The Depositional Environments and Shallow Subsurface Architecture of the Northeastern Caspian Sea

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Abstract

The Caspian Sea is often used to study the effects of sea-level change on coasts (Kroonenberg et al. 2000, Overeem et al. 2003, and Hoogendoorn et al. 2005). The results of these studies are of great importance to study the effect of eustatic sea-level change and analogue studies for hydrocarbon reservoirs. Relatively less research has been done in the Northeastern Caspian Sea, which is the flattest and shallowest part of the sea. The subsurface architecture and processes that lay down sediments in this area are relatively unknown. Therefore the objective of this study is to increase the knowledge of the depositional environments and shallow subsurface architecture of the Northeastern Caspian Sea. The results of this study can be used for the development of North Caspian offshore-petroleum-production facilities, as an analogue for reservoirs deposited in similar conditions and to study the effects of sea-level change on low gradient coasts.

For this project 6 new offshore cores have been drilled to a depth of 10m in the Northeastern Caspian Sea. In addition, a large amount of shallow subsurface data has been gathered from cores, cone penetration tests and seismics, during the development of offshore hydrocarbon-fields in the area. Sedimentological descriptions of the new cores were made, samples were dated with radiocarbon techniques and a biostratigraphical analysis was performed. The depositional environments for the different lithofacies were determined and a sequence stratigraphical framework was developed. With the available data a 3D-subsurface model was build with Petrel software to visualize the spatial sediment distribution and test the developed concepts.

Four lithofacies, deposited in different depositional environments were found in the cores. The oldest deposits are dated before 48.000y BP, they have been deposited as aeolian dunes during a lowstand and were reworked during a transgression by coastal processes. On top, lagoonal sediments are deposited between at least 48.000y BP and 42.000y BP during multiple centennial to millennial or less frequent sea-level cycles. These deposits are overlain by transgressive barrier-like sands which have been radiocarbon dated around 27.000y BP. The sediments in the top of the cores have been deposited after the Mangyshlak or Derbent regression in an environment with a low sediment influx and frequent sea-level changes. The study shows that the depositional environment in a low gradient environment with frequent sea-level changes mainly depends on the relative position to the coast, the supply of sediments and the speed of sea-level change. It increases our knowledge of the depositional environments and subsurface architecture of the Northeastern Caspian Sea.
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1 Introduction

The Caspian Sea is the largest continental water body on Earth and located at the border of Europe and Asia, East of the Caucasus, and North of the Elbruz mountain range. The sea forms a completely enclosed basin. Because the Caspian Sea is not connected to any ocean, the water level is dependent on the inflow of river water and evaporation. This causes the water level to be very dynamic. Sea-level cycles at multiple scales have taken place frequently in the past. The last full sea-level cycle occurred between 1929 and 1995 and had an amplitude of 3m (Kroonenberg et al. 2000).

Sea level regulates the balance between accommodation space and sediment supply and governs the position of the coast. The Caspian sea-level change has a large effect on the depositional environments in coastal zones. The depositional environment is defined as the combination of physical and chemical processes that lay down sediments. It has a large influence on the spatial distribution of sediments, and in this way through time on the sedimentary architecture.

The effects of the Caspian sea-level change on sedimentary processes have already been studied in the Dagestan barrier coast by Kroonenberg et al. (2000), in the Volga delta by Overeem et al. (2003) and in the Kura delta by Hoogendoorn et al. (2005). There is a large interest for these studies for predicting the effects of eustatic sea-level change on coasts worldwide. The North Caspian is also used as an analogue for hydrocarbon reservoirs deposited in similar settings with frequent sea-level changes, such as the South Caspian basin. The conditions under which the hydrocarbon reservoirs of the South Caspian basin formed are comparable with the present environment in the North Caspian Sea with frequent sea-level changes.

Less research has been performed on the Northeastern part of the Caspian Sea. This area has an extremely low on- and offshore gradient, 40 kilometres offshore the waterdepth is just 3m. The low gradient, in combination with the frequent sea-level changes cause large shifts of the coastline. Satellite images show that during the latest sea-level rise the coast shifted 40km landward. The low gradient, shallow waterdepths and frequent sea-level changes have profound effects on the sedimentary processes. In the last couple of years, more interest has been shown for the geology of the Northeastern part of the Caspian Sea. Multiple hydrocarbon fields are developed in the area, including the Kashagan field, one of the world's largest new discoveries. The field is developed from manmade artificial islands. Information about the subsurface and the depositional environment is crucial for the construction of the offshore production facilities. In the past years, many cores have been drilled in the shallow subsurface. This data provides valuable information about the depositional environments and subsurface architecture of the area in the recent past.

The objective of this research is to increase the knowledge of the depositional environments and shallow-subsurface architecture of the Northeastern Caspian Sea. This was achieved by analyzing 6 newly drilled cores, existing geotechnical cores and cone penetration test data. The depositional environments were determined and a sequence stratigraphical framework was developed. Finally a 3D-subsurface model was build to visualize the data and test the developed concepts.
The results of this study can be useful for the development of North Caspian offshore-petroleum-production facilities. It can also be used as an analogue for hydrocarbon reservoirs deposited in similar conditions, for example the South Caspian basin Productive Series. The results of this research can also be useful to study the effects of eustatic sea-level change on other low gradient coasts.
2 The Caspian Sea

2.1 - Regional geology of the North Caspian Basin

The North, South and middle Caspian basins are located within the Alpine-Himalayan collisional zone (Figure 1). It separates the Neogene continental collision of the Caucasus Mountains in the west from large scale faulting in the Kopeh-Dagh fold belt to the east. This boundary is marked by the Absheron Ridge which separates the South Caspian basin from the Middle and North Caspian basins (Knapp et al. 2003).

The North Caspian basin formed in a rift zone. The time of rifting remains disputable and theories suggest that rifting took place during the Late Proterozoic, Early Ordovician or Middle Devonian (Table 1). The spreading resulted in cratonic blocks moving away from the Russian craton and opening of the oceanic crust. In the Late Palaeozoic, subsidence resulted in the deposition of carbonate and clastic sediments on the basin’s margins, and deepwater shales and turbidites basinward (Table 1). These sediments are overlain by Permian evaporates. On top of the Permian salt, Mesozoic and Paleogene sediments reach up to a thickness of 6km (Ulmishek 2001b). The major oil and gas reserves of the North Caspian, including the supergiant Kashagan field, are located in the subsalt carbonate reservoirs. Deformation of the Permian salts played an important role in the formation of these oil and gas fields. The most active halokinesis occurred during the Late Permian-Triassic time, but growth of the salt-domes continued later and some of them are exposed on the present day surface (Ulmishek 2001b).

During the Late Tertiary, the Caspian basins where part of the Eastern Paratethys Sea, which formed during the Eocene by the closure of the Tethys Sea (Table 1). During the Late Miocene the Caspian developed into a totally enclosed basin. At the same time sea level dropped extremely (600m to 1500m). Only the southern part of the present Caspian Sea was filled with water at this time. Multiple explanations exist for the dramatic sea-level fall and isolation of the Caspian Sea. Some attribute the isolation of the Caspian Sea to the sea-level drop itself and correlate it with the Messinian sea-level drop in the Mediterranean region (Reynolds et al.1998). Others suggest that Caspian was already an enclosed basin and sea-level drop was caused by rapid subsidence of the South Caspian basin initiated by the Arabia-Eurasia collision (Kroonenberg et al. 2005).

As a response to the change in base level, rivers such as the Paleo-Volga, Paleo-Amu Darya and Paleo-Kura flowed into the South Caspian basin. The south flowing rivers carved out a canyon up to 200m deep in the North Caspian plateau. The Paleo-Volga delta at this time was located near the Absheron peninsula and deposited the main reservoir formations of the South Caspian basin, known as the Productive Series. The modern Volga-delta, which has the same drainage area and enclosed basin as the Paleo-Volga is often used as an analogue for these reservoirs (Kroonenberg et al. 2005).

The Miocene lowstand was followed by multiple transgressions and regressions in the Pliocene, Pleistocene and Holocene caused by climatic changes (Kroonenberg 2008). But a lowstand as during the Miocene was never reached again.
From the Miocene on, there has been no active subsidence in the North Caspian Basin. Also the sedimentation rates slowed down. The deposits of the Late Pliocene, Quaternary and more recent times do not exceed a thickness of 300m in the North Caspian Basin. The only recent tectonics in the North Caspian is caused by halokinesis of the Permian salts (Kroonenberg et al. 2005).

**Figure 1**
Map of the Caspian Sea; 1) the North Caspian basin, 2) the Middle Caspian basin, 3) the South Caspian basin; 4) the Kara-Bogaz-Gol Bay. Isobaths are shown in meters (Kosarev 2005).
<table>
<thead>
<tr>
<th>Era</th>
<th>Period</th>
<th>Epoch</th>
<th>Ma</th>
<th>Major events</th>
</tr>
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<tr>
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<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Pleistocene</td>
<td>1,8</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Pliocene</td>
<td>5,3</td>
<td>Incision of Paleo-Volga in NCB + deposition of reservoirs in SCB</td>
</tr>
<tr>
<td>Cenozoic</td>
<td>Tertiary</td>
<td>Miocene</td>
<td>23,7</td>
<td>Enclosure of CB + dramatic sea-level drop</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Oligocene</td>
<td>33,7</td>
<td>Development of Paratethys Sea</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Eocene</td>
<td>54,8</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Palaeocene</td>
<td>65</td>
<td></td>
</tr>
<tr>
<td>Mesozoic</td>
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<td>Cretaceous</td>
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<td></td>
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<tr>
<td></td>
<td>Jurassic</td>
<td></td>
<td>206</td>
<td>Active halokinesis of Permian salt in NCB</td>
</tr>
<tr>
<td></td>
<td>Triassic</td>
<td></td>
<td>248</td>
<td>Deposition of Permian Evaporites in NCB</td>
</tr>
<tr>
<td></td>
<td>Permian</td>
<td></td>
<td>290</td>
<td>Deposition of carbonate reservoirs in NCB</td>
</tr>
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<td></td>
<td>Pennsylvanian</td>
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<td></td>
<td></td>
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<tr>
<td></td>
<td>Mississippian</td>
<td>354</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Palaeozoic</td>
<td>Devonian</td>
<td></td>
<td>417</td>
<td>? Start development of NCB as a rift</td>
</tr>
<tr>
<td></td>
<td>Silurian</td>
<td></td>
<td>443</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Ordovician</td>
<td></td>
<td>490</td>
<td>? Start development of NCB as a rift</td>
</tr>
<tr>
<td></td>
<td>Cambrian</td>
<td></td>
<td>543</td>
<td></td>
</tr>
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</table>

Table 1
Geological time scale with the major geological events of the North Caspian basin (NCB) and important events in the South Caspian basin (SCB).
2.2 - Caspian sea-level change
Changes in Caspian sea level take place at much shorter time scales than in oceans. For example, from 1930 until 1995 a complete sea-level cycle took place with amplitude of 3m (Kroonenberg et al. 2000).

Because the Caspian Sea is an enclosed basin, sea level depends mainly of the balance between inflow from rivers and evaporation of surface water. Minor climate changes in the rivers drainage areas and Caspian region have large effects on the sea level. These climate changes are again depending on regional and worldwide climate changes caused by atmospheric circulation patterns and solar activity. The link between global climate change and the Caspian sea level is not clear, but it is known that sea-level change occurs in at least 5 different time scales; glacial/interglacial, centennial/millennial, decadal, seasonal and during storm surges (Kroonenberg et al. 2008).

2.2.1 Glacial/interglacial sea-level change
Glacial and interglacial periods during the Quaternary have been the reason for numerous major Caspian sea-level changes. Two reasons for glacial/interglacial initiated sea-level change exist; the melting of glaciers (Mamedov, 1997) and the cut-off of north-flowing rivers by ice sheets, causing the rivers to flow southwards into the Caspian Sea (Mangerud 2001). According to these theories highstands can occur during an early glacial and at the beginning of an inter-glacial. However, the exact time periods of the Caspian sea-level cycles are still disputable.

The highstands during the Quaternary can be distinguished from marine terraces along the Caspian coasts. Evidence for highs of at least +50m and lows around -120m Above Mean Sea Level (AMSL) has been found (Dumont 1998). Lowstands are harder to distinguish in the geological record. When the sea level falls below -34m AMSL the entire northern part of the Caspian Sea is exposed and the Volga, Ural and Emba rivers carve out valleys which are filled up with sediments during transgressions (Kroonenberg et al. 2008).

The main Holocene and Late Pleistocene transgressive and regressive phases are plotted in Figure 2. Evidence for the Baku transgression around 400,000y Before Present (BP) and the Khazar transgression around 200,000y BP was found in outcrops along the lower Volga and marine terraces along the western Caspian coast (Kroonenberg et al. 2008). The regression which followed is called the Atel regression.

There is much debate about the subsequent Early Khvalyn transgression Inter Khvalyn regression (or Yenetovan regression) and Late Khvalyn transgression. Rychagov (1977) dated the Early Khvalyn transgression using marine terraces along the Dagestan coast. He concluded that this highstand took place during the early glacial around 70,000y BP (Kroonenberg et al. 2008). According to Mangerud (2001) the glacial highstands in the Caspian Sea are related to the overflow of glacial lakes to the north of the Caspian Sea, caused by the cut-off of north-flowing rivers by ice sheets. Mangerud's study (2001) supports the theory of a highstand during the early Weichsel glacial period around 90,000y BP (Kroonenberg et al. 2008). Mangerud (2001) also found evidence for a minor highstand during the late Weichsel glacial period, which is probably the Late Khvalyn highstand (Kroonenberg et al. 2008). The lowstand between these two highstands is then the Inter Khvalyn or Yenetovan
regression. Svitoch (2006) dated the Early Khvalyn transgression between 15.000y BP and 11.000y BP. He argues that the Caspian Sea at this time, overflowed to the Black Sea. Mamedov (1997) dates the Early Khvalyn transgression around 32.000y BP and 24.000y BP. And the Late Khvalyn transgression between 16.000y BP and 8.000y BP. Together these studies indicate highstands around 90.000yBP, 70.000y BP, 24.000y BP and 11.000y BP.

The Mangyshlak regression was the latest glacial/interglacial regression and took place probably during the early Holocene around 10.000 or 8.000 years ago. During this lowstand the sea level dropped to at least -50m AMSL (Mamedov 1997). Since this lowstand, the Caspian sea level has risen to the present level with smaller scale transgressions and regressions in between. The present day transgression is called the Novocaspian transgression.

![Figure 2](none)

**Figure 2**

2.2.2 Centennial to millennial sea-level change
During the Holocene, the Caspian sea level has fluctuated multiple times between roughly -40m and -20m AMSL. Global climate change, caused by differences in solar activity are thought to be the main reason for these centennial to millennial sea-level fluctuations (Kroonenberg et al. 2007).

The best information for Holocene sea-level changes comes from the dating of cores from the Volga and Kura deltas and the Dagestan barrier lagoon coast. Rychagov (1977, 1997) published the first Holocene Caspian sea-level curve by C14 dating of molluscs from deposits in barriers and incised valleys in Dagestan. More recently a new Caspian sea-level curve was created with data from the Dagestan barrier coast, Volga Delta and Kura Delta. The oldest points in this new Holocene Caspian sea-level curve (Figure 3) were dated with AMS dating of sediments from the Volga delta.
The data indicate a lowstand around 8.000y BP. Between 8.000y and 5.000y BP no data are available. The data points between 5.000y and 3.000y BP suggest a constant sea-level rise until a highstand around 2.600y BP is reached. The proof for this highstand was found in barriers along the Dagestan coast by Kroonenberg et al. (2007). In the Middle Ages, the sea level reached a lowstand down to -34m AMSL or even -45m AMSL. This regression is known as the Derbent regression. The proof for this regression was found in the Kura delta by Hoogendoorn et al. (2005). Evidence for a highstand around 300 years ago was also found in the Dagestan barrier coast (Kroonenberg et al. 2000).

The two latest highstands can be correlated with two well known periods with colder temperatures in Europe; the Subboreal between 5600y BP and 2400y BP, and during the period known as the 'little ice age' in the 17th and 18th century. The Derbent regression seems to be related to the warmer period of the Middle Ages (Kroonenberg et al. 2007).

![Figure 3](image)

**Figure 3**
The Holocene Caspian sea-level changes (Kroonenberg et al. 2008).

2.2.3 Decadal sea-level change

The Caspian sea level has been measured since the early 19th century. The sea-level graph of the last century is shown in Figure 4. From 1930 to 1995 a complete cycle with a sea-level rise of 3m took place. The change is related to the enclosed nature of the Caspian Sea and the fact that the Volga River provides 80% of the water inflow in the Caspian (Kroonenberg et al. 2007). Evaporation is also an important factor in the Caspian sea-level change, especially in the shallow waters of the North Caspian and Kara-Bogaz bay. The amount of rainfall in the catchment area of the Volga is strongly dependent on variations in the North Atlantic Oscillation. The
strength curve of the North Atlantic Oscillation fits the Caspian sea-level curve of this century very well (Kroonenberg et al. 2008).

During the 1975 lowstand of the Caspian Sea, the eastern coastline of the North Caspian shifted up to 40km seaward. On satellite images from that time, white evaporate deposits can be seen at the former sea bed. These deposits can be expected in the subsurface of these areas.

![Graph showing sea-level changes in the 20th century.](image)

**Figure 4**
The 20th century Caspian sea-level changes (after Kosarev 2005).

2.2.4 Seasonal sea-level change
Seasonal changes of the Caspian sea level are related to snow melt and precipitation in the river’s drainage areas and increased evaporation during summer times. The sea level is monitored from space every 10 days by the Topex/Poseidon satellite in the past and more recent by the Jason-1 radar satellite (Figure 5). The difference in water level between summer and winter is around 0.4m.
2.2.5 Storm surges

The water body of the Caspian Sea is too small for tides. However, storm surges can cause local sea-level changes. When the winds in the North Caspian are coming from the west the water is piled up in the eastern part of the sea and can locally rise up to 2.5m. More common surges with a sea-level rise of more than 0.4m are observed 5 to 20 times per year (Kosarev and Yablonskaya 1994). When the winds are coming from the east, the water is blown away from the shallow areas where the sea level can drop with 1m meter in extreme cases. These surges commonly occur during the spring and fall and can flood or expose large parts of coastal lowland. In the summer the weather is normally calmer (Kosarev and Yablonskaya 1994). During the winter season the surges are damped out, because of ice cover in the northern part of the Caspian Sea (Kosarev and Yablonskaya 1994).

Figure 5
The Caspian sea-level changes from 1993 until the 20th of Jan 2009 after measurements from the Topex/Poseidon and Jason-1 satellites (after United States Department of Agriculture 2009)
2.3 - Physio-Geographical conditions of the Northeastern Caspian Sea

2.3.1 Present day depositional environments

The northeastern part of the Caspian Sea is the shallowest part of the sea and makes up about 29% of the entire area of the sea, though its volume only comprises 1%. The average water depth in this area is 6m. The maximum depth does not exceed 10m and 20% of this area has a water depth less than 1m (Figure 1). The offshore slope is around 0.007°. This shallow gradient has a profound effect on coastal processes.

Based on its morphology the coastal zone of the Northeastern Caspian Sea can be divided into four areas, coastal plain, mud flats, a reed-bed barrier zone and the open sea bordering the reed-bed barriers (Figure 6). Because of the gradual change from sea to land with reed in between and the dynamic sea level, the coastline is hard to define, it can shift over tens of kilometres in a couple of years due to sea-level change.

The coastal plain (Figure 6) comprises sandy and clayey (semi-) deserts. It is drained by rivers like the Ural and Emba. The Emba is only flowing to the sea during spring when the snow melts. Sediments are transported by winds and in the past formed longitudinal dunes (Figure 7A) (Leont'ev and Foteeva 1965). These dunes were first described by Baer (1855) and are also known as Baer mounds. They consist of sandy clays and contain small shell fragments (Baer 1855). The dunes can be found in the whole North Caspian coastal region. Some authors doubt the aeolian origin of the dunes, Badyukova (2005) attributes the dunes to be the result of sheet flow during the Khvalyn highstand, when the Caspian Sea was in contact with the Black Sea. From satellite images the size of the dunes was measured, they have an average length of 4km, an average width of 300m. In the field thicknesses between approximately 3m to 20m were observed. Numerous small lakes can be found between the dunes, these are filled with evaporates and are locally called salonchaks (Figure 7C). During winter times snow covers these areas and the ground is frozen. The vegetation is scarce and only salt tolerant vegetation can grow here.

The drowned mud flats or lagoons (Figure 6) are characterized by their flatness and shallow water depths, not exceeding 1.5m. Sediments are transported by moderate water currents. Deposits are nearly reaching the water surface in many areas, especially close to the reed beds (Figure 7D). This area is protected from waves by the reed beds resulting in a low energetic environment. Therefore mainly fine sediments are found in this area. Partly submerged vegetation grows in this area, especially close to the reed-beds barriers.

The reed-bed barriers (Figure 6) consist of partly submerged barriers, overgrown with perennial reeds. These reed beds can only grow in very shallow water and have moved landward during the last sea-level rise. Also submerged vegetation grows in this area. Both have an important land stabilizing function and act as a filter between the terrestrial and marine ecosystems (Kaplin and Selivanov 1994). The areal part of the reeds dies during the winter but persist for multiple years before falling on the ground to decompose. These reed zones are a potential trap for sediments (Masataka et al. 1998). Sedimentation processes in these areas are driven by wind
surges and wave flooding processes in the form of small washover fans, build across the reed-bed supported sandbanks (Figure 7B). On the seaward side, the vegetation is delineated by a sharp line, probably due to effects of inflow currents of wind surge waters. Channels are represented by openings in the reed beds and extend into the mudflats.

In the open sea, bottom relief is created by wave action, wind-induced surge currents, and circular offshore currents. The floor relief shows wide linear sandy ridges created by wind surges and waves and are locally called shalygi. The depth at which they occur ranges from 1.5 m to 3m and the longitudinal axes of these shalygi generally follow the contours of the coastline (Kamlesh and Dessinov, 2000). During large regressions, river systems drain the area and create relief. The deepest part of the Northeast Caspian Sea is occupied by de Ural Furrow and has a maximum depth of 10m (Figure 1). The Ural Furrow is a canyon carved out by the Ural River during regressions. The depression is also associated with salt tectonics and a major regional fault extending north-eastward from the Astrakhan Bay to Atyrau (Leont’ev et al. 1977). During recent regressions the Ural Furrow was a lake, which was drained to the Caspian Sea along a valley that goes through the shallow area of the Tyulenie islands to the Mangyshlak Furrow (Leont’ev et al. 1977).
Figure 6
Morphological zones in the northeastern part of the Caspian Sea and locations of photos A, B, C and D of Figure 7 (after Google Earth 2008 and Kamlesh and Dessinov 2000).

Figure 7
The photo locations are indicated in Figure 6. A) Satellite image of longitudinal dunes east of Atyrau Kazakhstan, the locations of, with evaporates or ice in the interdunal areas (Google Earth, 2008). B) Reed island with overflow deposits (Google Earth 2008). C) The coastal plain with a salonchak south of Kolsary, Kazakhstan. D) A North Caspian lagoon with reed islands (Bolattiguzzo 2004).
2.3.2 Hydrology

The prevailing wind direction in the North Caspian Sea is easterly. Westerly and northwesterly winds are also common (Kosarev 2005). The creation of waves by winds is restricted by the low water depths. Northwesterly, easterly and southeasterly travelling waves are prevailing. The maximum wave height increases with the water depth. During storms the wave heights near the Volga delta can be approximately 0.5m while at the shelf edge near the boundary with the middle Caspian waves can be around 4m (Kosarev 2005). Bottom compensatory currents can occur in the opposite direction of the wind direction.

The main water currents are affected by the wind, coastline, bottom geometry, river discharge and the Coriolis Effect. The water transport from the Volga River takes place along the western shore to the south towards the Absheron peninsula. This inflow of fresh water in combination with the Coriolis Effect drives the general anti-clockwise current direction of the North Caspian Sea. The outflow from the Ural River further increases this cyclonic current (Kosarev and Yablonskaya 1994, Kosarev 2005).

The continental climate in the Caspian region, in combination with the shallow water depths, results in large temperature shifts between summer and winter. During mid summer the average water temperature is 24°C and in some shallow parts, water temperatures rise up to 35°C. Ice usually starts to form in November. In severe winters the whole northern part of the Sea is covered with ice. During milder winters only the areas within the 2m to 3m isobaths are covered with ice (Kosarev and Yablonskaya 1994).

The salinity of the North Caspian changes from north to south. In the north, the salinity is around 0.1 parts per thousand (ppt) in the south the salinity is around 11 ppt (Aubrey et al. 1994). This is explained by the influx of river water from the Volga and Ural rivers in the north. These rivers have a strong effect on the salinity because of seasonal and yearly differences in freshwater inflow (Kosarev and Yablonskaya 1994). The Caspian Sea has a high sulphate and calcium content. These high concentrations combined with a high water temperature in the summer makes gypsum (CaSO₄·2H₂O) a common deposit in the Caspian Sea region.

2.3.3 Rivers

The Volga provides 80% of the total water supply to the Caspian Sea. The river drains an area of approximately 1,380,000 square km. Its annual discharge is 7,835 m³/s over 1888-1980, the peak runoff occurs in May. Its sediment output is around 38 million m³/y (Castelltort and van den Driessche 2003).

In the upper delta plain five major distributaries start fanning out from the apex and incise into late Pleistocene and early Holocene east-west elongated eolian longitudinal dunes, the so called Baer hills. Deposition during annual floods largely takes place in the interdunal areas (Overeem 2003). The most dynamic parts of the delta are the lower delta plain and the delta front. In these areas most sedimentation takes place and deltaic sequences are deposited on top of Pleistocene and early Holocene topography. In the lower delta plain the distributary channels are spread over 800 outlets along the total coastline of 200km. Most of these outlet channels are not wider than 10-20m. The lobate shape and multi-scale terminal distributary channels are typically for river-dominated deltas in shallow water basins (Overeem 2003). Fine
particles are carried far away from the delta before they settle. Although the Volga delta is located in the west of the North Caspian, its huge suspended sediment output can potentially be a source for fine sediments to the Northeastern Caspian Sea.

The Ural River starts in the Ural Mountains and flows into the Caspian Sea south of the city of Atyrau and has a length of 2430km. The Ural delta discharge varied between 3km$^3$ and 20.5km$^3$ per year in the last 50 years. 80% of the discharge comes from April to August, with the minimum just before ice appears in November (Kosarev and Yablonskaya 1994). The sediment outflux of the Ural is around 2,2 million m$^3$/y (Castelltort and van den Driessche 2003). The Ural delta begins just below the town of Atyrau and it spreads almost 40km to the south and southeast to Coere, an area of about 600km$^2$. The drainage network has two main branches, the Zolotoi and Yaitsky. The main discharge, up to 75 to 80 %, flows along the Zolotoi and only 20% goes along the Yaitsky. The Zolotoi branch constitutes the river part of the Ural Caspian shipping channel, which further downstream merges with the marine part of the channel. This channel connects the mouth of the river with the Ural furrow, which is the deepest part of the North Caspian (Kosarev and Yablonskaya 1994). Given the sediment output and location, the Ural delta can be a potential source for sand and clay to the study area.

The Emba River starts in the Mughodzar hills and has a length of approximately 640km. It flows through the north of the Ust Urt plateau and splits up in 2 channels near the town of Kolsary. Further downstream, it flows through a series of shallow lagoons. During the 18th century highstand the channels were used for shipping. Now, the river only reaches the sea during the spring when the snow melts. The rest of the year most parts of the downstream river beds are dry or contain standing water. Although the outflux from the Emba river is negligible at present time it might have had a higher outflux in the past, and can have supplied sand and clay for the study area.

2.3.4 Sediment composition and provenance

Khrustalov and Ryshkov (1975) studied the provenance of bottom minerals of the North Caspian Sea using heavy minerals. They concluded that sediments of the western part of the North Caspian Sea, around the Volga River delta are logically the results from discharge of terrigenous material by the Volga River. The Sediments in the Ural Furrow and around the Ural delta are mainly Ural River sediments and of lesser importance, Volga River sediments, aeolian sediments and suspended material from the Western half of the Northern Caspian. The eastern part of the North Caspian Sea, south and east of the Ural Furrow are mainly controlled by the supply of terrigenous material from aeolian sediments from the Buzachi Peninsula. Large quantities of Emba River sediments around this river’s mouth indicate a strong outflux of suspended load and bed-load from this river in the past (Khrustalev and Ryshkov 1975).

Because of the frequent sea-level changes and low gradient large vertical facies shifts in the stratigraphical record can be expected (Overeem 2002, Hinds 2004). For example satellite pictures made, during the last lowstand in 1975 show white evaporates on the former sea bed (Figure 8). While at present times, the water depth reaches up to 3 m in the area. Molluscs and phytoplankton are abundant in the waters North Caspian. Therefore sediments with a biogenic origin can be
expected, especially in the areas with a low terrigenous sediment influx, such as the eastern part of the North Caspian.

Figure 8
1975 Landsat satellite image of the eastern shore of the North Caspian Sea. White evaporite deposits can be seen on the former sea bed. The borehole locations are indicated in red. The present coastline is located approximately at the landward edge of the white deposits.
3 Data

For this project 6 new cores in PVC liners (CDS boreholes) have been drilled in the Northeastern Caspian Sea. These cores were used to make detailed sedimentological logs and descriptions of the subsurface. Samples for radiocarbon dating and biostratigraphical analysis were also taken from these cores. In addition, data from; existing geotechnical boreholes, cone penetration tests and seisms, bathymetrical data and satellite images have been used. During the core analysis in Kazakhstan a short fieldtrip was organized to investigate the Emba River and surrounding coastal plain. An overview of the study area with boreholes, CPT locations, seismic surveys and bathymetry is shown in Figure 9 and Table 2.

![Figure 9](image)

**Figure 9**
Map showing the study area, bathymetry, boreholes CDSA till CDSF, seismic surveys and the clusters of geotechnical boreholes and cone penetration tests.

<table>
<thead>
<tr>
<th>cluster</th>
<th>geotechnical boreholes</th>
<th>CPTs</th>
</tr>
</thead>
<tbody>
<tr>
<td>A/F</td>
<td>12</td>
<td>8</td>
</tr>
<tr>
<td>Aktote</td>
<td>4</td>
<td>0</td>
</tr>
<tr>
<td>C</td>
<td>5</td>
<td>0</td>
</tr>
<tr>
<td>D</td>
<td>16</td>
<td>32</td>
</tr>
<tr>
<td>G</td>
<td>5</td>
<td>0</td>
</tr>
<tr>
<td>Kairan</td>
<td>6</td>
<td>0</td>
</tr>
</tbody>
</table>

**Table 2**
Number of boreholes and CPTs in each cluster.
3.1 - Liner cores (CDS boreholes)

3.1.1 Coring procedure

The locations of the 6 new boreholes (CDS boreholes) are shown in Figure 9. The cores were taken below the 1975 lowstand coastline, which follows the 2m isobaths (Figure 8). This was done because major facies shifts are expected in this area. The drilling pattern was chosen in such a way that cross-sections in multiple directions could be made. This is important because it can give information about the lateral extend and sediment source of the deposits. The cores were drilled from a drilling barge with a Wireline-Push (WIP) sampler (Figure 10). For the liner cores the WIP sampler was equipped with a 70mm PVC liner, inserted in the core barrel and locked in place with the cone of a core catcher (Figure 10).

The recovery from the first 2m in core CDSB-1A was very poor. Therefore a second core was taken within a few meters from the first core. This 2m deep core is referred to as CDSB-1B. In core CDSE-12 a hard layer was hit at 8m and was too hard for the thin walled corebarrel. Therefore a thick walled barrel was used without a PVC liner. The two cores were stored in the thick steel core barrel itself. After sampling, the cores were stored in the PVC liners in a cooled container. The ends of the liners were sealed with a cap, and in some cases filled with wax to prevent the core from falling apart.

The liner cores were opened and described in the Fugro geotechnical laboratory in Atyrau, Kazakhstan. A vibrating saw was used to open the liners. The two deepest cores in borehole CDSE-12, stored in a steel core-barrel, had to be pushed out with a pneumatic device. The cores were cut in half with a wire saw. Samples for C14 analysis and texture analysis were taken with a modified plastic syringe. Photos of each core were made before and after sampling. For each borehole a log was made showing the sedimentary features of the soil.
Figure 10

Drawing of the wire-line push sampler locked in a bottomhole assembly (Fugro 1996) and corebarrel with PVC liner. The WIP sampler consists of a downhole jacking unit and a 1m long core barrel with a diameter of 51mm or 76mm. The sample procedure uses a piggyback drilling system. During piggyback drilling a hole is drilled with conventional open-hole rotary-drilling equipment. When the coring depth is reached the sampler is lowered into the drill-pipe and locked into position at the bottomhole assembly. By increasing the mud pressure in the drill-pipe the core-barrel is pushed into the ground with a constant rate of penetration. After the sample has been taken, the WIP sampler can be retrieved to the surface with a wire-line. Depending on the expected soil type the core barrel can be equipped with a core catcher to prevent the sample from falling out. The cone of the core catcher was screwed on the corebarrel to hold the PVC liner in place.
3.1.2 - Core descriptions

The liner cores are described in lithological logs and photo logs. The logs indicate the main grain size, secondary soil content, organic content, colour according the Munsell colour classification for soils, layering, laminations and other observations such as signs of oxidation and overconsolidation. The lithological logs, photo logs and a legend are shown in Figure 11.

Figure 11
Sedimentological logs of cores CDSA, CDSB-1A, CDSB-1B and CDSC.
Figure 11
Sedimentological logs of cores CDSD, CDSE and CDSF.
3.2 - Sample analysis

Samples for texture analysis, radiocarbon dating and biostratigraphical dating were taken from the CDS liner cores. The texture samples are taken at places in the liner cores which are representative for a specific interval, or sequence in the cores. The C14 and biostratigraphical samples are taken near lithological boundaries. In total 137 C14 samples and 178 texture samples were obtained from the 6 cores. The 19 most promising C14 and biostratigraphical samples were taken to The Netherlands for further analysis. Due to logistical reasons the other samples arrived too late to be included in this study.

3.2.1 Biostratigraphical analysis

From the 19 samples the 13 best ones were analyzed on microfossils by StrataData in the UK. Each sample was washed through a 125µm sieve and the residue dried at 100°C. The dried residue was then examined for microfossils and estimates were made of the relative abundance of each species. The type and amount of each species are typical for a specific environment. The results of the biostratigraphical analysis can be found in Table 3.

<table>
<thead>
<tr>
<th>Borehole</th>
<th>Depth (m)</th>
<th>Abundant species</th>
<th>Species Environment</th>
</tr>
</thead>
<tbody>
<tr>
<td>CDSA-11</td>
<td>7,1</td>
<td>Caspiolla sp., Bacuniella Dorsoarcuata</td>
<td>Caspian marine</td>
</tr>
<tr>
<td>CDSA-11</td>
<td>8,75</td>
<td>Paracytherideis Naphtatscholana, Leptocythere</td>
<td>Caspian marine</td>
</tr>
<tr>
<td>CDSB-1</td>
<td>1,1</td>
<td>Cyprideis Torosa</td>
<td>Backshore</td>
</tr>
<tr>
<td>CDSB-1</td>
<td>1,45</td>
<td>Cyprideis Torosa</td>
<td>Backshore</td>
</tr>
<tr>
<td>CDSB-1</td>
<td>8,85</td>
<td>Dreissena sp., Didacna sp., Caspia sp.</td>
<td>Marginal salt marsh</td>
</tr>
<tr>
<td>CDSC-15</td>
<td>8,25</td>
<td>Bacunella dorsoarcuata, Paracyprideis Naphtatscholana, Caspiolla sp.</td>
<td>Marginal brackish</td>
</tr>
<tr>
<td>CDSC-15</td>
<td>8,9</td>
<td>Paracyprideis Naphtatscholana, Leptocythere sp.</td>
<td>Caspian marine</td>
</tr>
<tr>
<td>CDSD-13</td>
<td>6,3</td>
<td>Caspiolla sp., Bacuniella dorsoarcuata</td>
<td>Caspian marine</td>
</tr>
<tr>
<td>CDSD-13</td>
<td>8,6</td>
<td>Caspiolla sp., Bacuniella dorsoarcuata</td>
<td>Caspian marine</td>
</tr>
<tr>
<td>CDSD-13</td>
<td>9,8</td>
<td>Paracyprideis Naphtatscholana</td>
<td>Caspian marine</td>
</tr>
<tr>
<td>CDSE-12</td>
<td>5,2</td>
<td>Bacunella dorsoarcuata, Caspiolla sp.</td>
<td>Caspian marine</td>
</tr>
<tr>
<td>CDSE-12</td>
<td>9,8</td>
<td>Dreissena sp., Cyprideis torosa</td>
<td>Marginal brackish</td>
</tr>
<tr>
<td>CDSF-14</td>
<td>9,6</td>
<td>Bacunella dorsoarcuata, Caspiolla sp.</td>
<td>Caspian marine</td>
</tr>
</tbody>
</table>

Table 3
Biostratigraphical analysis results.

3.2.2 Radiocarbon dating

Suitable specimens were selected from the residues made during the biostratigraphical analysis for C14 dating, which was carried out by Beta Analytic Inc. In most cases bivalves were used as these easily provided enough material for analysis. In a few cases ostracods were used where they looked ‘fresher’ or where bivalves were absent, every attempt was made to avoid the use of obviously reworked specimens (Athersuch 2008).

Most of the samples have a radiocarbon age between 48.000y BP and 27.000y BP (Table 3). The radiocarbon ages of these samples were converted to calendar years BP using Fairbanks et al. (2005) online marine reservoir calibration database. However, many of the samples were older than 48.000 years, which is outside the
resolution for radiocarbon dating. The samples which are dated older than 48,000\text{y} are shown as “< 48,000”, they have not been calibrated to calendar years. One sample (CDSB-110cm) was too young to be calibrated with the marine reservoir age correction, it has a radiocarbon age of 390 uncorrelated years. According to its depth this sample must be older than 0 calibrated (cal) years. This raises the question if the standard global marine reservoir age can be used in these sediments. A freshwater calibration of the same sample would give an age of 490 cal years BP. However, there is no suggestion from the C13/C12 ratio (-0.4) that this is a freshwater dominated environment (Athersuch J., 2008). It is more likely that the sample is not in situ and comes from higher up in the core. Overall the data seems robust, the dates agree with the principle of superposition. An overview of the results can be found in Table 4.

<table>
<thead>
<tr>
<th>Borehole</th>
<th>Depth (m)</th>
<th>Species for 14C</th>
<th>14C age (cal. years BP)</th>
</tr>
</thead>
<tbody>
<tr>
<td>CDSA-11</td>
<td>7,10</td>
<td>Ostracods</td>
<td>44.186 +/- 782</td>
</tr>
<tr>
<td>CDSA-11</td>
<td>8,75</td>
<td>Dreissena</td>
<td>46.781 +/- 1.134</td>
</tr>
<tr>
<td>CDSB-1</td>
<td>1,10</td>
<td>Bivalves</td>
<td>modern</td>
</tr>
<tr>
<td>CDSB-1</td>
<td>1,45</td>
<td>Ostracods</td>
<td>27.764 +/- 185</td>
</tr>
<tr>
<td>CDSB-1</td>
<td>8,85</td>
<td>Bivalves</td>
<td>&lt;48.000</td>
</tr>
<tr>
<td>CDSC-15</td>
<td>8,25</td>
<td>Dreissena</td>
<td>&lt;48.000</td>
</tr>
<tr>
<td>CDSC-15</td>
<td>8,90</td>
<td>Bivalves</td>
<td>&lt;48.000</td>
</tr>
<tr>
<td>CDSD-13</td>
<td>6,30</td>
<td>-</td>
<td>not dated</td>
</tr>
<tr>
<td>CDSD-13</td>
<td>8,60</td>
<td>Ostracods</td>
<td>42.357 +/- 645</td>
</tr>
<tr>
<td>CDSD-13</td>
<td>9,80</td>
<td>Bivalves</td>
<td>&lt;48.000</td>
</tr>
<tr>
<td>CDSE-12</td>
<td>5,20</td>
<td>Ostracods</td>
<td>45.130 +/- 185</td>
</tr>
<tr>
<td>CDSE-12</td>
<td>9,80</td>
<td>Dreissena + other bivalves</td>
<td>&lt;48.000</td>
</tr>
<tr>
<td>CDSF-14</td>
<td>9,60</td>
<td>-</td>
<td>not dated</td>
</tr>
</tbody>
</table>

Table 4
Radiocarbon dating results.
3.3 - Geotechnical boreholes and cone penetration tests

Three different hydrocarbon fields are being developed by AGIP KCO in the Northeastern Caspian Sea; Kashagan, Kairan and Aktote. The development of these fields takes place from artificial islands. A large amount of subsurface data has been gathered for the construction of these installations. The existing geotechnical boreholes and Cone Penetration Tests (CPTs) were all drilled to support the design and construction of offshore production and processing facilities near the artificial islands and pipeline routes in the North Caspian. The reports of these geotechnical surveys have been provided by AGIP KCO and Fugro Engineers BV to use for this project. In total, 48 geotechnical borehole reports and 40 cone penetration tests were used (Figure 9, Table 2).

The geotechnical boreholes and CPTs were all drilled in the neighbourhood of artificial islands. Therefore 7 different clusters of geotechnical boreholes and CPTs exist; 4 around the Kashagan East development near the A, C, D, F, G-islands and 1 around both the Kairan and Aktote artificial islands. Within these clusters the boreholes are closely spaced (50m to 500m), the different clusters are multiple kilometres apart. A map showing the location of the different clusters can be found in Figure 9.

3.3.1 - Geotechnical boreholes

The geotechnical boreholes are all drilled with a WIP sampler (Figure 10). After the sampler has been withdrawn from the drill-pipe, the core barrel can be unscrewed from the WIP sampler. The sample is pushed out of the core-barrel using a pneumatic device. In this way, samples of 1m can be obtained in the case of a fully recovered core. When the procedure is completed, the hole can be drilled 1m further with rotary-drilling and the next sample can be taken to get a continuous core. The depth of each 1m sample is taken by measuring the drillstring.

At the drill-site soil laboratory the soil-types are classified and samples for geotechnical tests can be taken. The geotechnical soil analysis is done according the American Society for Testing and Materials or ASTM soil classification method. This classification method is designed for geotechnical purposes. Descriptions of sedimentological features, like contacts and crossbedding, are not included in the classification. The results are presented in a geotechnical log with descriptions of the encountered strata and the results of geotechnical tests.

3.3.2 Cone penetration tests

The cone penetration test was developed to investigate soft soils for geotechnical purposes. The test uses a steel conus which is pushed into the ground with a constant velocity. During the test the cone resistance “qc” and sleeve friction “fs” are measured. Additional to these two measurements the pore pressure “u” can be measured which can be used to correct the cone resistance for pore pressure effects.

The locations of the cone penetration tests are in clusters around the artificial islands. The locations of these clusters are shown in Figure 9. Some CPT tests were taken within a couple of meters from the geotechnical cores, others stand alone. In the North Caspian Sea the CPTs are carried out with a piggyback technique. With open hole rotary drilling the desired test depth is reached. The CPT apparatus is lowered in
the drillstring and locked into position at the bottomhole assembly. When the mud pressure is increased the cone is pushed into the undisturbed soil below the drillbit.

There are multiple empirical relationships between lithology and the CPT parameters. Cheng Hou et al. (1990) was applied to a lot of different areas in the world and generally provides good results, therefore this method was applied to our CPT data first. Cheng Hou’s relation is explained below.

\[ B_p = \frac{\Delta u}{q_t - \sigma_e} \]

Where \( B_p \) is a pore pressure ratio, \( \Delta u \) the excess pore pressure, \( q_t \) the cone resistance corrected for pore pressure effects and \( \sigma_e \) the pressure of the water column above the cone.

\[ N_h = \frac{500B_p}{\log(q_t/2\sigma_e)} \]

Where \( N_h \) indicates the coarseness of the sediment

The following cut-off was applied.

For sand: \( N_h < 1 \)
For silt: \( 1 < N_h < 20 \)
For Clay: \( N_h > 20 \)

Because some of the CPTs were taken next to geotechnical cores, they could be compared to check the lithology. Cheng Hou’s method did not lead to a good match with the geotechnical cores, even after optimizing the cut-off values (Figure 12). This is probably because of a high variation in pore pressure in some of the intervals. Because this method did not work, a simple friction ratio cut-off was tried. The friction ratio was calculated, and then a cut-off was applied to distinguish the different strata. The calculation for friction ratio RF and the cut-off values used are shown below.

\[ RF = \left[ \left( \frac{f_s}{q_c} \right) \times 100 \right] \]

Then a simple cut-off was applied on the friction ratio RF.

For sand: \( RF < 2 \)
For silt: \( 2 < RF < 2.5 \)
For Clay: \( RF > 2.5 \)

Due to the large difference in sleeve friction between the clays, this method did not give satisfactory results either (Figure 12). However, the patterns of the friction ratio could be clearly recognized. These were used for the correlation of the CPTs and cores. Therefore the CPTs can only be used for correlation purposes and not for the determination of lithology.
Figure 12
Lithological log and friction ratios of borehole D53 and D54 in cluster D. The Logs on the left show the lithology as observed in the cores. The continuous log displays the friction ratio. The two discrete logs right of the friction ratio show the lithology derived from the CPT data with optimized cut-off ratios, on the left Cheng Hou and on the right the friction ratio cut-off. The derived lithology does not show a good match with the observed lithology in the cores. The friction ratio can be correlated well between the two boreholes.
3.4 - Seismics

The seismic data used for this project was provided by AGIP KCO. The data was obtained for 2 well site surveys 2km northeast and 2km southeast of Kashagan D-island. The seismic surveys were carried out with a Sub Bottom Profiler and Sparker.

A Sub Bottom Profiler can be used to identify and characterize sediments below the surface. The technique is similar to reflection seismics. A transducer emits sound pulses which are reflected against the seafloor. Some of the pulses will penetrate the seafloor and reflect against contacts between the different strata. The reflections are recorded by a receiver. The gathered data provides information about the shallow subsurface (Agip KCO 2007a).

A Sparker is an electric seismic source used for single channel reflection seismic surveys. The Sparker is fired by the release of a high voltage electrical charge in the seawater. The charge creates an ionized gas or plasma which collapses. This creates a high frequent seismic pulse in the water. The Sparker has a small size and weight which makes it very suitable for shallow marine surveys. The reflected seismic waves are recorded by a receiver which is towed behind a ship. The resolution of the data are between 10m and 120m below the seafloor (Agip KCO 2007a).

The seismic data was interpreted by Agip KCO. Channel-like features were recognized at depths between 15 and 25 m in the subsurface. Two figures are included; Figure 13 shows an example of a single-channel-seismic line, oriented N-S. Figure 14 shows a map of the interpreted channel-like features.

**Figure 13**
Seismic single channel line N-S, with two interpreted surfaces near the Kashagan field. The green surface indicates the reflector of the features interpreted as channels, the brown a deeper erosional surface. The vertical axis is indicated in two way travel time in milliseconds, but correspondents approximately with depth below seafloor in meters (Agip KCO 2007a).
3.5 - Bathymetry

A dataset with bathymetric isolines of the Caspian Sea, created by the Caspian Environment Program, was used (Kerimov 1999). This dataset was digitized in 1999 from the Navigation Map published by the Navigation and Oceanography Department of the Defence Ministry of the USSR in 1987, scale 1:1.000.000, Mercator projection and was corrected from the satellite “Resurs” made in 1997. The resolution of the image is 150m, with isolines at 0m (-29m AMSL), 5m, 10m, 20m, 50m, 100m depth and deeper at every 100m interval. The map was digitized with Arc View GIS 3.1 (ESRI). The depths of the isolines were created using the Spatial Analyst 1.1 extension of Arc View GIS. We converted the data to UTM with the PCTrans software which was developed by the Hydrographical Survey of the Dutch Royal Navy. The isolines were imported in Schlumberger Petrel reservoir modelling software as polygons.

In addition, water depths measured during the coring and CPT surveys were used. These water depths were obtained by measuring the water depth in the moonpool before drilling. By adding the Caspian sea level at the day of measurement, the water depths were converted to m AMSL. For this conversion the Topex Poseidon radar satellite Caspian sea-level curve was used, which measures the Caspian sea level every 10 days. In the shallow waters of the Northeast Caspian, sea level can differ from the mean Caspian sea level, due to wind surges. Therefore the corrected water depths are still not very accurate, and only useful when averaged. The water depths for each borehole were imported in Petrel together with the polygons from the digitized bathymetrical map (Figure 15). A surface was created between both data sources and smoothened to get a detailed and realistic seafloor map.
Figure 15
Bathymetrical map of the study area in the Northeastern Caspian Sea. The map was constructed with data from the 1987 navigation map of the USSR defence ministry and measurements taken during coring and cone penetration testing.
3.6 - Emba River fieldtrip
During the coring analysis in Atyrau, Kazakhstan a short field trip was made to the Emba river area and coastal plains south of the town of Kolsary (Figure 16). The goal of the fieldtrip was to get an impression of the depositional environments of the Emba River delta and the surrounding coastal plain. This area was flooded during the latest highstand. The former seabed now forms the coastal plain. The features and geology of this coastal plain might be a good analogue for the offshore deposits. The Emba River itself is possibly a source for sediments and might have flowed through the offshore study area during lowstands.

The main branch of the Emba River splits up in multiple branches near the town of Kolsary. The river was visited just before the point where it splits up. Two of the smaller branches were also visited. The visited locations are plotted in Figure 16. Around the Emba River vegetation consists of bushes and grasses. During the fieldtrip in October the riverbed of the main branch was dry with the exception of some pools, holding still water. The main branch is approximately 100m wide and is incised 2m into the surrounding landscape. It shows steep edges on the outer bends, and point bars on the inner bends. The steep outer river edge consists of silt and fine sands. Crossbedding and climbing ripples were visible in these steep outer bends. The bottom of the river bed consists of fine to medium sands and showed black organic bands. Vegetation was spotted in the pools with standing waters and might be responsible for the organic black bands encountered in the subsurface.

The two smaller visited branches downstream are approximately 80m and 20m wide. In these branches there was no running water present. A hole dug in the river bed showed similar sediments as found in the main stream. More organics were found in the subsurface of the deepest parts of the river, probably a direct result of more frequent presence of water in the lower parts of the riverbed.

Away from the Emba River the vegetation becomes sparser. The landscape exists of low fixed dunes and salt lakes in the interdunal areas. The bottom of these lakes contain silty clays with gypsum layers and organic bands. The edges of the lakes are bordering the longitudinal hills which have in some cases steep edges created by erosion.

An outcrop next to the road showed that the dunes are deposited on top of silty clays with gypsum layers. The dune at the outcrop was approximately 4m thick and consists of fine to medium sands. Some of the intervals showed crossbedding. Through the whole outcrop small scattered shell fragments were present.
Figure 16
Satellite image showing the visited areas during the fieldtrip. Emba 1 is located at the main branch of the Emba River, Emba 2 and 3 are the smaller branches after the split up. Gypsum indicates the spot of a visited gypsum lake and sand indicates the spot of the dune outcrop (Google Earth 2008).

Figure 17
River bed sediments at the main branch of the Emba River at location Emba 1. The river bed consists of medium and fine sands with black organic bands.
Figure 18
*Photo of one of the smaller Emba branches at location Emba 3. The channels are incised approximately 2m in the surrounding plain.*

Figure 19
*Close-up of the crossbedding found in a dune outcrop at location “Sand”. Small scattered shell fragments were present in the whole interval.*
4 Sedimentological data interpretation

4.1 - Lithofacies descriptions & depositional environments

Four different lithological facies could be distinguished in the cores; a light grey sand facies, a brown sand facies, a silty clay facies and an overconsolidated sand facies (Figure 20). Sedimentological descriptions of these lithofacies were made and described in this paragraph. With the observations, interpretations on the depositional environments for the four different lithofacies were made and also described in this paragraph. The depositional environment describes the combination of physical, chemical and biological processes under which sediments are deposited.

Table 5 gives an overview of the different lithofacies and associated depositional environments.

<table>
<thead>
<tr>
<th>zone nr.</th>
<th>facies name</th>
<th>depth below seafloor (m)</th>
<th>description</th>
<th>depositional environment</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>grey sands</td>
<td>0-1,5m</td>
<td>fine to coarse sands with shell debris and gypsum bands</td>
<td>Open sea environment</td>
</tr>
<tr>
<td>2</td>
<td>brown sands</td>
<td>1,5-4m</td>
<td>fine to medium layered sands with small shell fragments</td>
<td>barrier</td>
</tr>
<tr>
<td>3</td>
<td>silty clays</td>
<td>4-9m</td>
<td>olive grey silty clay with organic rests and gypsum bands</td>
<td>lagoon</td>
</tr>
<tr>
<td>4</td>
<td>overconsolidated sands</td>
<td>9-?m</td>
<td>fine to medium sands with small shell fragments</td>
<td>reworked aeolian</td>
</tr>
</tbody>
</table>

Table 5
Table with the 4 lithofacies and their descriptions, depositional environments and zone numbers.

Figure 20
A representative picture of the four lithofacies, from left to right; the grey sands, the brown sands, the silty clays and the overconsolidated sands. In the grey sands, a dark shell fragment layer is visible. The top section contains layered sands with gypsum, silt and clay bands. The brown sands are very uniform and show small scattered shell fragments. The silty clays show a colour change from grey to brown in the top, laminations of silt are visible as well as black organic bands and spots. The overconsolidated sands are laminated with shell rests and broken up because they were pushed out a steel core barrel.
4.1.1 Grey sand facies description

The grey sand facies consists of light grey medium to fine shelly sands, interbedded with thick fragmented shell layers and gypsum bands. The fine and medium sands are layered with silty and clayey layers with a thickness of a few centimetres. The sorting of the sands between these layers varies from very good to very poor. Shell layers are found in the top of the lithofacies and have a thickness up to 30cm. In some of these layers the shell content reaches up to 80%. Mm to cm scale gypsum laminations and small individual crystals are found in the whole interval. The facies is present in all the cores and has a thickness ranging between 0.5m and 1m. The contact between the grey sands and the underlying brown sands is horizontal and sharp (Figure 21). The contact was observed in core CDSB and CDSE.

![Figure 21](image)

The contact between the grey sands and brown sands in core E and B. The colour change is sharp, in core E a gypsum band is deposited at the contact.

4.1.2 Depositional environment of the grey sands

The grey shelly sands are the most recent deposits in the North Caspian Sea. The lithofacies was encountered in all the CDS as well as the geotechnical boreholes. A high shell content of up to 80% shows that the sediment influx was low during the deposition. The fact that most of the shells are fragmented indicates that the sediments are reworked. Waves and storm surges are the most likely processes that can rework sediments in such a way.

The sand and silt in this lithofacies is probably provided by alongshore currents and maybe reworked material or distal Ural Sediments or Volga. During the snow melt in the spring, Emba river sediments might also contribute to the sediment influx. The layering between finer and coarser sediments is probably caused by a combination of the relative position to the coast and the type and amount of sediment influx. If the relative position to the coast changes, due to small scale sea-level changes the water depth and the energetic level of the environment changes too. In shallow water close to the coastline, the energetic levels are generally higher and only coarser sediments are deposited, in a low gradient environment.
Laminated gypsum layers generally form by precipitation in a body of water. The water depth at this moment, 2m to 4m, is too deep for gypsum layers to form. They must have formed during periods when the water level was lower. This situation occurred for the last time during the 1975 lowstand. Satellite pictures made during this lowstand, show that the boreholes were located at the shoreline, surrounded with white sediments, probably gypsum (Figure 8).

The contact between grey sands and the underlying brown sands is sharp and the colour change between the two lithofacies is large. It is therefore interpreted as an erosional contact (Figure 21).

4.1.3 Brown sand facies description
The brown sand facies is layered and consists of fine to medium-fine sands with bands of small shell fragments. In some cores black organic bands, macro rests and gypsum nodules were found at different depths. The sorting of the sands varies over the lithofacies but can overall be seen as poor. The facies was found in all cores except for core CDSC-15 and has a thickness ranging between 3m and 4m. The contact between the brown sands and the silty clays below was observed in core CDSA, CDSB, CDSD and CDSE. The contact is very sharp and irregular which are signs of erosion (Figure 22). Shell lags between 1cm and 10cm thick are found above these contacts. In cores A and B two erosive surfaces were found 15cm above each other. In these cores the brown sands are eroded into grey clay with shell fragments, the grey clay is eroded into the silty clay lithofacies below (Figure 22).

The brown sand facies was not encountered in core CDSC-15. Instead of the brown sands a 1m thick grey clay interval with regular mm scale bands of shells, gypsum and silt was found (Figure 24). These clays were deposited on top of a grey fine sand layer of 0.5m thick. Below the fine sands two clay layers, both of approximately 1m thick were found. The top one has a dark colour caused by a high organic content and is laminated with numerous mm scale black bands. The lower clay interval has a light grey colour and contains shell lags. The contact with the silty clay lithofacies below is formed by a 10cm thick shell lag deposit.
4.1.4 Depositional environment of the brown sands

Medium-fine sands, are generally deposited in a high energetic environment. The sands are very continuous parallel to the coast (Figure 23, Figure 24). The layers of shell fragments and the poor sorting indicate a coastal depositional environment. The observed microfossils in this layer suggest a backshore position and deposition in a salt-marsh-like environment (Athersuch 2008). At the bottom of this interval lag deposits of shell debris were found. Lag deposits are the residual accumulation of coarse material left behind by the winnower of finer material. These types of deposits are typical for the shoreface area.

In core CDSC, which was drilled close to the shore, the sands were not encountered. Instead the top of the interval consists of silty clays with regular gypsum, shell and silt bands. This indicates a less energetic depositional environment with low water levels and occasional events that deposit shells and silt layers. The sediments in this core are interpreted as being deposited in a lagoonal environment, like the modern mudflats. This area is protected from waves by barrier islands so sedimentation of finer sediments can take place. The shell and silt layers have been deposited during storms as backbarrier deposits.

The lateral continuity of the brown sands, shoreface deposits more seaward and lagoonal deposits more landward, all indicate a barrier-like depositional environment (Figure 23, Figure 24). For a barrier-like system a high sediment supply is needed. At the present moment, the sediment influx in the system is very low. The Emba River is the closest source of sediment but has a very low output; however, it could have had a higher output in the past. The Ural delta is in the present situation too far away to supply sand to the system. However, during regressions the river incises into the Ural furrow and fills it with sediments during a subsequent transgression (Leont’ev et al. 1977). A transgressive barrier system could have transported the sediments landwards towards the migrating coastline by barrier overstep. It is very likely that the increased influx of sediments from the Ural furrow initiated the formation of barrier islands. The remnants of this system are still visible in the present day barrier islands in front of the coast.

The contact of the sands with the silty clays below is sharp and irregular, these are signs of erosion. In some cores multiple erosional surfaces where found near the contact with the silty clay facies. The erosional surface can be created by two different events; by subaerial exposure and as a ravinement surface. Subaerial exposure can be caused by a sea-level drop leaving the sediments exposed on the surface. Continental processes can erode the exposed sediments. A ravinement surface is the result of near shore erosion during a transgression in a barrier system. When a barrier system migrates landward during a transgression, the overwash deposits and the lagoonal deposits below can be eroded by shoreface processes. This system is explained in Figure 25. The eroded sediments are transported by currents and deposited on top of the ravinement surface (Cataneo 2003).
Figure 23
SW-NE borehole correlation between boreholes F14, D13, E12 and A11. This image shows the continuity of the brown sands parallel to the coast.

Figure 24
E-W borehole correlation between borehole clusters and boreholes D, F, C, B1, A11 and Kairan. This correlation shows the low continuity of the brown sands perpendicular to the coast.
Illustration of a wave dominated sediment poor barrier/lagoon system. The majority of changes occur during storms (Storms 2002). Erosion takes place near the storm wave base and mobilises sediment, the resulting unconformity is indicated in red and called a ravinement surface. Depending on their grainsize, the sediments are transported by wave generated bottom currents, distributed along the coast and deposited at the barrier front or shallow marine area. A part of the mobilized sediments might be transported by washover fans to the backbarrier. During transgressions, retrogradation of the coastline takes place and the barriers move landward.
4.1.5 Silty clay facies description
The olive brown silty clay facies is interbedded with few silt layers and bands with shell rests. Black organic bands and spots with macro rests are common in some of the intervals. Gypsum nodules and lenticular gypsum crystals are found, as well as traces of roots. Sharp and gradual colour changes occur and range from olive brown to olive grey. This facies was present in all cores and has a thickness of at least 3.5 m or more. Towards the bottom of the core the colour changes, from olive brown to olive grey. In core CDSA-11 and CDSB-1, above the contact with the overconsolidated sands, layers with black shell fragments were found. The contact between the silty clays and the overconsolidated sands was only recovered in core CDSB. The grainsize changes gradually and it is best noted from the colour change over 30cm (Figure 26).

![Figure 26](attachment:image)

The contact between the silty clays and overconsolidated sands was only found intact in core B. The colour change is gradual but the change in grain size is more abrupt.

4.1.6 Depositional environment of the silty clays
The silty clays can only be deposited in an area with a low energetic level. Shell bands and a calcareous composition indicate a marine or coastal origin. Rests of rootlets, black organic bands, black spots and macro rests are found in the whole interval. Especially the horizontal organic bands and macro rests indicate that plants grew in the area during deposition and not after deposition. The lenticular gypsum found in this layer forms when sulphate in dissolved organic material reacts with carbonate, it also indicates high organic content (Cody 1979). The microfossils found in this lithofacies are typical for a Caspian marine or marginal brackish environment (Athersuch 2008). These properties indicate a lagoonal depositional environment protected from the open sea. During storms coarser sediments can be transported into the lagoon forming silt bands. Colour changes from olive grey to olive brown point to a different organic or calcareous content. The grey colloured deposits probably formed during deposition in a deep calcareous-rich part of the lagoon, and the brown colloured deposits in the shallow vegetated parts. On the other hand, the
colour change can also be explained by changing climatic conditions causing periods with more or less vegetation.

The silty clays were found in all the 6 CDS cores. The thickness of at least 4m deposition in the same depositional environment indicates aggradation. Because there is no subduction in the North Caspian Sea this has to be the result of a sea-level rise.

The erosional surface found between the silty clay and brown sand facies can be interpreted as a ravinement surface, and as a consequence the silty clays and brown sands can be interpreted as part of the same barrier system. In this case they are deposited in the same transgressional phase. On the other hand, the erosional surface between the two lithofacies can also be interpreted as a subaerial exposure, a regressive phase between the deposition of the two facies would have occurred. This would imply a larger gap in time between the deposition of the two lithofacies. The occurrence of root rests, found in the silty clays support for the subaerial exposure hypothesis. However, they could also be deposited in a lagoonal type of environment. At this moment both options should both be equally considered.

4.1.7 Overconsolidated sand facies description
The overconsolidated sand facies is layered with medium-fine brown sands. This facies has a lot of similarities with the brown sand facies but is overconsolidated. Bands of shell fragments occur, as well as scattered shell rests. The bottom of this facies was not found. This facies was only encountered in cores CDSA-11, CDSB-1 and CDSE-12.

4.1.8 Depositional environment of the overconsolidated sands
The overconsolidated sands were only found in 3 of the 6 CDS cores and in the deep geotechnical cores. The overconsolidated sands in the cores were badly sorted and contained bands with small shell fragments. The first impression of these sands would point to a coastal depositional environment.

However, the seismic reflection of these sands in the Kashagan area and in the whole area between the Kashagan to Karain cluster, show E-W oriented channel-like shapes on maps interpreted from shallow seismics (Figure 27). Therefore these sands were previously interpreted as channel fill deposits. The seismic maps show an extremely dense channel pattern with a low sinuosity (Figure 27). To create such a pattern a very big braided river system would be required. Such a river system is not present in the area at the present moment, and it is not very likely that there was one in the past, because of the geomorphological conditions of the area.

The shapes, sizes and distribution of the channel-like patterns have much more in common with aeolian-dune deposits. Such deposits can be found in the whole onshore North Caspian region and known as Baer hills. The shapes sizes and E-W orientation of the Baer hills are similar to the channel-like features on the seismic maps (Figure 27). It is more likely that the channel-like features are deposited as longitude aeolian dunes during a lowstand on the dry seabed.

The sands are badly sorted and contain small scattered shell fragments and bands of shell fragments. Therefore, the sedimentological properties of these sands do not look aeolian. However, coastal dunefields are often totally reworked by shoreface processes. The paleotopography of the dunes disappear and the sands are
redistributed along the coast. In our case, the seismics show dune shapes but the CDS cores show coastal structureless sands. A combination of both processes seems very likely. A similar situation can be found in Jurassic formations near Ghost Ranch, New Mexico (Cheikh et al. 2000). In outcrops, aeolian dunes buried beneath subaqueous strata were found. The remnant dune paleotopography is buried by onlapping large structureless sandstones. These structureless sandstones are the result of mass wasting of the upper portions of the dunes and are deposited as sediment-gravity flows that filled the space between the dunes (Cheikh et al. 2000).

The reworking of dunes by water is also visible in our onshore analogue near the Baer hills. Here salt lakes formed in the interdunal areas, and erode the edge of the dunes (Figure 27). Small channels formed between the dunes, distributing the sands. Shells found in the cores just above the overconsolidated sands indicate a marginal salt marsh environment, such as the modern Salonchaks, which corresponds with the idea of dunes being flooded.

A reason for the conservation of the paleotopography of the aeolian dunes in the Northeast Caspian can be the gentle slope, which may cause dissipation of waves and a less energetic environment close to the shore. The overconsolidated sands are overlain by lagoonal deposits. The presence of a lagoon, protecting the dunes from reworking by coastal processes, can also help to preserve the shape of the dunes. It is thus very likely that the overconsolidated sands in the cores are originally deposited as aeolian dunes, and are partly reworked by coastal processes during a transgression.

![Figure 27](image)

*Figure 27*

*Satellite image of coastal dunes southeast of Atyrau, locally known as “Baer Hills” and a depthmap of the seismic reflections from the overconsolidated sands near the Kashagan field (modified from Agip KCO 2007a). The size, shape and orientation of both features show remarkable similarities.*
4.2 - Sequence stratigraphy

Sequence stratigraphy is used as a tool to link stratigraphy to time, sea-level change and sediment supply. It places the subsurface architecture in a chronostratigraphical context. The sequence stratigraphical framework describes genetically related units that result from the interplay of accommodation and sedimentation. These units are bounded by bounding surfaces.

4.2.1 Vertical Stacking patterns and genetic units

Because the sediments in the Northeastern part of the Caspian sea have been deposited in an environment with large shifts of the coastline and sea level changes in multiple time scales, evidence of higher order sequences can be found within the major sedimentary units, represented as alternations of thin gypsum laminations and thick clastic layers. This makes it hard to define discrete genetically related units, and even harder to separate different orders of sequences from each other. Therefore a more continuous approach was followed that uses depositional water depth instead of discrete genetically related units. The depositional water depth is here defined as the interpreted water depth, in which a sediment has been deposited.

The change in depositional water depth in a vertical succession is an indicator for base level changes and sediment supply. When the rate of deposition and rate of accommodation differ throughout time, 3 different vertical stacking patterns of sediments can form aggradation, retrogradation and progradation (Posamentier 1988, Van Wagoner et al. 1990, Catuneanu et al. 2009).

With the depositional environments of the different facies in mind, the water depths during deposition can be estimated. This was done to reconstruct tentative vertical stacking patterns and the sea level during deposition. The water depths of the different environments were estimated by comparing the present morphology of the northeastern Caspian Sea with the depositional environments found in the cores. For the present day marine environment a water depth between 2m and 3m was used, for the shoreface environment between 1.5 and 2m, for lagoonal deposition between 1 and 1.5m and for gypsum layers between 0.5 and 1m.

Evidence for changes in waterdepths is also found within the different lithofacies. Within the silty clays, colour changes between grey and brown occur. The brown collared silty clays contain overall more organic matter. The colours are changing gradually from grey to brown upward and have a sharp contact on top. The brown colour is probably an indicator for an increased organic content and the sharp contact of subaerial exposure. This in turn is an indication for changed climatic conditions or variations in water depth. Plants can only grow in shallow waters, so one can assume that the brown clays are deposited in shallow water. The depositional water depths are changing when the colour is changing. In this sense it is likely that the brown clays are deposited in a more proximal position. The colour change from grey to brown indicates progradation. Up to three of these colour sequences occur in the silty clays lithofacies. Together these units indicate aggradational stacking pattern.

In the brown sands, some subtle fining or coarsening-up sequences occur together with some sharp contacts between different colours of sand. They are not very clear and the grain size and colour changes are very subtle. The sorting of the sands also varies. The trends were not comparable and visible in all the cores, and the grains
size changes are very subtle, hence indicate changes in sediment supply and depositional environment.

In the grey sands, gypsum bands occur, indicating shallow water conditions. These gypsum layers clearly indicate a drop in base level. The gypsum layers can be interpreted as being formed near subaerial unconformities.

For every log, a graph was made showing the waterdepth during deposition as a function of the depth in the core (Figure 28). From this graph the stacking patterns in different scales can be read. When the water depth stays the same, aggradation takes place. Assuming no tectonics, the sea level is rising while the coast stays in the same position. When the inferred-depositional-water depth decreases stratigraphically upward, progradation is likely to take place. The sea level drops or is outpaced by sedimentation and the coastline moves basinward. When the depositional water depth increases upwards in a vertical deposit, retrogradation takes place. The sea level rises and the coastline moves inland.

The four main lithofacies can be recognized from the curve in Figure 28 by looking at the depositional water depths. The top 1m of the curve represents the grey sands deposited in a water depth around 2.5m with regular changes in water depth. Between 1m and 4m the brown sands have been deposited in a water depth around 2.5m, but with less sea-level changes. Between 4 and 9m the silty clays have been deposited in a water depth between 1m and 1.5m, the changes here were interpreted from the colour changes. Below 9m the overconsolidated sand were observed in 3 of the 6 boreholes. The depositional trends were interpreted for all cores in 3 different orders. A schematic overview of these interpretations can be found in Figure 31.
Figure 28
Graph showing the water depths during deposition for all the CDS cores based on the sedimentary environments. The spikes in the top 1m indicate gypsum bands. The top part of the cores has been deposited in a marine and near shoreface setting, below 4m the silty clays are deposited in a shallow-lagoonal environment. Below 9 m the overconsolidated sands were deposited during a lowstand in subaerial conditions.

With the depositional water depth and the depth in the core, the Caspian sea level could be calculated as a function of the core depth. The depositional water depth was added to the depth in the core relative to mean sea level. A plot of the Caspian sea level versus the depth in the core was made in this way (Figure 29). This graph illustrates that almost all sediments are deposited during base level rise. This means that during regressions no sedimentation or even erosion occurred.
Figure 29
Graph showing the Caspian sea level during the deposition of the sediments, as a function of the core depth, based on the sedimentary environments and water depths during deposition and assuming no subduction. The graph shows that the overall trend during deposition was transgressive.

4.2.2 Sequence-stratigraphic surfaces
Figure 29, illustrates that the sediments are almost all deposited during base-level rise. This supports the idea of the sediments being deposited in a transgressive depositional system. However, it is not possible that all the sediments are deposited in a single phase of base-level rise. We know that the Caspian sea level has not been rising continuously over the past 48,000 years. Major regressions occurred, and must have caused falls of the base-level. These sea-level cycles are not completely preserved in the sedimentary record, and are separated by phases of no deposition or erosion. This means that the sediments are deposited during different base-level cycles and therefore not genetically related to each other. Different genetically-related sediment bodies are separated by sequence boundaries, which are represented by contacts between the different lithofacies. This concept is explained in Figure 30.
Figure 30
This figure illustrates the concept of a sequence boundary separating transgressive deposits in the case of a barrier-lagoon complex. Initially a barrier-lagoon complex is deposited during a transgression. During a subsequent regression the deposits are subaerial exposed and eroded. During a following transgression a new barrier-lagoon complex is formed on top of the erosional surface which forms a sequence boundary. Only transgressive deposits are preserved in the stratigraphical record.

The overconsolidated sands are interpreted as aeolian sands that are partly eroded and reworked during a transgression. The contact with the silty clays above is conformable. The silty clays are deposited in a lagoonal environment. The surface between the overconsolidated sands and silty clays is therefore interpreted as a maximum regressive surface. This surface is diachronous and younging towards the basin’s margin (Catuneanu et al. 2009).

The next major shift in facies occurs between the silty clays and the brown sands. There are two possible interpretations for this surface; either a transgressive ravinement surface or a subaerial exposure. The erosive surface is correlatable over a very large distance, which is not conventional for a ravinement surface. However, the area has a very low gradient, which might make the ravinement surface continuous over large distances. If the surface is a ravinement surface, it is diachronous and becoming younger towards the basins margin. If, on the other hand the surface is a subaerial exposure it is a sequence boundary, which means that the silty clay and brown sands have been deposited in a different sequence (Catuneanu et al. 2009).

Above the brown sands the grey sands are deposited. The contact between the brown sands and the grey sands is most likely erosional. The depositional environment of the grey sand facies is interpreted as a low sediment influx marine environment with large vertical facies shifts. Between a barrier deposit and a shallow marine deposit one would expect a correlative surface, marking a gradual change.
from shoreface to marine deposits. However, this is not the case here. There is a sharp contact between the deposition of the brown sands and the grey sands caused by a strong change in sediment supply. This is also indicated by the colour difference between the two sands. The two lithofacies are probably not genetically related to each other. Although there are not much signs of subaerial exposure, the surface is interpreted as a subaerial unconformity. The erosion can be created by coastal erosion, which does not leave clear marks of subaerial exposure. Following this interpretation a hiatus between the deposition of the brown sands and the grey sands is likely present.

4.2.3 Sequences

A sequence is defined as a succession of strata deposited during a full cycle of change in accommodation or sediment supply (Catuneanu et al. 2009). With the sequence stratigraphical surfaces in mind, we can conclude that the sediments in the cores are deposited during at least three and maybe four large scale sequences, depending on the interpretation of the boundary between the silty clays and brown sands. The aeolian sands in the bottom are deposited during a regression in the first sequence. The overlaying silty clays are deposited during a transgression in a subsequent sequence, the brown sands are deposited in a phase of transgression of the same sequence or the next sequence and finally the grey sands are deposited during a transgression of the last new sequence (Figure 31).

Evidence of smaller scale more frequent sequences was found within the different lithofacies, the gypsum layers, colour changes of the clay and shell bands are all indications of higher order cycles of relative sea-level change. They result in rapid vertical facies shifts within the main lithological units.
Figure 31

The CDS cores with interpreted depositional trends; transgressive in blue, regressive in red and aggradation in green. The trends are interpreted in 3 different orders; Order 1 the largest, at the scale of the lithofacies, Order 2 within lithofacies and smaller-scale Order 3. The sequence boundaries are indicated as SB and the possible ravinement surface as RS.
4.2.4 Sequences and the Caspian sea level

The C14 dates and sequence-stratigraphical interpretations can be compared with the known Caspian sea-level fluctuations (Dumont 1998, Kroonenberg 2008, Mamedov 1997 and Rychagov 1997). Caspian sea-level curves from literature exist in multiple timescales. The sediments in the study area are deposited between at least 48,000 years BP and present. The sea-level data from Rychagov (1997) and Kroonenberg (2008) were combined to get the resolution of this period (Figure 32). Before the Late Khvalyn highstand around 16,000y BP, the curve is less detailed. There is a large uncertainty about the timing of the highstands and lowstands during this period as was explained in chapter 3. The Inter-Khvalyn lowstand was dated around 30,000 BP and the Inter-Khvalyn regression has been estimated to take place between 40,000 and 70,000 years ago.

Our best data points for the curve are the ones dated with C14 (Table 9). The depositional water depths of these deposits are estimated and converted to meters AMSL. These points can be plotted in the graph and represent the high confidence points (Figure 32).

The overconsolidated sands were probably deposited during a lowstand, when the sea level was below their present depth in the core. One C14 analysis obtained from marine molluscs shells in the overconsolidated sands yielded a date older than 48,000y BP. Molluscs do not originate in aeolian sands, this date refers to the time when the sands were already partially eroded and reworked by coastal processes.

Most of the C14 samples came from the silty clays. It is assumed that the silty clays are genetically related and deposited in a single major transgressional event. In core CDSE-12 two dated samples in the silty clay lithofacies were taken 1,65m apart from each other, with a time interval of 2595 years. These two dates were used to estimate the depositional rate during the deposition of the lithofacies; 1.65m/2595y gives a depositional rate of 0,635mm per year. This depositional rate was extrapolated, with the dated C14 dates as anchor points, for all the cores.

The unconformity between the silty clays and brown sands marks a hiatus. If the time gap is large, it is likely that the surface is a subaerial unconformity, when the time gap is smaller it probably represents a ravinement surface, depending on how much has been eroded away (Figure 25). The extend of the surface is also important to consider. A subaerial unconformity would be noticed throughout the whole basin, while a ravinement surface is created locally. A ravinement surface would imply that the silty clays and brown sands are deposited in the same transgressional period. One date from the brown sands was known in core CDSB-1. This date is however questionable, it is dated from reworked shell fragments at 27,764y BP. The sample taken just 35 cm higher in the same facies had a C14 age of 390 years and was too young to calibrate. It is however plotted in Figure 32, and fits with the extrapolated depositional rate of the silty clays.

The grey sand facies is deposited recently. The exact sedimentation rates are unknown and the sediments are not dated. So it is hard to say when the deposition of this lithofacies started. The erosional contact with the brown sands shows a period of no deposition or erosion. If the brown sands are deposited during the Late-Khvalyn transgression, it is very likely that the contact marks the subsequent Mangyshlak regression. The deposition of the grey sands would then have started at
the beginning of the Novo-Caspian transgression. The gypsum bands are then deposited during the regressions in the Holocene. If it is correct that the contact between the brown sands and the grey sands represents the Mangystal regression, approximately 1m of sediments would have been deposited during the Holocene.

When comparing the Caspian sea-level curve known from literature, in Figure 32, with the points from our data, a number of things can be noted. The silty clays are all dated between <48.000y BP and 42.357y BP. According to the existing Caspian sea-level curve this is during the Inter-Khvalyn regression. In our interpretation, the silty clays are deposited during a transgression. The underlying overconsolidated sands are according to our interpretation, deposited during a lowstand and flooded before 48.000y BP. This raises some questions; was the Inter-Khvalyn lowstand and the beginning of the Late-Khvalyn transgression earlier than known from literature, before 48.000 years ago (Figure 33)? However, an alternative explanation would be to place a lowstand around 20.000 years ago and to include a transgressive phase before this lowstand (Figure 34).

An important question is whether the silty clays and the brown sands are deposited in the same transgressional phase. In other words, is the contact between these facies a ravinement surface or a subaerial unconformity? If it is a subaerial unconformity representing a regression, a lowstand could have taken place between the deposition of the two facies, around 30.000BP (Figure 35). If it is a ravinement surface, it is more likely that the lowstand occurred before the deposition of the silty clays, around 60.000 BP or after the deposition of the brown sands, around 20.000BP.

<table>
<thead>
<tr>
<th>Borehole</th>
<th>Sample depth (m b.s.f.)</th>
<th>Present water depth (m)</th>
<th>Water depth during deposition (m)</th>
<th>14C age (cal. Y BP)</th>
<th>Caspian sea level during deposition (m AMSL)</th>
</tr>
</thead>
<tbody>
<tr>
<td>CDSA-11</td>
<td>7,1</td>
<td>2,4</td>
<td>1</td>
<td>44.186 +/- 782</td>
<td>-36,5</td>
</tr>
<tr>
<td>CDSA-11</td>
<td>8,75</td>
<td>2,4</td>
<td>1,5</td>
<td>46.781 +/- 1.134</td>
<td>-37,65</td>
</tr>
<tr>
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<td>3,6</td>
<td>2,5</td>
<td>Modern</td>
<td>-30,2</td>
</tr>
<tr>
<td>CDSB-1</td>
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<td>3,6</td>
<td>2,5</td>
<td>27.764 +/- 185</td>
<td>-30,55</td>
</tr>
<tr>
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<td>3,6</td>
<td>1,5</td>
<td>&lt;48.000</td>
<td>-38,95</td>
</tr>
<tr>
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<td>2,7</td>
<td>1,5</td>
<td>&lt;48.000</td>
<td>-37,45</td>
</tr>
<tr>
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<td>1</td>
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</tr>
<tr>
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<td>1,3</td>
<td>Not dated</td>
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<tr>
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<td>-39,8</td>
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<tr>
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<td>-41</td>
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<tr>
<td>CDSF-14</td>
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<td>3,2</td>
<td>1,5</td>
<td>Not dated</td>
<td>-39,3</td>
</tr>
</tbody>
</table>

Table 6
Table showing the water depths needed for the calculation of the Caspian sea level during the deposition of the dated samples. The present Caspian sea level is -28m AMSL.
Figure 32
Composed Caspian sea-level curve (grey line). The black dots show the C14 dated deposits from this research. The red dots indicate the deposits which were too old for dating and are at least 48,000 years old (after Dumont 1998, Kroonenberg 2008, Mamedov 1997 and Rychagov 1997).
Figure 33
Possible Caspian sea-level curve with an early Inter-Khvalyn lowstand, around 60,000y BP. The contact between the silty clays and brown sands is interpreted as a ravinement surface. The coloured lines show the extrapolated not dated depositions from our cores; they are extrapolated with a constant depositional rate. The three different phases of deposition are indicated in grey (after Dumont 1998, Kroonenberg 2008, Mamedov 1997 and Rychagov 1997).

Figure 34
Possible Caspian sea-level curve with a later Inter-Khvalyn lowstand around 20,000y BP. The contact between the silty clays and brown sands is interpreted as a ravinement surface. The coloured lines show the extrapolated not dated depositions from our cores; they are extrapolated with a constant depositional rate. The three different phases of deposition are indicated in grey (after Dumont 1998, Kroonenberg 2008, Mamedov 1997 and Rychagov 1997).
Figure 35
Possible Caspian sea-level curve with a transgressive phase between the Early-Khvalyn highstand and Inter-Khvalyn lowstand. The contact between the silty clays and brown sands is interpreted as a subaerial unconformity surface. The coloured lines show the extrapolated not dated depositions from our cores; they are extrapolated with a constant depositional rate. The three different phases of deposition are indicated in grey (after Dumont 1998, Kroonenberg 2008, Mamedov 1997 and Rychagov 1997).
5 Modelling

To visualise the available data and test the developed depositional model and sequence stratigraphic framework, two modelling approaches were used; a process based response model and a static 3D-subsurface model were built.

5.1 - Process-response modelling

To evaluate whether the developed depositional models are physically possible BARSIM modelling software was used. BARSIM is a process based response model that simulates coastal evolution and deposition of multiple grain-size classes (Storms 2002). The model simulates the development and stratigraphy of wave dominated shorelines in two dimensions by numerically calculating erosion and deposition during sea-level change. The result shows the stratigraphy in a cross section of a coast. Key parameters in this simulator are slope, sea-level curve, sediment supply and wave height.

The main goal of the BARSIM simulation was to test if the brown sands and silty clays could have been deposited in the same transgressional sequence with an erosive contact in between. And if not, what kind of processes created the stratigraphy observed in the cores. This is important to determine whether or not a regression occurred between the deposition of both deposits. This was achieved by alternating the key parameters in BARSIM.

5.1.1 Input parameters

Because there are no tides in the Caspian Sea, initially the tidal movement was set to zero. In the simulator however, tides transport the fine sediments into the lagoon. When the tides are set to zero there is no deposition in the lagoon. In the Caspian Sea, sediments are transported into the lagoon during storm surges. Therefore the tidal movement in the simulator was set to 1m and does not simulate the tide but the frequent storm surges. The slope in the model was set to 20m over a 100km distance, which corresponds with 0,011°. Wave dissipation is not included into the model. The result is that the influence of waves might be overestimated during the simulations. The model’s standard values for the grain size distribution were used.

5.1.2 Simulation results

The first simulation was run with a linear sea-level rise of 25m in 10.000 years, with varying rates of sediment supply. Four different rates of sediment supply were used; Case A 0 m2, Case B 5 m2, Case C 10 m2 and Case D 15m2 per 1000 years. Figure 36 shows the result of the simulations after 10.000 years.

Case A and B show a retrograding barrier system. The sediment supply can not keep up with the increased accommodation space. The first barrier is overstepped and a second barrier is formed 40km closer to the coast, with a small and shallow lagoon. In front of the barrier the initial profile and the first barrier are eroded. In this retrograding system the barrier sands and lagoonal deposits are not preserved, but reworked, and a ravinement surface is formed.

Case C and D show an aggradational barrier system. The sediment supply is high enough to stabilize the barrier during sea-level rise. The lagoon widens and is filled with a thick package of lagoonal sediments. In front of the barrier, shoreface sediments are deposited. The simulation shows that for the deposition of a thick and
extensive lagoonal facies, a large sediment influx is needed. The barriers developed in this simulation are narrow, perpendicular to the coast, and do not have an erosive base. It is very unlikely that the barrier island is preserved during a subsequent regression.

To test the effect of a subsequent regression on the barrier islands a second simulation was performed. In this simulation a transgressive and regressive phase are simulated. After a 25m sea-level rise in 10,000 years a 25m forced regression of 10,000 years is simulated. This was again done with 4 different rates of sediment supply; Case A 5 m², Case B 10 m², Case C 15 m² and Case D 20 m² per 1000 years. The results after 20,000 years and the used sea-level curve are shown in Figure 37.

With a low sediment supply in Case A the initial barrier is not preserved, and a second barrier is formed close to the shore during the transgression. During the regression onlapping shelf deposits are deposited in front of the barrier.

With a higher sediment supply in cases B, C and D, the initial barriers keep up with the sea-level rise for a longer time. Aggradation occurs and the lagoons get wider. As a result a larger part of the barrier and lagoon deposits of the initial barrier are preserved. During the ongoing sea-level rise the first barrier is overstepped and partly eroded. On top of this erosional ravinement surface onlapping shelf deposits are deposited during the forced regression.
Figure 37
Simulation results of a 25m transgression followed by a 25m regression in 20,000 years in a low gradient environment with different sediment supplies. A) 5 m$^2$, B) 10 m$^2$, C) 15 m$^2$ and D) 20 m$^2$ per 1000 years. The used sea-level curve and a stratigraphic column, at the location of the red line in simulation D, are also plotted.

To test what the effects of a transgression followed by a forced regression and a new transgression are, another simulation was performed. Here a 25m transgression was followed by a short 25m regression and a new 25m transgression in 10,000 years. This was again done with 4 different rates of sediment supply; Case A 5 m$^2$, Case B 10 m$^2$, Case C 15 m$^2$ and Case D 20 m$^2$ per 1000 years. The results after 20,000 years and the used sea-level curve are shown in Figure 38.

With a low sediment supply in Case A and B, the initial barrier is again not preserved, and a second barrier is formed closer to the shore during the transgression. During the regression onlapping shelf deposits are deposited in front of the barrier. These are reworked again during the final transgression.

With an increasing sediment supply in cases C, D and E, initially the barriers keep up with the sea-level rise for a longer period. The barriers build up a wider lagoon, but are overstepped and partly eroded. The regression occurs faster than the one in the previous simulation. As a result, less sedimentation of onlapping shelf deposits takes place. During the final transgression a new barrier complex is formed on top of the initial partly eroded barrier complex, which is again overstepped and partly eroded. The result is a sediment package of lagoonal clays with extensive sand deposits on top, separated by an erosive surface.
Figure 38
Simulation results of a 25m transgression followed by a fast 25m regression and a second 25m transgression in 20,000 years in a low gradient environment with different sediment supplies. A) 0 m², B) 5 m², C) 10 m², D) 15 m² and E) 20 m² per 1000 years. The used sea-level curve and a stratigraphic column, at the location of the red line in simulation D, are also plotted.

5.1.3 Process response modelling conclusions
The process based response model shows the effects of different sediment supply rates and sea-level curves on a low gradient barrier coast. A large sediment influx is needed to create aggradation of the barrier and extensive lagoonal deposits. The barriers are narrow and perpendicular to the coast. The effect of the low gradient is that the barriers overstep over larger distances, in this simulation over 40km. Erosion in front of the barrier takes place and leaves an extensive ravinement surface. During forced regressions onlapping shelf deposits are deposited. During short regressions less sediments are deposited. During a second transgression, new barriers are deposited on top of the older eroded strata. In the final simulation barrier sand was deposited on top of eroded lagoonal facies from the previous transgression.

The stratigraphy found in the cores from the Northeastern Caspian Sea show eroded lagoonal facies with extensive barrier sands on top. The process based simulation shows that it is very likely that the browns sands and silty clays have been deposited during different transgressions. It is unlikely that these two lithofacies have been deposited during a single transgression.
5.2 - 3D-subsurface model

To make a static subsurface model, Schlumberger’s Petrel software was used. This software is designed to assemble oil reservoir data from different sources and build reservoir models. In this case the program was used to create a static 3-dimensional subsurface model to visualize the results of the core analysis and sequence stratigraphy in 3 dimensions and evaluate the depositional model. The modelling procedure starts with the import of data and well correlation, followed by the creation of a 3-dimensional grid. The procedure is finalised with facies modelling, to populate the grid with sedimentary bodies.

5.2.1 CDS borehole correlations

The logs of the 6 new cored CDS boreholes (liner cores) were imported in Petrel as bitmap logs. A discrete lithological log was created and the lithologies were digitized. The lithological logs contain the following strata; shell debris, coarse sand, medium coarse sand, medium fine sand, fine sand, coarse silty sand, fine silty sand, clayey sand, sandy silt, silt, clayey silt, sandy clay, silty clay and clay. The main sequence stratigraphical surfaces were correlated between the boreholes, which are between 10km and 12km apart. Four different horizons were used; the seabed, the top of the brown sands, the top of the silty clays and the top of the overconsolidated sands. The horizons separate the four described lithological zones. The correlation panels of all the logs can be found in the Appendix.

5.2.2 Geotechnical borehole correlations

The reports of the geotechnical boreholes consist of a geotechnical log with the depths and description of the different observed strata. The CDS cores are described in more detail than the geotechnical logs. Therefore differences in resolution and interpretation of soil textures occur between the two different types of logs. For each borehole a discrete lithological log was created with the same lithological discrete facies as for the CDSE boreholes.

The boreholes around the Kairan and Aktote fields are located near the CDSE boreholes and were correlated first (Figure 9). The 4 horizons could be well recognized.

After the Karian and Aktote clusters, the nearest cluster of geotechnical boreholes was cluster C, 20km apart from the CDS boreholes. The overconsolidated sands were only found in the deepest borehole of cluster C and are located at 13m below seafloor. In the CDS, Karian and Aktote boreholes the overconsolidated sands were found at a depth of approximately 9m below seafloor. So the top of the overconsolidated sands is deepening towards the west. The top of the silty clays was very clearly recognizable on all the logs because there is a large contrast in grain size between the two lithofacies. It was harder to differentiate the grey sands from the brown sands in the geotechnical logs. The difference between these two zones is best noticed from the colour change which was not always reported in the geotechnical logs.

In cluster G, 9km to the west of cluster C, the overconsolidated sands were found in the deepest core at a depth of 13m below the seafloor (Figure 9). The boundary between the silty clays and brown sands was still very well recognizable from the lithology. The top of the brown sands was harder to correlate. No clear colour
change or lithological change was described in the geotechnical cores. Where possible the top of the brown sands was correlated. The correlation panels of all the logs can be found in the Appendix.

5.2.3 CPT correlations

For the clusters D and F also digital CPT data was available. We received the CPT data in ASCII (American Standard Code for Information Interchange) format. The cone resistance, sleeve friction, friction ratio and pore pressure were imported in Petrel as continuous logs. Empirical relations to calculate lithology directly from the CPT data did not work as described in Chapter 4. The patterns in the CPT logs were a good tool to correlate the logs. Especially the friction ratio was used for correlation purposes. The boreholes with both CPT and a geotechnical logs were already correlated. After this was done the logs with only CPT data were correlated between the boreholes. Figure 39 shows a well section with logs from cluster F. The correlation panels of all the logs can be found in the Appendix.

Figure 39
Well section showing the borehole logs in cluster F with the 4 interpreted horizons, sea bed, top brown sands, top silty clays and top overconsolidated sands. The boreholes with available geotechnical logs A11 and A12 were correlated first. The patterns of the friction ratio could be well correlated between all the boreholes. The silty clays are thicker in the boreholes of cluster F than in borehole CDSB-1.
5.2.4 Surfaces and zones

Now the correlations of the 4 interpreted horizons were made, surfaces could be created in Petrel to show the correlated horizons in three dimensions. A boundary around the study area was made with a square polygon (Figure 9). Surfaces were created at the seafloor, the top of the brown sands, the top of the silty clays and the top of the overconsolidated sands from the interpreted horizons (Figure 41 and Figure 42). The algorithm used to create the surfaces is convergent interpolation. This method is a control point orientated algorithm which will converge upon the solution iteratively adding more and more resolution with each iteration. This means that general trends are retained in areas with little data while detail is honoured in areas where the data exists (Petrel 2008). This method was used in this model, because there is a high data resolution in the geotechnical borehole clusters, and no data between the clusters. The result of the interpolation keeps the high resolution in the clusters and maintains a general trend between boreholes located further apart. The resulting surfaces were smoothed, because in the areas with a lot of boreholes the resolution was too high what resulted in peaks in the surface which were not desired in the final model. These peaks were probably caused by measurement errors, small sea-level changes and local differences in sea level during water depth measurements and small scale seabed relief.

Figure 40
3D image of the 4 non-smoothened surfaces. This close-up near D cluster shows the high resolution in the clusters and local peaks, which were smoothened. The depth is exaggerated 500 times.
Figure 41
3D-view of the 4 smoothened surfaces created from the correlated logs and CPTs. Depth is exaggerated 500 times.
5.2.5 Gridding and upscaling

Between the surfaces a grid was made with a top limit at the seafloor and a bottom limit at 20m below sea level. The bounding polygon defines the spatial limits of the model (Figure 9). The gridblocks have a size of 50m × 50m which provide enough resolution between the closely spaced boreholes. The 5 surfaces divide the model in 4 different zones representing the main facies; Zone 1, the grey sands; Zone 2, the brown sands; Zone 3, the silty clays and Zone 4, the overconsolidated sands. In each zone, layers were created for property modelling. The number of layers depends on the thickness of the thinnest bands and the total thickness of the layer. The grey-sands zone is just 1m thick on average. The vertical homogeneity is very low. Therefore 10 different layers were created in this zone, each having a thickness of 0.1m. The brown-sand zone, silty-clay zone and overconsolidated-sand zone are thicker than the grey-sand zone. There are some thin shell layers which are
important for the model. To create a good resolution and to keep the thin shell layers in the model 30 layers were created in each of the 3 facies.

Each gridcell in the final model will have a single value for lithology. To match the well logs with the grid model, the logs need to be upscaled. This was done with the ‘most of’ method within the cells. This method selects the discrete value which is most represented in the log for each particular borehole. The result honours the resolution of the available data.

<table>
<thead>
<tr>
<th>Zone</th>
<th>Lithology</th>
<th>Number of layers:</th>
</tr>
</thead>
<tbody>
<tr>
<td>Zone 1</td>
<td>Grey sands</td>
<td>10</td>
</tr>
<tr>
<td>Zone 2</td>
<td>Brown sands</td>
<td>30</td>
</tr>
<tr>
<td>Zone 3</td>
<td>Silty clays</td>
<td>30</td>
</tr>
<tr>
<td>Zone 4</td>
<td>O.C. sands</td>
<td>30</td>
</tr>
</tbody>
</table>

Table 7

*Table with the number of layers for each zone.*
5.3 - Data analysis

After completing the structural model, which is a framework for the facies model, the next step is to populate the model with lithological properties. For proper facies modelling knowledge of the spatial distribution of the lithologies is necessary. This knowledge is partly derived from the conceptual model and partly from the data itself. The spatial characteristics of the lithological data were analysed with variograms, vertical distribution curves and thickness maps within each zone.

5.3.1 Variance

A variogram is a graph that describes the natural variation present in the data in a specific region. The semi-variance is defined as the expected squared increment of the values between two locations. If the semi-variance is calculated for a large amount of data, a variogram can be made, with on the x-axis the distance between the two data points, and on the y-axis the semi-variance. From a variogram the range, which is the distance beyond which point data points no longer have any similarity, the nugget which represents the variance where the distance is zero and the sill which is the variance where the variogram flattens out to random similarity can be read.

Because geological data are usually anisotropic, variograms can be calculated in different directions. In Petrel it is possible to calculate the variance in three directions, in two directions in the horizontal plane and one in the vertical direction. The sample pairs for the variograms are chosen in the preferred direction by using a "search cone". Because the continuity is the best parallel to the coast, the major variogram direction was chosen to be 32 degrees compass angle (NNE). The minor direction is automatically perpendicular to this direction. Variograms in all three directions were made for every lithology in all the zones. An exponential or Gaussian curve was fitted to the datapoints. Figure 43 shows the variogram of shell debris in Zone 2. A list of all the ranges and nuggets is presented in Table 8. These curves were later used for the facies modelling to calculate the spatial distribution of the modelled lithofacies. Because object based modelling was used for the overconsolidated sands and it was only penetrated in a few boreholes, no variograms were made for this zone.
Figure 43
Variogram in the major direction (32 degrees) of shell debris in Zone 2. The semivariance of the different bins are plotted as grey dots. The blue line represents the interpreted variogram. The bars indicate the amount of sample pairs used to construct the semivariances. The variogram has a range of 804m and a nugget of 0.

<table>
<thead>
<tr>
<th>Zone</th>
<th>Lithology</th>
<th>Nugget</th>
<th>Major Range (m)</th>
<th>Minor Range (m)</th>
<th>Vertical Range (m)</th>
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<td>12695</td>
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<td>Sandy silt</td>
<td>0,00</td>
<td>903</td>
<td>720</td>
<td>2,3</td>
</tr>
<tr>
<td>2</td>
<td>Silt</td>
<td>0,00</td>
<td>560</td>
<td>500</td>
<td>1,3</td>
</tr>
<tr>
<td>2</td>
<td>Sandy clay</td>
<td>0,00</td>
<td>5270</td>
<td>3440</td>
<td>2,4</td>
</tr>
<tr>
<td>2</td>
<td>Silty clay</td>
<td>0,00</td>
<td>30455</td>
<td>27980</td>
<td>3,6</td>
</tr>
<tr>
<td>2</td>
<td>Clayey sand</td>
<td>0,00</td>
<td>226</td>
<td>30</td>
<td>3,6</td>
</tr>
<tr>
<td>3</td>
<td>Silty sand, fine</td>
<td>0,06</td>
<td>365</td>
<td>351</td>
<td>1,2</td>
</tr>
<tr>
<td>3</td>
<td>Sandy silt</td>
<td>0,12</td>
<td>395</td>
<td>380</td>
<td>1,9</td>
</tr>
<tr>
<td>3</td>
<td>Sandy clay</td>
<td>0,02</td>
<td>1512</td>
<td>1005</td>
<td>1,2</td>
</tr>
<tr>
<td>3</td>
<td>Silty clay</td>
<td>0,41</td>
<td>43100</td>
<td>31681</td>
<td>1,9</td>
</tr>
</tbody>
</table>

Table 8
Variogram properties for each facies in each zone.
5.3.2 Vertical lithology distribution

Vertical variations in lithology were observed within the facies in the different boreholes. To see if the vertical lithological variations are random or show trends, the proportion of each lithology was calculated in every layer within the different zones. This was done for all the boreholes together. The result is a vertical lithological distribution curve (Figure 44). Towards the top, from layer 50, in the silty clays the silt and sand content increases. This probably indicates an increase of washover deposits towards the top and supports the idea of a retrograding lagoon being overridden by a barrier. There is an increase of shell fragments towards the top of the grey sands, this might indicate more wave activity, lots of time to concentrate the shells, and sediment starvation. The lithology of the brown sands seems to be homogenous distributed in a vertical direction. The curve of the overconsolidated sands below layer 78 does not have enough datapoints, therefore below this layer the curve is not representable. Above layer 78 the overconsolidated sands seem to be homogenous distributed in a vertical direction. The vertical distribution of lithologies were used to distribute the lithologies during the facies modelling.
Figure 44
The vertical lithological distribution curves of the 4 different facies. The vertical axis shows the layer number of in the model which increases with depth. The horizontal axis shows the proportion of each lithology. The total thickness of each facies ranges in the model between the following values; Grey sands (0,8-1,2m), Brown sands (1,4-3,8m), Silty Clays (3,8-5,5m) and the Overconsolidated sands (5m). The curve of the overconsolidated sands is only representable above layer 78, below this layer there are not enough datapoints.
5.3.3 Lateral lithological distribution

Thickness percentage maps for each lithology in every facies were created to search for trends. This was done by making a thickness well attribute for each lithology in all the boreholes. The thickness of the lithologies was converted to points. These points were used to create surfaces with convergent interpolation. The thickness maps were divided by the total thickness of the zone, to create thickness percentage maps. This was done for each lithology in all the facies. Maps of these surfaces indicate the thickness of a lithology in a specific facies. Although these maps are the result of interpolations between the boreholes, they indicate lithological trends in a horizontal direction. To distribute the lithologies, the thickness maps were used during the facies modelling as trend. The summed thickness maps for sand clay and are shown in Figure 45, Figure 46 and Figure 47.

One of the most notable trends became visible when the thickness of all the sands was calculated for zone 2 (Figure 46). In this zone, most of the sands are concentrated in a line parallel to the coast on the west side of the CDS boreholes. During the interpretation of the cores it was already noticed that borehole CDSC-15 contains backbarrier deposits, while the cores to the west contain barrier sands. It is very likely that this concentration of sands parallel to the coast indicates the remains of a barrier. Also within the silty clays a trend of more silt was recognized basinward, which are interpreted as overwash deposits from a barrier nearby. The distribution of sand and clay in the grey sands has the highest continuity parallel to the coast, with an increase of sand basinward.
Figure 45
Thickness maps of zone 1, the grey sands. Sand was observed in parallel to the coast. The clays are located closer to the shore.
Figure 46
Thickness maps of zone 2, the brown sands. The concentration of sands parallel to the coast indicates the remains of a barrier complex. Towards the east in the clay content increases and towards the west the silt content increases.
Figure 47
Thickness maps of zone 3, the silty clays. No sand was observed in this zone. Silt was observed in the northeastern part of the study area.
5.4 - Facies modelling

After a 3-dimensional grid model with layers was made and the lithological data was analyzed, facies modelling was used to “fill in” the gridblocks with strata. The facies modelling was done in such a way that the observed strata in the upscaled well logs, the variograms, the vertical distributions and the thickness maps were honoured. Each zone was deposited in a different depositional environment. Therefore it is important that each zone is modelled separately.

Facies modelling can be done with either deterministic or stochastic methods. Deterministic facies modelling has the advantage that the output will be the same after each run. The disadvantage is that with little data the result is too continuous and does not represent the natural variation created by sedimentological processes. Therefore modelling was done with stochastic modelling methods. These methods use a random seed, which means that every outcome will be slightly different, even with the same input data. The resulting models do not represent the real subsurface tough honouring the well data used, but give a better indication of the variation encountered in the subsurface.

5.4.1 Overconsolidated sands

The overconsolidated sands were modelled with object based modelling. Object based modelling populates a facies model with stochastically generated and distributed objects. The geometrical properties, like size and orientation, can be assigned with a statistical distribution. This method is especially suited for this zone because the zone consists of longitudinal dune features and a background of silty clays. Also the geometrical properties of the dunes are known. Their exact position is however not known, except for the areas with seismic surveys and well locations. The modelling algorithm therefore places the object first at the well locations, and fills the area between the boreholes with randomly placed objects, but honours the total percentages.

The seismic maps (Figure 14, Figure 27 and Figure 50) of this facies and the satellite images of the aeolian sands onshore were used to determine the statistical distribution of the shape, sizes and orientations of the dune features. The seismic maps of the “channel shaped features” were used to make a distribution of the widths and lengths of the dunes. The histograms show that the length and width as a fraction of the length are nearly normally distributed (Figure 48). The average length of the dunes, measured from the seismic maps, is 777m with a standard deviation of 435m. The average width of the dunes is 186m with a standard deviation of 98m. The thickness of the sands found in the cores was estimated between 5 and 10m. The orientation is east-west with a deviation of 20 degrees. The silty sand percentage in this facies found in the cores is 22% and the silty clay facies around 75%.

In the object based modelling option in Petrel these values were used to model rounded ellipses of silty sand with a background of silty clays (Figure 49). A map of the top of the silty sand objects was made to compare the modelled shapes with the seismic depth maps (Figure 50). This map shows that the modelled objects match well with the shapes seen on the seismic map.
Figure 48
Histograms of the length and width distribution of the “channel features” observed in the seismic depth maps. The width is indicated as a fraction of the length.
5.4.2 Silty clays

The silty clays are deposited in a lagoonal environment. The main lithofacies in the silty clay zone are silty sands, sandy silt, sandy clay and silty clay. To model this zone, truncated Gaussian simulation was used. Truncated Gaussian simulation is a stochastic modelling technique whereby the result is dependant on the upscaled well...
log data, the defined variograms, trends and a random seed. This method was chosen because it is the stochastic modelling algorithm which honours the observed spatial variations and horizontal and vertical trends found during the data analysis. The result has a distribution which is typical for the real case.

The variograms were made during the data analysis step for every lithology within the silty clay zone. It is also possible to include trends in the distribution with this modelling method. The vertical proportion graphs created during the data analysis step were the trends used to distribute the lithology in a vertical direction. The thickness maps created during the data analysis step were the trends used to distribute the lithologies in a horizontal direction. The fractions of these trends are used as the probability of the occurrence of a lithofacies in a gridcell. The total probability is calculated as the global fraction times the probability in x,y direction times the probability in the z direction.

The result of the modelling is visible in Figure 51. The main lithology in this facies is obviously silty clay. The high sandy silt content in the northwest corner of the model is the result of the vertical trends and thickness maps. This corresponds with the conceptual model of a retrogradational lagoon. In such a model the sandy silt could represent a coarsening trend towards the barrier, resulting from overwash deposits.

Figure 51
Facies model of Zone 3, the lagoonal silty clay facies. Note that the depth is exaggerated 1,000 times.

5.4.3 Brown sands
To model the brown sands zone the same modelling method as for the silty clays, truncated Gaussian simulation with different parameters, was used. This modelling algorithm was used because all the trends and variograms created during the data analysis can be used with this algorithm. The lithofacies in this zone are, shell debris, coarse sands, medium coarse sands, medium fine sands, fine sands, coarse silty sands, fine silty sands, sandy silts, silt, clayey silt, clayey sand, sandy clay and silty clay. The fractions of these lithofacies, as found in the upscaled logs were honoured. The variograms for this zone were used to define the spatial variations. The vertical
proportion curves were used to distribute the lithofacies in a vertical direction. For the horizontal plane the lithofacies thickness maps were used.

The result of the modelling is visible in Figure 52. The distribution of the lithologies is again mainly dependant on the thickness maps and the vertical distributions. Most of the fine sands are located in a north-south running band parallel to the coast in the middle of the model. Landward on the eastside of the model more silty clay and silt is modelled. Basinward towards the west, also an increase in silt content is modelled. The depositional environment in which this zone was deposited is a barrier-like environment. The process based model indicates that sands deposited in a barrier system can be reworked during multiple phases of sea-level change. The paleotopography of a barrier complex is thus not expected. However the model and the boreholes clearly indicate an increased fines content coastward and basinward. The sediments coastward are therefore probably deposited in a backbarrier/lagoonal environment and the fines basinward as barrier front/shallow marine deposits.

![Figure 52](image_url)

*Figure 52*
*Facies model of Zone 2, the brown sands deposited in a barrier-like environment. Note that the depth is exaggerated 1.000 times.*

5.4.4 Grey sands

To model the grey sands zone, the same modelling method as for the silty clays and brown sands, truncated Gaussian simulation with different parameters was used. The present lithofacies in this zone are, shell debris, medium fine sands, fine sands, coarse silty sands, fine silty sands, sandy silts, silt, clayey sand, sandy clay and silty clay. The fractions of these lithofacies, as found in the upscaled logs were honoured. This method was used because it provides the option to use all the trends and variograms. The variograms generated during the data analysis for this zone were used to define the spatial variations. The vertical proportion curves, created during the data analysis step were used to distribute the lithofacies in a vertical direction. For the horizontal plane the lithofacies thickness maps were used.

The result of the modelling is visible in Figure 53. The distribution of the facies in the model is mainly the result of the thickness maps and vertical distribution. The influence of the variograms is also large in this zone. Very high ranges parallel to the
coast in the order of 20km to 40km were measured and used in the model (Table 8). The ranges perpendicular to the coast measured and used are around 5km, which is a factor 4 to 8 lower. This high contrast in ranges causes the sediments to be spread out parallel to the coast in the facies model. The sediments are deposited in a low sediment influx shallow marine environment with frequent sea-level changes. From a conceptual point of view the distribution parallel to the coast is very likely, in the sense that frequent small scale sea-level changes create large facies shifts over small distances. Alongshore currents can also distribute sediments parallel to the shore. The model shows an increased fines content in the shallow areas coastward. This might be caused by the dissipation of waves and the resulting low energetic environment in these shallow areas.

![Figure 53](image)

*Figure 53*  
Facies model of Zone 1, the grey sands deposited in the present shallow marine environment. Note that the depth is exaggerated 1.000 times.

### 5.5 - Sensitivity analyses

#### 5.5.1 Uncertainty

The interpolated surfaces, thickness maps and facies models are all interpolated between the wells and have a degree of uncertainty. The only points in the model which are certain are the borehole locations. At further distances from the boreholes a larger uncertainty exists. The correlations itself provide a large uncertainty. The model assumes a layer-cake model. Therefore the assumption was made that the lithofacies can be correlated between the wells. Variations within the lithofacies were modelled with the facies modelling and not correlated between the boreholes. To quantify the uncertainty caused by the interpolation of well tops, a map for each point in the model with the distances to the closest borehole locations was created (Figure 54). This map gives the level of uncertainty of the model. Close to the boreholes, the purple colour indicates a low uncertainty. Far away from the boreholes no data are available and the uncertainty of the model is high, these areas are coloured red.
The uncertainty of the model is not only dependent on the distance to a borehole but also on the anisotropy of the subsurface. The sediment bodies are generally deposited parallel to the coast. The uncertainty between two boreholes located in a line parallel to the coast is therefore lower than the uncertainty between boreholes located in a line perpendicular to the coast. This anisotropy was described with variograms in the data analysis paragraph. The average observed range parallel to the coast is 11,000m, perpendicular to the coast the average observed range is 5000m. This information was used to create a kriging variance map of the study area (Figure 55). The kriging variance is a good measure of uncertainty and is dependant of the spatial arrangement of the borehole locations. It indicates the expected variance between a point in the study area in relation to the nearest boreholes. The model has a high uncertainty in areas with a high kriging variance and a low uncertainty in areas with a low kriging variance.
5.5.2 Range sensitivity

The range and nugget were determined with a sample variogram for each lithology in each zone. The shape of the sample variogram relies on the direction of the search cone, lag distance and search radius. For some lithologies the trends in the sample variograms are not clear. The values for the range and nugget are thus dependant on an interpretation and not always well defined. To test what the influence of the variograms on the model, a sensitivity analysis was performed. This analysis was done by creating multiple realisations of the facies model, with different values for the range or nugget.

A quantitative property is needed to compare these models. A good parameter to do this is the net over gross ratio. It indicates the ratio between sand and the total volume of the model. The property net over gross was calculated with the property calculator option in Petrel. The volume calculation process was used to calculate the net and gross volumes in the 4 zones.

A workflow was created in which the facies modelling, net to gross calculation and volume calculation processes could be repeated with a variable range. During each loop a random range between 300m and 30,000m was used. By repeating this loop 15 times, 15 facies models with different ranges were created. The results of the volume calculations show the net and gross volumes of each model.

Figure 55

Kriging variance map of the study area. A high variance indicates a high uncertainty. A low variance indicates a low uncertainty.
The Petrel workflow which was used to repeat the calculations for the sensitivity analysis. In this case the nugget is the variable parameter indicated by $z_{on}$. In the numeric expression step a random value between 0 and 1 is chosen for the nugget. The facies modelling steps, net to gross calculation and volume calculations are then calculated. These steps are repeated in a loop, and the results are stored in a new case.

For each zone a graph was created with the net volume versus the range (Figure 57). The graphs indicate the effect of the range in the different zones. In Zone 1 the net volume decreases as the range increases. In Zone 2 the net volume increases as the range increases with a Gaussian trend. In Zone 3 the influence of the range on the net volume is the lowest, the net volume decreases slowly when the range increases. In the bar graph, the influence of the range on the total net volume in the model is visible. The influence of the range on the total net volume, in the facies model, is almost insignificant.

The percentages of the lithology are nearly completely determined by the thickness maps, vertical distribution and total percentages. The range is more important for the homogeneity of the facies model. In Figure 58 two realization of the facies model for Zone 1 are shown with a range of 100m and a range of 50,000m. The low-range model has a lower homogeneity and a more “speckled” distribution than the high-range model, which has a “patchier” distribution.
Figure 57
The net volume versus the range in each zone for 15 different realisations of the facies model.

Figure 58
Facies models of Zone 1 with two different ranges. For the realization on the left model a range of 100m was used, for the right realisation a range of 50,000m was used. Both models have a nugget of 0.01.
5.5.3 Nugget sensitivity

The same sensitivity analysis was performed for the nugget. A random nugget between 0 and 1 was used to test the influence on the net over gross ratio.

For each zone a graph was created with the net volume versus the nugget in Figure 59. The graphs indicate that the effect of the nugget is different in the different zones. In Zone 1 the net volume decreases as the nugget increases. In Zone 2 the net volume increases as the nugget increases. In Zone 3 the net volume increases as the nugget increases. In the bar graph, the influence of the nugget on the total net volume in the model is visible. The influence of the nugget on the total net volume, in the facies model, is clearly insignificant.

The percentages of the lithology are almost totally determined by the thickness maps, vertical distribution trends and total percentages. However, the nugget is important for the homogeneity of the facies model. In Figure 60 two realization of the facies model for Zone 1 are shown with a nugget of 0.01 and 0.75. The low nugget model has a very large homogeneity. The edges between horizontal lithologies are very sharp. The high nugget model has a very low homogeneity.

![Graphs of net volume versus nugget for Zones 1, 2, 3, and 4](Figure 59)

*Figure 59*

*The net volume versus the nugget in each zone for 15 different realizations of the facies model.*
Figure 60

Facies models of Zone 1 with two different nuggets. For the realization of the left model a nugget of 0.01 was used, for the right realization a nugget of 0.75 was used. For both models a range in x and y direction of 50000m was used.
6 Results
This chapter gives a summary of the results. The complete workflow and a detailed description of the results can be found in the chapters 4 “Sedimentological data interpretation” and 5 “Modelling”.

6.1 - Data interpretation results
Four lithofacies were recognized in the cores named the grey sands (0-1.5m bsf), the brown sands (1.5-4m bsf), the silty clays (4-9m bsf) and the overconsolidated sands (Table 9). Using sedimentological, biostratigraphical and seismic data, interpretations for the depositional environments of these lithofacies were made. The overconsolidated sands were deposited as longitudinal aeolian dunes and reworked by coastal processes. The silty clays were deposited in a lagoonal environment and the brown sands in a barrier-like environment. The grey sands were deposited in the most recent depositional environment, characterized by a low sediment influx and frequent small scale sea-level changes. Radiocarbon dating of molluscs was performed and resulted in dates between 27.000y BP and at least 48.000y BP (Table 9).

Sequence stratigraphical analysis shows that the overconsolidated sands have been deposited during a lowstand and reworked during a subsequent transgression. The silty clays and brown sands have been deposited during at least one transgression or more. The contact between both units can be interpreted as a ravinement surface or a subaerial unconformity. Between the deposition of the brown sands and grey sands erosion occurred. From this the conclusion can be drawn that the sediments encountered in the cores are deposited during at least three but more likely 4 large scale sea-level sequences.

When placing the results from the sequence stratigraphy and dating in a historical context, multiple sea-level graphs can be constructed, based on the dated samples and known sea-level history (Figure 61) (Dumont 1998, Kroonenberg 2008, Mamedov 1997 and Rychagov 1997). The Inter Khvalyn regression can be placed during the deposition of the overconsolidated sands, between the deposition of the silty clays and brown sands or after the deposition of the brown sands. The erosional surface between the brown sands and recent grey sands is probably created during the Mangyshlak regression or Derbent regression.
### Table 9

*Table with the 4 facies and their descriptions, depositional environments, C14 dates and depositional trend.*

<table>
<thead>
<tr>
<th>Facies name</th>
<th>Depth m bsf</th>
<th>Description</th>
<th>Depositional environment</th>
<th>C14 dates</th>
<th>Depositional trend</th>
</tr>
</thead>
<tbody>
<tr>
<td>grey sands</td>
<td>0-1.5m</td>
<td>fine to coarse sands with shell debris and gypsum bands</td>
<td>present environment</td>
<td>not dated</td>
<td>multiple small scale cycles</td>
</tr>
<tr>
<td>brown sands</td>
<td>1.5-4m</td>
<td>fine to medium layered sands with small shell fragments</td>
<td>barrier</td>
<td>27.000BP</td>
<td>transgression</td>
</tr>
<tr>
<td>silty clays</td>
<td>4-9m</td>
<td>olive grey silty clay with organic rests and gypsum bands</td>
<td>lagoon</td>
<td>&lt;48.000BP-42.000BP</td>
<td>transgression</td>
</tr>
<tr>
<td>Overconsolidated sands</td>
<td>9-?m</td>
<td>fine to medium sands with small shell fragments</td>
<td>reworked aeolian</td>
<td>&lt;48.000BP</td>
<td>lowstand</td>
</tr>
</tbody>
</table>

**Figure 61**

*Image showing the results in a sea-level graph; The C14 dated samples (black dots); The C14 dated samples older than 48.000y (red dots); The erosional surfaces found in the cores (red spotted lines); The results of the biostratigraphical analysis (coloured bars); Three possible Caspian sea-level curves for the period between 60.000y BP and 20.000y BP (the three dotted lines); The three different phases of deposition (indicated in grey); The Composed Caspian sea-level curve from literature (grey line) (after Dumont 1998, Kroonenberg 2008, Mamedov 1997 and Rychagov 1997).*
6.1 - Modelling results

A subsurface model was build to visualize the 3D-spatial sediment distribution and test the conceptual model. Correlations between the new CDS cores, geotechnical boreholes and cone penetration tests were made (Figure 62). The 4 lithofacies can be correlated over large distances, in the range of 10-30km. Between the correlated sequence boundaries, surfaces were created. The correlations show that the deposits have the highest continuity parallel to the coast.

To determine the variation in lithology distribution within the main lithofacies, variograms were made for every lithology in each facies. These were later used for facies modelling.

Vertical proportion curves for each lithology in each facies were made using all the data from the geotechnical and CDS-cores. They indicate that the lithologies within the overconsolidated sands are uniform distributed vertically. Within the silty clay lithofacies, the silt content increases towards the top of the zone. Within the brown sand facies, a slight increase of silty clay towards the bottom was noticed. In the grey sands an increase in shell debris towards the top is noticed.

The lithology distribution in the horizontal plane was determined with thickness proportion maps of the geotechnical and CDS-cores. In the silty clay lithofacies, they indicate an increase in silt basinward. In the brown sand facies a high sand content parallel to the shore is observed with more silty clays coastward and silt in a basinward direction.

To populate the model with sedimentological bodies, facies modelling was performed. The overconsolidated sands are modelled with object based modelling, with properties derived from the seismic data. The other zones are modelled with truncated Gaussian modelling, based on lithology thickness maps, vertical lithology proportion trends and data driven variograms of the lithology.

The object-based modelled overconsolidated sands, show similar shapes, sizes and orientations as the longitudinal-dune features observed in the seismics (Figure 62, Figure 63). The facies model of the silty clays shows and increase of coarser sediments towards the top of the unit on the seaward side of the study area (Figure 62). These coarser deposits can be related to overwash deposits of a barrier complex seaward of the study area. Lithology thickness maps within the brown sand facies show an accumulation of sand parallel to the coast, with a higher clay fraction coastward and a higher silt fraction seaward (Figure 62, Figure 63). This sand accumulation is interpreted as a relict barrier complex and enforces the concept of deposition in a barrier-like environment. The model of the grey sands shows a high heterogeneity, especially perpendicular the shore, and has a high shell fragments content in the top layers.

A kriging variance map was created to indicate the uncertainty of the model caused by the location of the boreholes. Close to the boreholes the uncertainty is low, far away from the boreholes the uncertainty is high. Volume calculations were performed to test the influence of the range and nugget on the model. These two parameters have a low influence on the volumes, but a large influence on the homogeneity of the model. A high range or a low nugget, results in a high homogeneity, a low range or a high nugget in a low homogeneity. The main
parameters influencing the lithology distribution are the thickness proportion maps and vertical proportion curves.

**Figure 62**
E-W cross-section of the facies model, perpendicular to the shore. The horizontal black lines indicate the surfaces which separate the 4 lithofacies. The depth is exaggerated 1.000 times.

**Figure 63**
SW-NE cross-section of the facies model, parallel to the shore. The horizontal black lines indicate the surfaces which separate the 4 lithofacies. The vertical black lines indicate the borehole locations. The depth is exaggerated 1.000 times.
7 Conclusions and discussion

7.1 - Conclusions

The objective of this research was to improve the knowledge about the depositional environments and subsurface architecture of the Northeastern Caspian Sea. Based on the results, the following conclusions can be drawn about the depositional environments that played a role during the sedimentation of the deposits found in the cores and their influence on the subsurface architecture.

This study shows that the sediments found in the subsurface of the study area have been deposited in 4 different depositional environments. The depositional environment depends on multiple factors in which the relative position to the coast is one of the most important factors. Another important factor is the supply of sediments, and the speed of sea-level change. It seems that only large scale sea-level cycles combined with changes in sediment supply have an effect on the major depositional environment, while more frequent sea-level cycles only cause a brake or temporary change in deposition.

The oldest dated sediments in the cores are deposited before 48.000y BP. Presumably the depositional environment in the study area at that time was aeolian. Longitudinal dune features formed, on the former seabed. During a transgression the dunes were flooded and partially eroded and reworked. But, features of their original shape were preserved and are still visible on seismic images.

On top of these dune features a lagoon formed, which was probably protected from the sea by barriers. The lagoon existed for at least 6.000 years between 48.000y BP and 42.000y BP. Proof for frequent small scale base-level changes was found in these silty clays, but only lagoonal sediments are preserved. Overall the lagoonal deposits form an aggradational sequence. The vertical and lateral homogeneity of this layer is therefore very high. The upper-seaward side of the lagoonal deposits contain more silty material, which is probably related to overwash deposits of a nearby barrier. Probably as a result of sea-level change or a decrease in sediment supply, retrogradation of the coastline finally resulted in backstepping of the barriers.

The top part of the lagoonal sediments was eroded by subaerial exposure or by coastal processes. This means that either a regression or a subsequent transgression could have taken place. Eventually transgressional barrier sediments were deposited on top of the lagoonal deposits. One radiocarbon date around 27.700y BP is known from these deposits. A sand package was deposited in the middle of the study area parallel to the coast. Basinward finer silty sediments were deposited in front of the barrier. The homogeneity of these deposits is high parallel to the coast and lower perpendicular to the coast.

Process response modelling can simulate the effects of different sediment supplies and sea-level curves on a low-gradient barrier coast. According to the model a large sediment influx is needed to create aggradation of a barrier-lagoon complex. It shows that the most pronounced effect of a low gradient is that barrier overstep takes place over a large distance. Erosion in front of a barrier system can also leave extensive erosional surfaces. From the modelling, the conclusion can be drawn that the stratigraphy of the lagoonal and barrier deposits, as found in the subsurface of
the Northeastern Caspian Sea, has most likely been the result of multiple centennial to millennial or smaller sea-level cycles and probably not the result of a single transgression.

During the Mangyshlak regression or Derbent regression a part of the barrier deposits eroded away. The grey sands lithofacies is deposited on top of this erosional surface. From this period onwards, the sediment influx was very low. Most of the sediments in this lithofacies are derived from reworking of older sediments and biogenic (shells) influx. During frequent small-scale regressions gypsum layers are deposited. The frequent sea-level changes and low sediment influx leave a sediment package which has a low vertical homogeneity. The horizontal homogeneity is the highest parallel to the coast, and lower perpendicular to the coast.

The subsurface model visualizes the 3D spatial distribution of the sediments based on core and CPT data and verifies the conceptual geological model. Facies modelling was performed with object based modelling for the overconsolidated sand facies, based on shapes observed on seismics, and sedimentary properties observed in the cores. The other 3 facies were modelled with truncated Gaussian modelling techniques, based on trend maps, vertical trends and variograms as observed in the core and CPT data. The trend maps and vertical distribution curves have the most pronounced influence on the distribution of the lithologies and the variograms on the heterogeneity.

The model shows that the conceptual model developed with the sedimentological and sequence stratigraphical analysis lines up with the spatial distribution of the sediments observed in the cores. It also represents the hypothetical subsurface architecture of the study area according the data and the conceptual model.
7.2 - Discussion

The presence of an aeolian dune facies between coastal deposits is surprising, although large vertical facies shifts in the stratigraphical record are typical for an environment with a low gradient and frequent sea-level changes (Overeem 2002, Hinds 2004). The interpretation of a reworked-aeolian depositional environment of the overconsolidated sands is based on the features seen on seismic images and the occurrence of onshore longitudinal dunes with similar dimensions and orientations in the North Caspian region. The initial fluvial interpretation of the features seems very unlikely because the extent, density and orientation of the features do not fit with channels of a fluvial system. The sedimentological evidence from the cores could also indicate coastal deposition of the sands. This can be explained by reworking of the sands in a coastal environment. The origin of the onshore-longitudinal dunes was also debated by other scientist. Baer (1855) and Leont’ev (1965) both assumed an aeolian origin of the onshore dunes. Others assume a sheet flood origin and reworking by aeolian processes (Badyukova 2005). More research of these reworked dunes is needed, especially when taking into account that aeolian deposits are known to be excellent hydrocarbon reservoirs. They could also occur for example in the South Caspian basin’s Productive Series, which were also deposited in a low gradient environment with frequent sea-level changes (Reynolds et al. 1998).

Radiocarbon dating shows that the silty clays have been deposited in a time span of at least 6,000 years. Comparing this period with the last 6,000 years, one can expect that multiple centennial to millennial and less frequent sea-level cycles occurred during this period. Despite these assumed sea-level changes, the silty clays are very homogene. The only evidence for a change in depositional environment are colour differences and changes in organic content, but the interpreted depositional environment remains lagoonal. The effect of these frequent sea-level changes on the depositional environment seems to have been very small during this period. A reason could be that during the frequent sea-level changes barriers and lagoons drowned passively or were exposed without evident erosion taking place.

Similar silty clays are found through the whole North Caspian region. They are often referred to as chocolate clays, because of their chocolate brown colour (Kroonenberg 1997, 2003, Badyukova 2007). There is much debate about the exact age of these clays and whether they are connected to the Early or Late Khvalyn highstands. The chocolate clays are dated between 11,000y BP and 30,000y BP and are interpreted as deep marine deposits (Svitoch 1976, 2007, Badyukova 2007). However, Badyukova (2007) states that the chocolate clays are deposited in a lagoonal environment. She concludes that they have been deposited in lagoons protected by barriers which formed during multiple transgressions at different levels. Therefore the chocolate clays cannot be correlated over the whole Caspian basin, and for this reason are dated at different ages. The depositional environment explained by Badyukova (2007) and the features of the chocolate clays resemble the silty clays found in our cores remarkably well, and supports the barrier-lagoon hypothesis. The C14 dates in the silty clays in our cores indicate deposition between at least 48,000y BP and 42,000y BP. A date from the barrier sands on top of the silty clays indicates deposition around 27,000y BP, which corresponds with the timing of the deposition of the chocolate clays. The transgressive barrier-lagoon environment in the Caspian Sea created by multiple sea-level cycles therefore could well have lasted for a long period, from at least 48,000y BP until 11,000y BP. During this period barrier and lagoonal sediments have been deposited at multiple levels.
Theoretically, the low energetic, protected environment of a lagoon can also be formed without the presence of barriers in the Northeastern Caspian Sea. The shallow gradient of the area can cause dissipation of waves, what results in a low energetic environment close to the shore without the need of a protecting barrier. Kaplin and Selivanov (1995) and Kroonenberg (2000) say that shores with a gentle slope (<0.03°) show passive drowning during a sea-level rise. According to them only shores with a slope between 0.03° and 0.3° show the formation of barriers. The present day coast of the Northeastern Caspian Sea has an average offshore slope of 0.007°, which would be too gentle for the formation of barriers and lagoons. At the present moment, indeed no active barrier formation takes place along the Northeastern Caspian shores. However, this study and Baduykova (2007) show that in the past, active barrier formation occurred in the study area. A reason for the formation of barriers in the Northeastern Caspian Sea could be that the area had a steeper slope in the past. Another reason could be that a different sediment budget resulted in the formation of barriers on a slope with a lower gradient than 0.03°. Therefore it would be useful to investigate the formation of barriers on a gentle slope with process-response modelling and include dissipation of waves in the model.

The contact between the lagoonal facies and barrier sand facies is erosional and could represent either a subaerial unconformity or a ravinement surface. Dating of samples above and below the contact could give an indication of the nature of this contact. This would indicate whether the silty clays and brown sands are deposited during the same transgressional sequence or if the change from silty clay to sand is caused by a regression. Because erosional surfaces extend laterally over a whole basin, while ravinement surfaces occur only locally, the lateral extensiveness can also give information about its nature. When the erosional surface is only present locally, it is probably a ravinement surface. If the erosional surface can be traced over the whole basin, it represents a subaerial unconformity.

The barrier sands are very continuous, both vertically as horizontally. This means that they have been created by large barriers, or by multiple, stacked barriers. Large barrier lagoon-complexes are formed during transgressions (Kroonenberg 2000). Two models exist over the retrogradation of a barrier. One, in which a barrier moves coastward continuously, without losing its shape and volume and one in which a barrier grows larger until it has become too steep and steps back over a distance and starts over again closer to the shore (Swift 1975, Sanders and Kumar 1975, Rampina and Sanders 1970, Swift et al. 1991). The process-response model shows that during the continuous retreat of a barrier not much of a barrier-lagoon complex is preserved. To preserve large parts of a barrier-lagoon complex, backstepping of the barrier is needed. Because the barrier sands in the study area are extensive, most likely this situation has occurred repeatedly during the deposition of the barrier sands.

The contact between the barrier sands and overlaying grey sands is erosional. The deposition of the grey sands could have started at the end of the Mangyshlak regression or after the more recent Derbent regression. In the first case, the sedimentation rate would be extremely low, in the order of 1cm/100 year, in the latter case the sedimentation rate of the grey sands would be around 1cm/5 year, which is still low. The low sediment influx could declare why there is no active barrier lagoon complex in the Northeastern Caspian at present day, hence a large sediment influx is needed to create a barrier-lagoon complex. During the last lowstand in 1975, satellite images show that the present day barriers in the Northeastern Caspian Sea
where exposed, and drowned passively during the last transgression. Probably the present day sediment influx is too low, and the sea-level cycle too short to create an active barrier-lagoon complex. The barrier islands in front of the present day coast are therefore relict barriers, from a period when the sediment influx was higher.

The difference between the present depositional environment and the environment during the deposition of the lagoonal and barrier facies is not sea level but sediment supply. Both facies are deposited during periods when the relative sea level did not change that much from the present day situation. However, the present deposits are deposited in a totally different environment than the lagoonal and barrier sand facies. The largest difference between both environments is the sediment influx. It seems that for a major change in depositional environment the type and amount of sediment influx has a large effect.

Potential sources for sediments during the deposition of the barrier and lagoonal facies are the Ural or Emba rivers. A backstepping-barrier complex could have transported the sediments from the Ural delta, which is located in the Ural furrow during lowstands, towards the study area. It might also be possible that coarse sediments were provided by erosion of older deposits, such as the overconsolidated sands, away from the study area. These sediments could have been transported by alongshore currents.

Many different Caspian sea-level graphs have been made in the past (Mamedov 1997, Svitoch 2006, Mangerud et al. 2001, Kroonenberg et al. 2008, Rychagov 1977, 1997). The exact timing and sea levels of the highstands and lowstands are much debated and there is a lot of inconsistency about their names. This study supports the possibility of a lowstand before 48.000y BP. This lowstand was followed by a major transgressive phase between 48.000y BP and 27.000y BP or two transgressive phases separated by a regression after 42.000y BP. Depending on the interpretation of the contact between the silty clays and overconsolidated sands.

The current findings enhance our understanding of the depositional processes in a low gradient, low sediment influx coastal environment with frequent sea-level changes. It adds knowledge about the subsurface architecture of the unique depositional environments of the Northeastern Caspian Sea and its sea-level fluctuations.
8 Recommendations

More dated samples from the Northeastern Caspian Sea could provide additional points for a Caspian sea-level curve. An accurate sea-level curve can give better information of the influence of sea-level change on sedimentary processes, and can be used for the whole Caspian region. Suited samples for dating of the cores are available but arrived too late for this project. Because many of the samples in the silty clays proved to be older than 48,000 y BP which is beyond the limit for C14 dating, uranium-thorium dating techniques are advised for the silty clays or older deposits. Thermoluminescence dating, which dates the time elapsed since a mineral has been exposed to sunlight, could be used to estimate the time of deposition and flooding of the aeolian sands, or other subaerial exposed sediments. To determine if the silty clays and brown sands have been deposited in the same sequence, dating just above and just below the contact between both could be done. This is important to determine whether a regression took place between the deposition of both units or not.

Texture analysis of samples can provide information about grain-sizes and sorting. Suited samples for grain-size analysis of the cores are available but arrived too late for this project. Grain-size trends can be used for a more detailed analysis of the depositional environments and sequence stratigraphical analysis. In this way aggradational, retrogradational and progradational trends could be recognized within the lithofacies. This gives information about the balance between sediment supply and base-level changes. The sediment sources for the deposits encountered in the subsurface are still unknown. To make a conclusion, the texture samples taken for this research could also be used to determine the provenance of the sediments, by genetic grain-size decomposition. Analysis of the mineral content can also provide information about sediment provenance.

The present day barriers and lagoon in front of the northeastern Caspian coast are a good analogue for the barrier and lagoonal deposits encountered in the subsurface. They might have been deposited during the same transgressive phase as the deposits encountered in the subsurface. Therefore, coring these barrier islands and lagoons would be very useful to test the depositional model and increase the knowledge about barrier deposits in shallow gradient environments.

At this moment it is not possible to include dissipation of waves in the used process-response modelling software BARSIM. Including dissipation of waves makes the model better suited for coastal process research in low gradient environments.

More research on the longitudinal dune features in the overconsolidated sands is needed to make a conclusion about their origin. This can be done by studying more seismic surveys in the region to determine the shape and extend of the features. Cores in the seismic survey areas can be used to verify the seismic reflector and used to do a better sedimentological study of these features. Onshore sedimentological studies of the longitudinal dunes can be used to compare with the features in the subsurface. Aeolian dunes are known to be good reservoir rocks. If these reworked dune sediments, occur in similar settings, for example the Miocene South Caspian basin, they can be a potential hydrocarbon reservoir rock.
Geotechnical cores could be used more often for sedimentological research, especially when adding descriptions of contacts between units and coarsing- or fining-upwards sequences. Geotechnologists can derive a lot of information from sedimentological information, for example the distribution of sediments. Less cores are needed for geotechnical purposes if more is known about the distribution and homogeneity of deposits.

The variations within the lithofacies were not correlated. Modelling the smaller scale variations within the lithofacies, will give a better facies model and would decrease the uncertainty of the model. It can also give more detailed information about the smaller scale sea-level changes. The used dataset would be very suited for a higher resolution correlation because of the close spacing within the clusters of geotechnical boreholes and CPTs.

The dating of chocolate clays in the Caspian Sea has lead to a lot of discussion over the age of the Khvalyn highstands. Because the chocolate clays are deposited in different lagoons, during multiple sea-level oscillations, they are not correlatable. Individually the dates obtained from the chocolate clays can be used. It is therefore recommended that the dates of the chocolate clays versus their depth bsf plus depositional depth are plotted in a sea-level graph.
9 References


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Appendix

Borehole correlations
This appendix contains all the borehole correlations and cluster maps.

Figure 64
*CDS borehole locations and cross-sections CDSA-CDSF and CDSB-CDSC.*
Figure 65
Cross-section of sedimentological logs CDSA-CDSF.
Figure 66
Cross-section of sedimentological logs CDSB-CDSC.
Figure 67
Cluster Aktote borehole locations and cross-section BHC4-BHC3.
Figure 68
Cross-section of geotechnical logs BHC4-BHC3.
Figure 69
Cluster Kairan borehole locations and cross-section CDSA-KAI1BH1.
Figure 70
*Cross-section of geotechnical logs CDSA-KAI1BH1.*
Figure 71
Cluster C borehole locations and cross-section BHKECNW-BHKECSE.
Figure 72
Cross-section of geotechnical logs BHKECNW-BHKECSE.
Figure 73
Cluster D borehole locations and cross-sections D25-D56, D46-D49, D53-D51, D54-D50 and D30-D37.
Figure 74
Cross-section of geotechnical logs and CPT friction ratios D25-D56
Figure 75
Cross-section of geotechnical logs and CPT friction ratios D46-D49
Figure 76
Cross-section of geotechnical logs and CPT friction ratios D53-D51
Figure 77
Cross-section of geotechnical logs and CPT friction ratios D54-D50
Figure 78
Cross-section of geotechnical logs and CPT friction ratios D60-D37
Figure 79
Cluster G borehole locations and cross-section BHKEGNW-BHKEGSE.
Figure 80
Cross-section of geotechnical logs BHKEGNW-BHKEGSE.
Figure 81
Cluster F borehole locations and cross-sections A13-A11 and A06-A09.
Figure 82: Cross-section of geotechnical logs and CPT friction ratios A-13-A11
Figure 83
Cross-section of geotechnical logs and CPT friction ratios A06-A09