the People's Republic of China, Determination of Bitumen from Rocks by Chloroform Extraction (SY/T 5118–2005).

Total hydrocarbon content and Tmax were obtained using rock–eval pyrolysis. The samples were heated to 300 ℃ for 3 min and thermally evaporated to determine the free hydrocarbon (S_1). The kerogen thermal cracking hydrocarbon (S_2) was measured at a temperature range from 300 to 600 ℃ and the corresponding peak temperature was Tmax. The tests followed the State Standard of the People's Republic of China, Rock Pyrolysis Analysis (GB/T 18602–2012).

Silver nitrate titration was used to test Cl^- content. The two main test procedures included titrating the powders solution with a standard solution of silver nitrate and diluting sample powders with distilled water. When the solution turned brick-red, the titration endpoint was reached. Potassium chromate was utilized as an indicator. The Cl^- content was determined according to the consumption of the silver nitrate solution. The tests followed the State Industry Standard of the People's Republic of China, Salt Mineral Exploration-Part 4: Deep Brine (DZ/T 0212.4–2020).

The distribution and evolution characteristics of salinity in the Western Qaidam Basin were constructed by restoring the paleosalinity with Cl^- content. Cl^- content was widely used to restore seawater salinity, and the specific relationship between them was deduced as

$$S\text{ ‰ (salinity)} = 1.80655\ Cl^{-}\text{‰ (Lyman 1969)}$$

Boron content was used to adjust the salinity calculation results from Cl^- content (Liu et al. 2018).

The burial process, thermal evolution process and oil and gas generation process of the Paleogene and Neogene have been recovered. In order to obtain the sedimentation and deposition rate, the reverse denudation method by Allen and Allen (2013) was applied to restore the burial process, where the thickness and timing of the denudation referred to Feng et al. (2021). The restoration of paleotemperature was realized by EASY %Ro model established by Sweeney and Burnham (1990). The geothermal gradient values were fitted until the simulated and measured Ro values were in agreement, at which point the thermal history was rebuilt. Chemical kinetics techniques were used to reconstruct hydrocarbon generation process (Tissot and Welte 1984).

In order to fully characterize the vertically and horizontally distributional characteristics of formation pressure in the Western Qaidam Basin, the equivalent depth method based on acoustic logging data (Magara 1968; Zhao et al. 2001) was used to predict the formation pressure of more than 30 typical wells.

Results

Distribution characteristics of formation pressure

In the Western Qaidam Basin, the formation pressure is from 3 to 15 MPa, which is lower than the hydrostatic pressure, for shallower strata with a depth less than 1300 m. The pressure coefficient for these shallower layers is 0.5–0.9, thus belonging to the negative pressure system. For the middle layers with a buried depth of 1300–2500 m, the formation pressure is 15–30 MPa, corresponding to a distribution near the hydrostatic pressure. The pressure coefficient at these middle layers is between 0.9 and 1.27, thus belonging to the normal pressure system. The formation pressure for the strata deeper than 2500 m is 30–80 MPa, which is higher than the hydrostatic pressure. The pressure coefficient at these deeper layers is generally greater than 1.3, and some even greater than 1.73, 62 percent of measured points belonging to overpressure and strong overpressure systems. Obviously, the formation pressure in the shallow and middle intervals increases with the burial depth. However, the correlation between the pressure and the buried depth gradually becomes worse for the middle and deep layers (Fig. 2).

In terms of the stratum distribution, most formation pressure data of the N_2^1 and the N_1 fall within the normal pressure system. The E_3^2 is overpressured roughly from 2300 m. The measured pressure coefficient of the E_3^2 begins to exceed 1.73, corresponding to the depth of 3750 m. The greatest observed pressure coefficient is 2.23 at a depth of 3840 m of Well S40. The formation pressure coefficient of the E_3^1 is totally higher than that of the E_3^2. Some pressure coefficient data points of more than 1.73 appear around 3300 m, and there are more data points with strong overpressure and greater depth for the E_3^1.

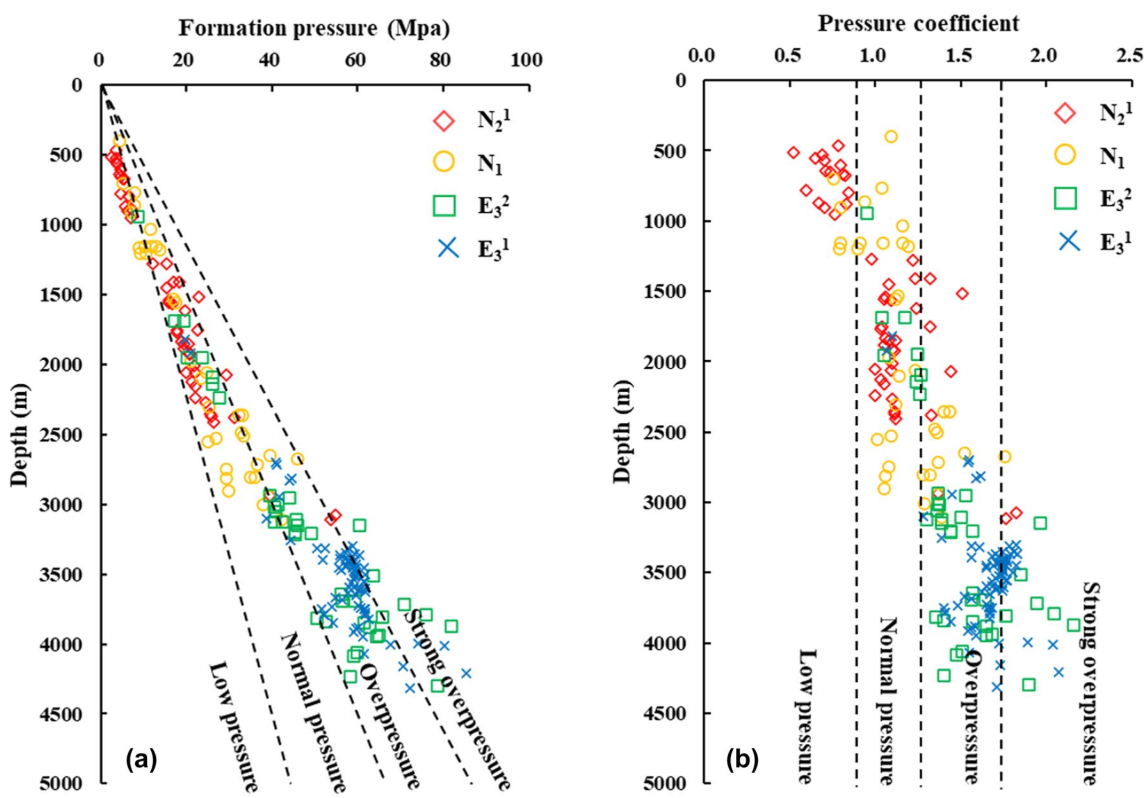

Fig. 2 Measured formation pressure vertical distribution in the Western Qaidam Basin: **a** formation pressure; **b** formation pressure coefficient

The NW–SE profile spans from the Honggouzi structure through the Dafengshan structure, including Well G6, Well L3, Well N10, Well F3 and Well F2. The Honggouzi and Xiaoliangshan structures are close to the basin edge, where the formation pressure coefficient is totally low because of poor preservation conditions. The overpressure occurs in deeper strata, including the E_3^2 and N_1, where the pressure coefficient is slightly greater than 1.3. The overpressure top of Well G6 and Well L3 is respectively 2780 m and 3300 m. The pressure coefficient in the Nanyishan structure is relatively large, and the maximum pressure coefficient in the E_3^2 can reach 1.75. The overpressure of Well F3 begins to appear from the N_2^1, the depth of the overpressure top is about 2200 m, and the overall pressure coefficient is 1.3–1.5. The lower stratum of Well F2 develops overpressure, with the overpressure top of about 2450 m and the pressure coefficient of 1.35–1.7 (Fig. 3).

The pressure coefficients are generally higher in the west while lower in the east of the Western Qaidam Basin, similar to the distribution characteristics of Cl^- content, indicating that the salinity has effects on the formation pressure (Liu et al. 2018; Li et al. 2021).

The E_3^2: The value of Cl^- content indicates that the lakes in most areas were freshwater and brackish water. The super salinization center was in the west of Yingxiong ridge where thin gypsum salt rock and mudstone were deposited. The average value of formation pressure coefficient in the E_3^2 ranges from 1.0 to 1.9, and the overpressure with a pressure coefficient more than 1.27 has the largest development range except for the edge of the basin. It is determined that the western Yingxiong ridge and Xianshuiquan areas are strong overpressure centers, where the overpressure coefficient can reach more than 2.0. The amplitude of overpressure decreases gradually from the center to the periphery (Fig. 4).

The N_1: most of the areas were composed of freshwater and brackish water environments. The high salinity centers included part of the western Yingxiong ridge and Honggouzi areas, but the scope reduced compared to the E_3^2. The salinization center shifted eastward, and the supersaline water environment no longer existed. The pressure coefficient of the N_1 is, respectively, greater in the north and south. The average pressure coefficient ranges from 1.0 to 1.7, which is lower than that of the E_3^2. The center of strong overpressure disappears and the range of overpressure decreases in the western Yingxiong ridge. There are two pressure centers, one is located at the Youquanzi structure and its vicinity, and the other near the Eboshan structure on the southeastern edge of the Western Qaidam Basin, where the highest pressure coefficient is in the center but gradually decreases from the center to the surroundings.

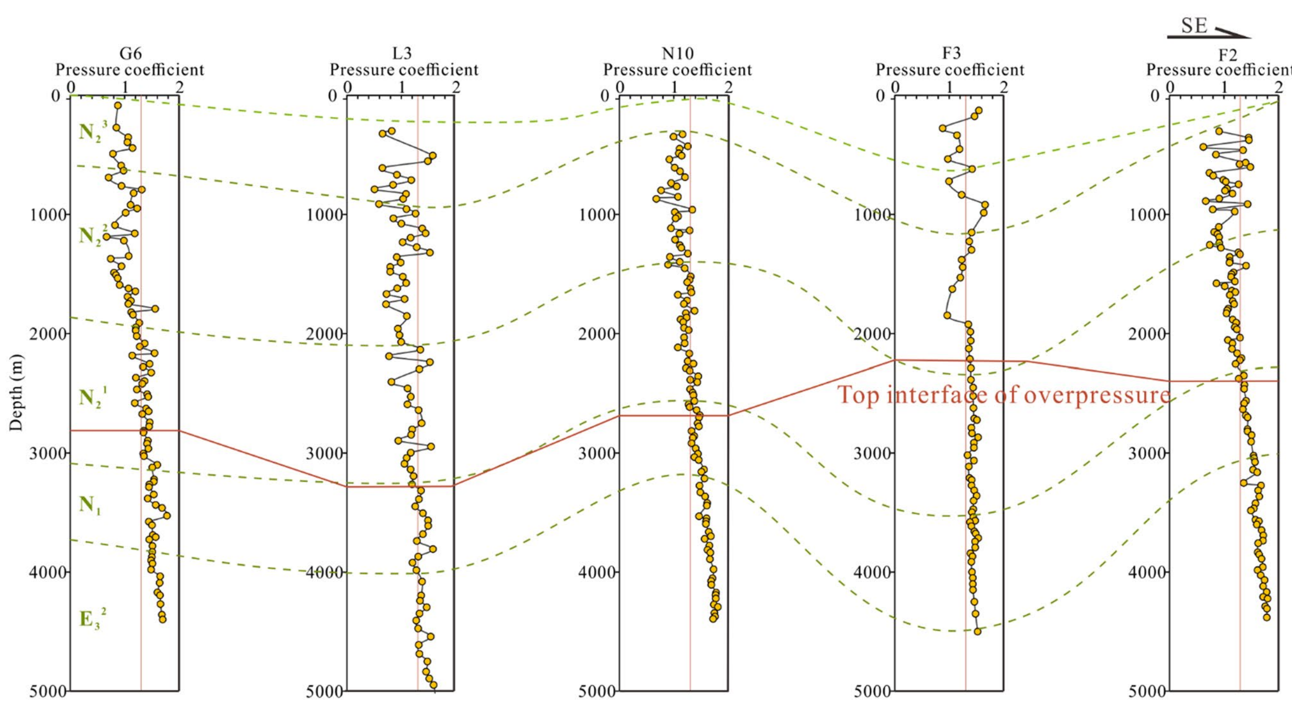

Fig. 3 Cross-well formation pressure coefficient profiles in the Western Qaidam Basin. The positions of profiles are shown in Fig. 1c

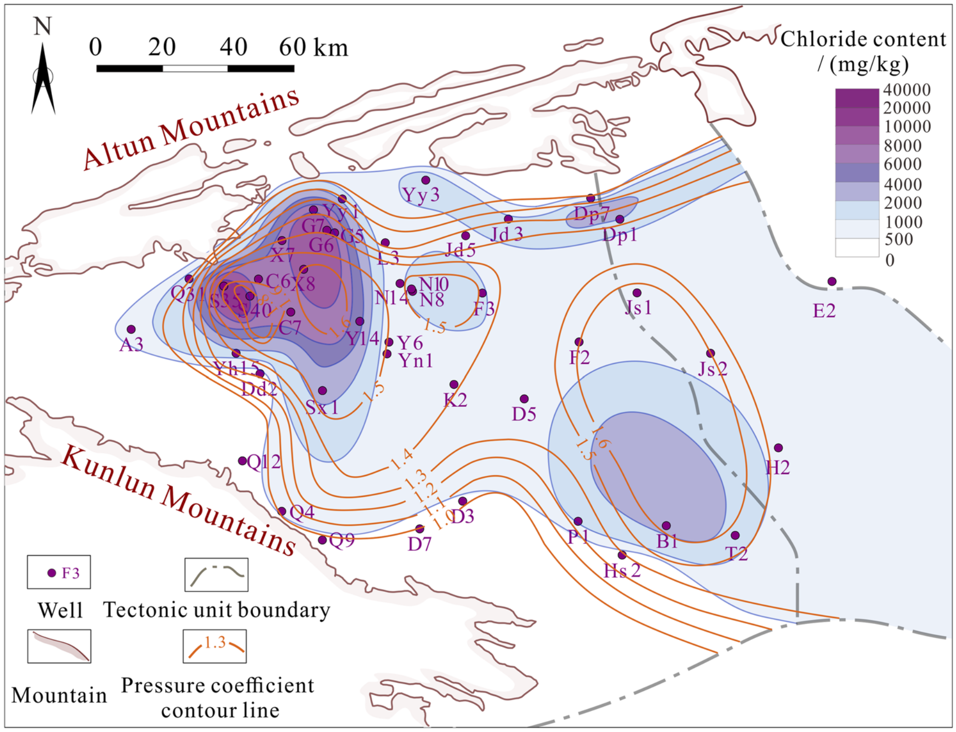

Fig. 4 Overlayed contours of formation pressure coefficient and salinity of the E_3^2 in the Western Qaidam Basin

Different elements' effects on formation pressurization

Disequilibrium compaction

Disequilibrium compaction has a significant impact on the generation of overpressure in the western Yingxiong ridge area, according to the identification of the overpressure mechanism above. In fact, the overpressure of a large number of Cenozoic young sedimentary basins with high sedimentation rate and fine lithology is resulted from the disequilibrium compaction of mudstone, which is largely one of the main mechanisms of overpressure (Zahra and Erfan 2020; Li et al. 2020). In the process of disequilibrium compaction, the rapid burying and compaction of the sedimentary strata result in an disequilibrium relationship between the sedimentation rate and drainage rate. The fluid in the pores of the strata is not discharged smoothly, which bears part of the continuous increase of overlying strata pressure, forming the overpressure by disequilibrium compaction. In the western Yingxiong ridge area, during the deposition period of the E_3^1 to N_2^2, the sedimentation rate was more than 60 m/ma; during the deposition period of the E_3^2, the western Yingxiong ridge area was one of the sedimentary centers of the Western Qaidam Basin, with the sedimentation rate of 400–700 m/ma, showing rapid subsidence and sedimentation. The rapid accumulation and loading of sedimentary layers are favorable for the disequilibrium compaction pressurization.

The gypsum salt rocks in the Western Qaidam Basin were mainly formed in the salinization environment of the Oligocene (E_3^1 and E_3^2) of the western Yingxiong ridge area. During this geohistory, the lakes expanded to the largest area. The sedimentary facies were mainly semi-deep and deep lakes, with the development of unique lacustrine carbonate rocks and gypsum salt rocks, thinly interbedded with fine-grained sandstones and mudstones. Combined with the previous large-scale flume sedimentation simulation experiments (Gou et al. 2014; Huang et al. 2015), it is preliminarily confirmed that the salinization basins in the arid and evaporating paleoclimate environment are more likely to develop rich salt minerals. Under the control of salinized water mediums, the fine clastic particles and argillaceous components are thinly interbedded, and the distribution range of fine sediments is also more extensive. On the one hand, the western Yingxiong ridge was situated in the sedimentary center of the basin during the Oligocene (E_3^1 and E_3^2), with a fast sedimentation rate and a large total thickness of sediments. Under the action of overlying load, the compressive stress decreased the pore volume, and the pore fluid bore part of the load, which also prevented the further compaction of the sediments. On the other hand, the fine-grained sediments, especially the gypsum salt rocks, formed tight cap rocks. The fluid could not be discharged in time under the overlying load, and leaded to the disequilibrium compaction of the interbedded sandstones and mudstones, resulting in the overpressure under the cap rocks.

Hydrocarbon generation

The effect of organic matter's hydrocarbon generation on the development of overpressure is that the solid kerogen with higher density is transformed into liquid hydrocarbon and gaseous hydrocarbon with lower molecular weight, forming new pore fluid with lower density and making the volume expansion rate of original pore fluid generally up to 25% (Hunt 1990).

The organic matter abundance corresponding to medium salinity is higher, while the organic matter enrichment is limited in the super salinity stage. In the salinized environment, the opportune salinity is conducive to the preservation of soluble organics (Wang et al. 2009; Liu et al. 2016). The higher the salinity within a particular range is, the better the type of organic matter. Most studies indicate that the existence of salts accelerates the generation of hydrocarbon and degradation of source rocks, and the hydrocarbon creation of source rocks is ensured by the catalytic influence of salts on the production of organic carbon (Zhang et al. 2018).

The E_3^2 and N_1 sedimentary period in the west of the Western Qaidam Basin was dominated by shore-shallow and deep lacustrine facies, forming two sets of hydrocarbon source rock-thick mudstones. According to the Well Y8 in the Youquanzi structure, the TOC contents of hydrocarbon source rocks increase from 0.7% to 1.8% and then decrease to 0.4% from the formation bottom to top. The TOC contents are higher in moderate saline water than in fresh and hypersaline water. The extractable organic matter contents are 0.1–0.2% with the similar trend of salinity to the TOC contents. The hydrocarbon source rocks are identified as medium through good source rock with higher hydrocarbon generation potential. The yield index of $S_1/(S_1+S_2)$ of organic matter reflects the degree of evolution reaching the mature and high mature stages (Fig. 5). The newly generated fluid cannot be discharged smoothly due to limited connectivity between source rocks and reservoirs, forming overpressure.

The intersalt formations are not the main hydrocarbon generating layers, and the pressure increase caused by oil and gas generation is not too large. The intersalt interval of Well S23 in the western Yingxiong ridge is 3000–4200 m, corresponding to the period of supersalinization, with the TOC contents being about 0.4%, less than 0.8%, and the maturity from 0.6% to 1.2%, and limited hydrocarbon generation. Therefore, intersalt hydrocarbon generation has a negligible impact on overpressure origin, and the formation pressure is lower than that of subsalt formation.

Tectonic compression

As one of the causes of overpressure, tectonic compression is controlled by stress in three directions, equivalent to three-dimensional compaction. When the stress is large enough, the formation is broken and some pressure is released. The Western Qaidam Basin was under the background of tectonic compression by the surrounding mountains during the Paleogene and Neogene. The compressive stress in the horizontal direction is greater than that in the vertical direction due to local or regional folding, fault and fault block activity, diapiric salt and mud dome movement in tectonic activity, which leads to the decrease in pore volume. The formation

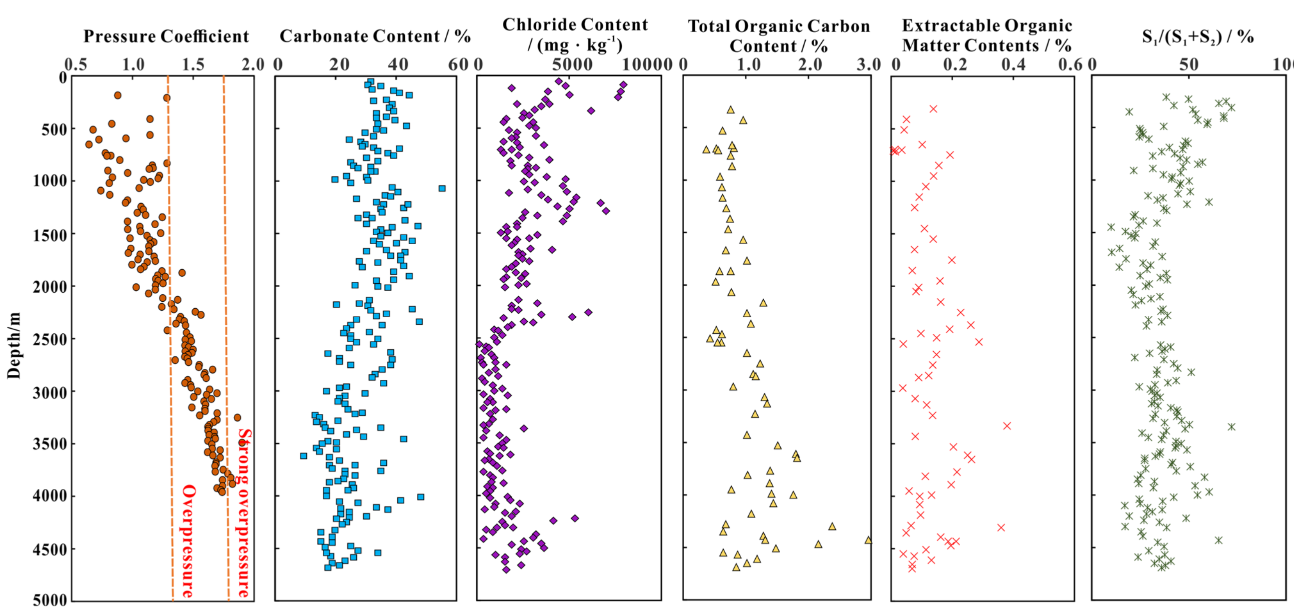

Fig. 5 Comparison of pressure coefficient, salinity index and organic matter characteristics of Well Y8

overpressure caused by structural compression occurs in the mudstone or adjacent sandstone or carbonate rock. After the tectonic movement stops, the overpressure disappears rapidly. The existence of an overpressure system caused by tectonic action is conditional (Muggeridge et al. 2005).

Different from clastic rocks, gypsum salt rocks tend to brittle deformation at low temperature and plastic deformation at high temperature. Gypsum salt rocks can reach very strong plasticity and fluidity at a depth of about 3000 m. The mechanical properties of salt rocks are more special than gypsum rocks. Under the same stress conditions at the same temperature, the salt rocks deform earlier. In the E_3^2 in the Western Qaidam Basin, there are various salt minerals, and some of them are accumulated to form gypsum rocks. With the increase of the proportion of Gypsum salts, the plastic creep effect increases when the stress is applied (Bian et al. 2020). The abnormal high pressure is developed in the western Yingxiong ridge area from 2500 m, and the temperature is more than 100 ℃ from 2800 m (Fig. 6). The proportion of evaporating salt rock in the upper part of the salt layer is relatively large, so it has a plastic rheological property. Under the influence of the peripheral orogenic belt and fault activity, the tectonic stress field has been left lateral compression and torsion since the end of Paleogene. Both gravity load and structural load drive salt rocks to flow to fault belts, where low formation pressure develops. Consequently, salt nappe structures are formed and the original fractures and faults can be filled. The lower part of the salt layer is relatively rich in mud, and the gypsum layer is vulnerable to brittle deformation. The faults and fractures stop development in the salt layers or are sealed up. The continuous pressure boosting and pressure relief under the action of various genetic mechanisms are dynamic equilibrium processes. The optimal sealing conditions formed by the plasticity and fluidity of the gypsum salt rock play a static sealing role in the formation and distribution of overpressure system.

Diagenesis

The influence of diagenesis on formation overpressure mainly includes dehydration of montmorillonite into illite, dehydration of gypsum into anhydrite, and pore and fracture filling with salt minerals. Salt rock has good thermal conductivity, which makes the heat flow concentrated in the mudstone section, resulting in the high temperature of mudstone. The increase of salinity also reduces the dehydration temperature of montmorillonite, making it easy for montmorillonite to transform and dehydrate. Therefore, the increase of fluid volume in the formation promotes the origin of overpressure (Wang and Ouyang 1993). As temperature and depth increase, a large amount of free water is released in the process of gypsum transforming into anhydrite without crystal water. Free water continues to accumulate in the surrounding rock pores, which raises the formation pore fluid pressure.

Under the control of saline water, gypsum rock, salt rock and mudstone are thinly interbedded in the western Yingxiong ridge area. The E_3^2 lithology is dominated by lacustrine micritic carbonate rocks with mixed genesis. After the formation was deposited, tectonic stress, the brittleness of the limestone, and fluid pressure all contributed to the development of many fractures. A high salinity environment was rich in evaporate minerals such as gypsum, halite and glauberite. The semi-consolidated salt minerals flowed and reprecipitated in the later evaporation deposition process, thus plugging some fractures and early dissolution pores.

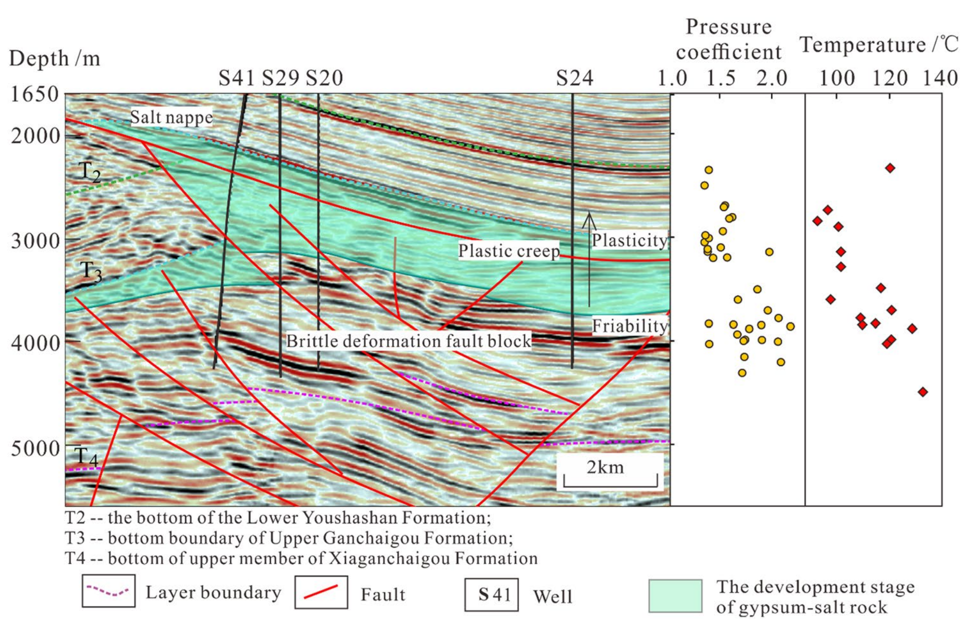

Fig. 6 Seismicity interpretation, vertical measured temperature and pressure in the section of Wells S41, S29, S20 and S24

The salt water with high density in the salinized lake was easy to infiltrate and cement in the process of sedimentation, resulting in the filling and plugging of the underlying reservoir pores. Finally, a high-pressure sealing layer was formed. Field outcrops and cores show gypsum salt rock widely covered in the upper E_3^2. Most fractures with a width of about 0.05–0.50 mm are also observed in the rock slices and are mostly filled with calcite and evaporite minerals. The formation is also rich in salt minerals and clay minerals (Fig. 7). Therefore, compared with the subsalt strata, there are more gypsum salt rocks in the intersalts, and diagenesis also contributes to the origin of overpressure.

Formation overpressure origin mechanism

Analyses of logging curves

Well Y8 is located in the Youquanzi structure. There is no gypsum salt rock in the E_3^2. The combination characteristics of acoustic, resistivity and density logging curves show that the curves are offset at about 1800 m. The acoustic values of 1800–2600 m interval in Well Y8 deviate from the normal trend line and the resistivity of hydrocarbon-bearing layer increases, while the variation trend of density logging parameters reflecting the volume properties of rock almost remains unchanged. However, the acoustic velocity between 2600 and 4000 m increases continuously or remains constant, and the resistivity and density increase slightly while deviating from the normal trend. It can be preliminarily identified that there is almost no disequilibrium compaction between 1800 and 2600 m, and fluid expansion or tectonic compression may occur. Disequilibrium compaction occurs between 2600 and 4000 m, and multiple overpressure composite sources such as fluid expansion or tectonic compression may exist at the same time.

Analyses of acoustic velocity-density cross-plot

The acoustic velocity–density cross-plot method is based on the study of the mechanical relationship in the formation compaction process, which can be divided into two types: loading curve relationship and unloading curve relationship (Bowers 1995; Ye et al. 2012). On the acoustic velocity–density cross-plot, the effective vertical stress increases while porosity decreases on the loading curve. The origin mechanism mainly includes differential compaction and tectonic compression. In the unloading curve relationship, the effective vertical stress decreases while the porosity increases, thus leading to the rise of pore pressure or the falling of overlying pressure. The origin mechanism mainly includes fluid expansion or overpressure conduction (Fig. 8a).

(a) Field outcrop, Ganchaigou section, gypsum salt rock

(b) Well S38-4, 3,739.62 m, micritic limestone

(c) Well S20, 4195m, argillaceous limestone, with well-developed fractures, most of which are filled with calcite and anhydrite

(d) Well S20, 4195m, argillaceous limestone, with well-developed dissolution pores, most of which are filled with calcite and anhydrite, and some of which are incomplete filled

(e) Well S41-2, 4089.41m, plaster mudstone, scanning electron microscope, gypsum

(f) Well S41-2, 4181.16m, silty mudstone, SEM, flaky clay mineral-coated quartz particles

Fig. 7 Characteristics of salt formation and salt minerals of the E_3 in the Western Qaidam Basin

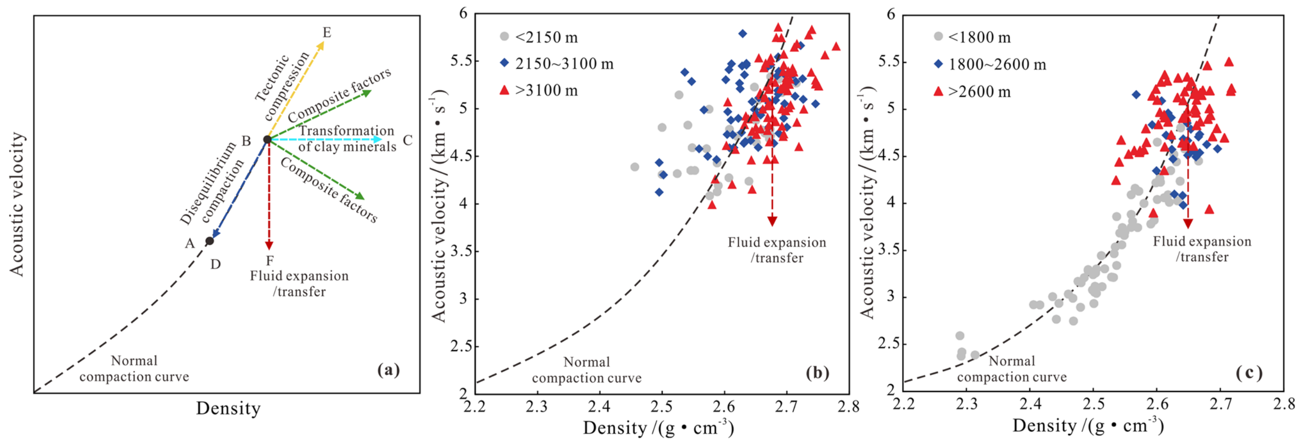

Fig. 8 Acoustic velocity–density cross-plot of mudstone of Well Y8. **a** Recognition model of main overpressure origin mechanism; **b** velocity–density cross-plot of Well 8; **c** velocity–density cross-plot of Well 8

From the acoustic velocity–density cross-plot of mudstone in Well Y8, it can be seen that 0–2150 m is the normal compaction section; 2150–3100 m is the overpressure section where the points of acoustic velocity and density mainly fall on the loading curve within the depth range of the normal compaction trend section. The variation trend of the loading curve is similar to that of the normal section, and deviates slightly from the loading curve to the direction of low acoustic velocity and density. Since the gypsum salt rock is reasonably developed, it has been determined that disequilibrium compaction is the primary cause of the overpressure and that hydrocarbon production pressurization is not a major factor. The wave velocities and density dispersion points of 3100–4500 m of overpressure fall on the loading curve, indicating that the overpressure is primarily attribute to disequilibrium compaction. The overpressure formed by tectonic compression falls on the normal loading curve corresponding to the depth of the overpressure section and deviates to the direction of high acoustic velocity and density. There are few points with this feature in this overpressure section, indicating that the possibility of tectonic compression pressurization is little to almost nonexistent (Fig. 8b).

In contrast, the observation and analysis of Well Y8 in the Youquanzi structure where gypsum salt rock is not developed shows that: the normal compaction section is 0–1800 m; the acoustic velocity and density points of 1800–2600 m fall on the unloading curve. The density data of this section fluctuate little, but the acoustic velocity decreases obviously. The points of the overpressure section conform to hydrocarbon generation pressurization on the unloading curve. The points of acoustic velocity and density between 2600 and 4000 m vary widely. Although the curve composed of points does not completely coincide with a certain overpressure mechanism identification curve, it is still distributed around the curve, indicating that formation overpressure is mainly due to disequilibrium compaction, partly to tectonic compression and superimposed diagenesis (Fig. 8c).

Calculation of contribution rate of various origin mechanisms

In consideration of former workers' perspectives (Hua and Lin 1983; Guo et al. 2005; Liu et al. 2019), disequilibrium compaction of sediments, hydrocarbon generation from organic matter, and tectonic compression are thought as the main reasons of formation overpressure in the Western Qaidam Basin.

For the Western Qaidam Basin, the sedimentation rate of the E_3^2 is 5–6 times that of other strata. The higher sedimentation rate has an influence on the generation of overpressure, and the formation is also the main layer of overpressure distribution. The E_3^2 lithology is mainly composed of thick grey and dark grey mudstone. The impermeable mudstone is responsible for the origin of disequilibrium compaction, which is also very favorable for the preservation of overpressure.

The contribution rate formula of disequilibrium compaction (Luo 2004) is as follows:

$$\Delta P_1 = P_f - P_w = (Go - Gw)[Z - (Ln \triangle t_o - Ln\Delta t)/c]$$

$$G_1 = \Delta P_1/\Delta P$$

where $\triangle P_1$ is the disequilibrium compaction pressurization, at; P_f is the formation pore fluid pressure, at; P_w is hydrostatic pressure, at; G_o is the formation lithostatic pressure gradient, kg/cm^2/m and $Go = 9.14*10^{-6}Z + 0.224$ in the Western Qaidam Basin; G_w is the formation static water pressure gradient, kg/cm^2/m and Gw = $1.73*10^{-6}$Z + 0.105 in the Western Qaidam Basin; Z is buried depth, m; Δt_o is the surface acoustic transit time, μs/ft; Δt is the target stratum acoustic transit time, μs/ft; c is the constant related to rock properties; G_1 is disequilibrium compaction pressurization contribution rate; ΔP is excess pressure, at.

Using the method of equivalent depth, we calculated the contribution rate of disequilibrium compaction to overpressure in the Western Qaidam Basin. The results illustrate that the average contribution rate of disequilibrium compaction can reach 60% (Table 1).

The analyses of the acoustic transit time curve of mudstone in Well N10 of the Nanyishan structure and formation pressure curve in Well Y14 of the Youquanzi structure show that disequilibrium compaction is the main cause of overpressure in the Western Qaidam Basin.

As a mechanism of overpressure, tectonic stress generally occurs in areas with strong tectonic compression, which affects the genesis of anomalous fluid pressure, mechanical seepage characteristics and rock properties. In a completely closed system, the pressure increment caused by tectonic stress is as follows (Luo 2004):

$$\Delta P_2 = \sigma_1 - S$$

$$G_2 = \Delta P_2 / \Delta P$$

where ΔP_2 is the tectonic compression pressure increment; σ_1 is the maximum principal stress in the lateral direction; S is the vertical overlying load; G_2 is the tectonic compression pressure contribution rate; ΔP is the total pressure increment.

If the formation is well sealed, tectonic stress would significantly increase the abnormal pressure. The values of σ_1 in the Western Qaidam Basin were simulated in light of the statistical model of maximum principal stress established by Li et al. (2001) using resistivity and the calculation formula of pressure increment due to tectonic stress in a completely closed system proposed by Wang et al. (2005).

Since the deposition of the N_2^1, the Western Qaidam Basin has experienced an intense tectonic movement, greatly promoting the overpressure in this area. The contribution rate of tectonic compression to overpressure was calculated (Table 2), indicating that compressive stress with a contribution of 20–30% also has a significant impact on the overpressure development in the Western Qaidam Basin.

Conclusions

With increasing depth, the pressure in the Western Qaidam Basin steadily rises, and the shallow, middle, and deep strata exhibit distinct characteristics of low pressure, normal pressure, and overpressure. The depth of the overpressure top interface is different in various structures. Areas with a high-pressure coefficient moved eastward with the salinization centers, indicating that salts and their salinity had impacts on the pressure evolution from the E_{1+2} to N_2^3.

The combination of disequilibrium compaction, hydrocarbon generation, tectonic compression, and diagenesis leads to the overpressure in the Western Qaidam Basin.

Table 2 Contribution rate of structure extrusion stress pressurization of Well F2

Well name	Depth (m)	σ_1 (at)	S (at)	σ_1–S (at)	Contribution rate
F2	3005	921.41	755.52	165.89	0.26
	2804	892.32	697.31	195.01	0.30
	2718	799.07	672.64	126.43	0.16
	2481	879.12	605.95	273.17	0.42
	2379	801.43	577.80	223.63	0.38
	2075	791.68	495.03	296.65	0.40

Table 1 Contribution rate of disequilibrium compaction pressurization in typical wells

Well name	Depth (m)	Test pressure (at)	Calculated pressure at)	Disequilibrium compaction pressurization (at)	Excess pressure (at)	Contribution rate
F2	3692	605.97	567.10	169.15	208.03	0.81
	3005	388.10	352.75	28.87	64.21	0.45
J5	3098	396.30	391.30	57.34	62.34	0.92
	2941	426.10	367.75	50.71	109.06	0.46
	2236	282.50	262.06	20.98	41.42	0.51
	2143	266.60	248.05	17.03	35.58	0.48
	2093	266.60	240.55	14.92	40.97	0.36
	1949	242.70	218.94	8.84	32.61	0.27
N9	3208	502.62	482.45	136.65	156.82	0.87
	3055	416.96	382.00	52.67	87.63	0.60

Disequilibrium compaction is the most crucial factor, with a contribution rate of more than 60%, followed by tectonic compression, with a contribution rate of 20–30%. Disequilibrium compaction has a great contribution to the overpressure in the intersalt and subsalt strata. Compared with the intersalt strata, the contribution of hydrocarbon generation is greater in the subsalt strata. In the process of diagenesis, fracture reducing by salt filling and fluid volume expanding by gypsum dehydration promote the overpressure.

The strong plasticity and sealing properties of the gypsum salt rocks also create favorable conditions for the development of overpressure. The formation overpressure in the western Yingxiong ridge area is vertically developed under the gypsum salt rock. The distribution thickness of gypsum salt rock is horizontally well coupled with the pressure coefficient. The salts and their concentrations not only control the formation overpressure through a variety of factors, but also play important roles in preserving the overpressure under the background of tectonic compression.

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