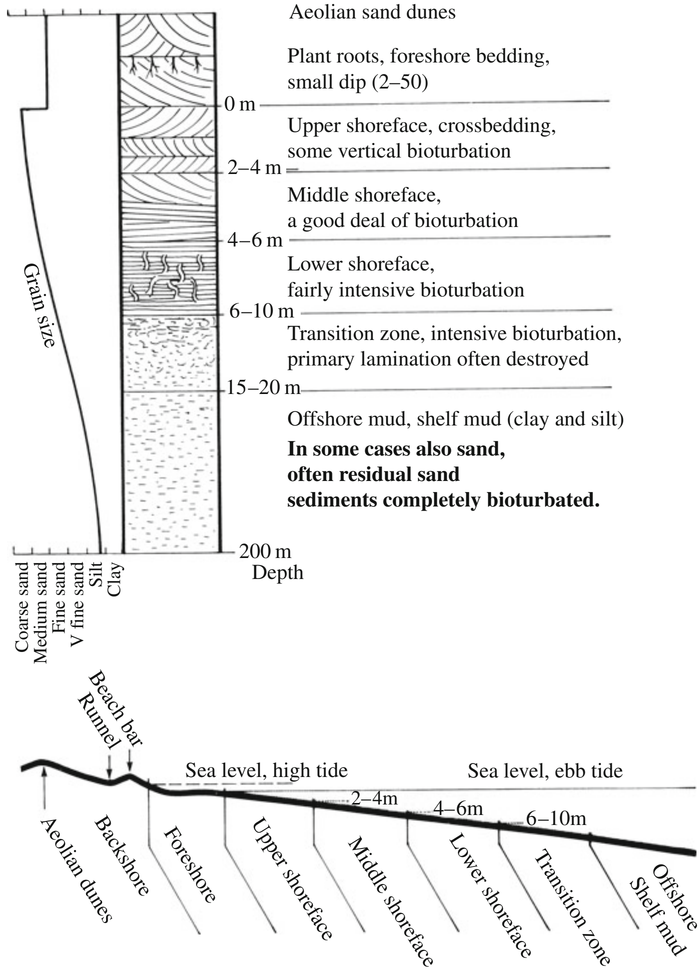

Fig. 2.39 Diagrammatic section through a sand deposit with some typical sedimentary structures. Just under normal fair-weather wave base, 5–20 m, "hummocky" stratification occurs

Bioturbation occurs in the lower part of this sequence. Below the wave base it usually takes the form of horizontal feeding traces, and in the lower and middle shorefaces, where there is relatively high wave energy, as vertical traces (*Skolithos* facies). As erosion and reworking intensify, the preservation potential of bioturbation is reduced and it becomes less frequent. The formation of sand bars and erosion surfaces on the upper shoreface results in cross-bedding, usually trough cross-bedding, representing flow in the upper part of the lower flow regime.

In the breaker zone there is an upper flow regime, producing a planar facies which in vertical section will appear as very low-angle cross-bedding. On the beach we often have a beach bar which is flooded only during storms, and a depression behind it called a runnel, which helps to drain the backshore area. Because the exposed beach is a rich source of sand, the wind will tend to blow sand from the beach and redeposit it as aeolian dunes, usually where it is trapped by vegetation. Aeolian sediments therefore often cap ancient beach profiles, but the aeolian dunes may also be eroded and not be preserved in the geological record.

2.38 Barrier Islands

Barrier islands are beach deposits which are separated from the mainland by a lagoon. They form long, thin sand ridges which are often only a few hundred metres to a couple of kilometres broad, and which rise up to 5–10 m above sea level.

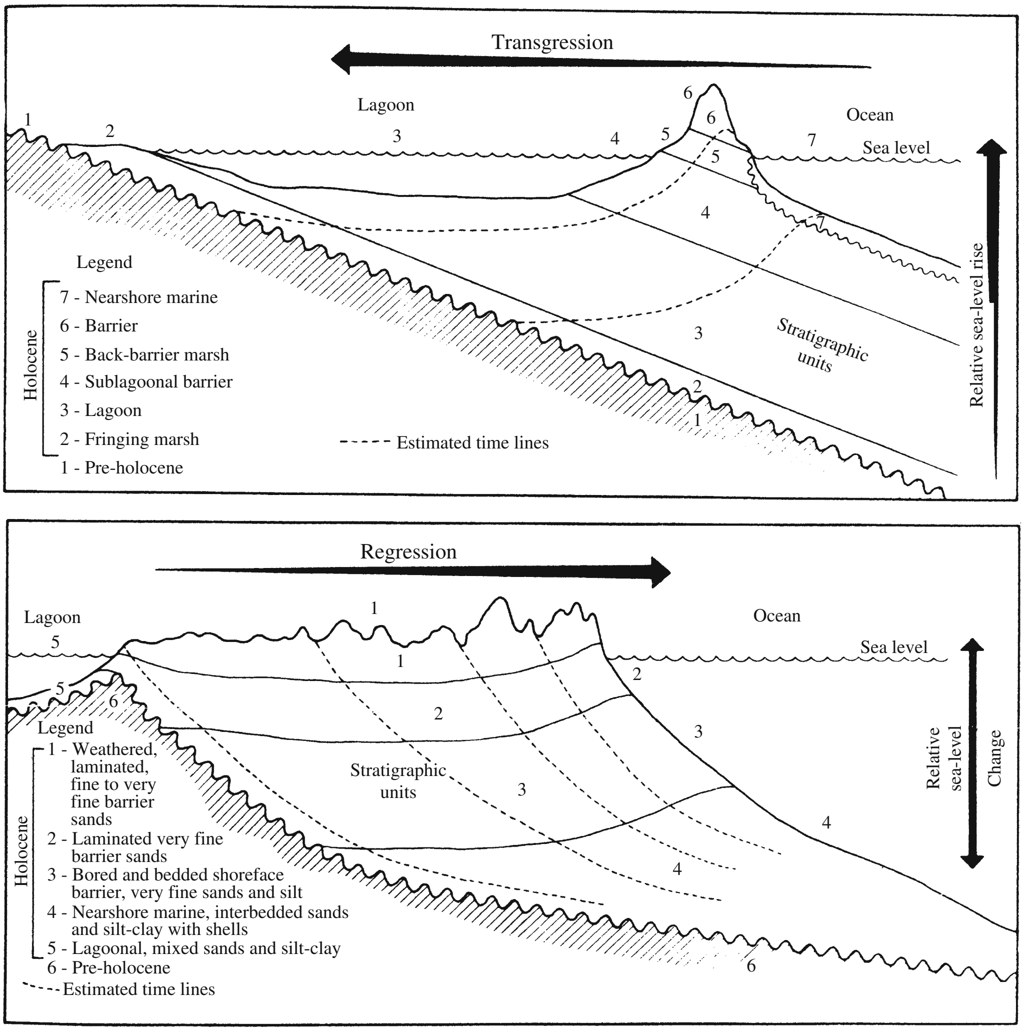

Fig. 2.40 Transgressive and regressive barrier systems. (from Kraf and John 1979)

A vertical section through the part of a barrier island facing the sea resembles an ordinary beach deposit (Fig. 2.40). We find a coarsening-upwards sequence from marine clay to beach sand, often with vegetation on top. On the lee side, facing the lagoon, there is little wave power and clay, mud and often oyster reefs are deposited in the lagoon. Barrier islands are very well developed along long stretches of the coast of North America, particularly off Texas, Georgia and North Carolina. There are several indications that barrier islands are due to transgressive conditions such as those of Holocene times. This is certainly the case along the eastern coast of the southern North Sea. When the ocean rises in relation to the land, beach deposits can continue to grow through gradual deposition of sand in the beach zone, but the areas behind them sink below sea level and form a protected lagoon with clay sedimentation. More localised transgressions may be caused by compaction and subsidence of sediments along a coastline where the sediment supply is insufficient to keep pace with the subsidence.

One prerequisite for forming barrier islands is an adequate supply of sand, so that the island can grow and keep pace with the transgression. This sand cannot be transported across the lagoon, and must be added along the length of the islands parallel to the coast (strike-feeding from deltas or eroding coastlines).

During storms or hurricanes the sea level may rise due to wind stress, and waves may break over and through the barrier island. An erosion channel is then formed through the island, and a *washover fan delta* develops at the rear, out into the lagoon. A delta of this type can form in a matter of hours during a hurricane. Barrier islands may extend for tens of kilometres, but they will not form a continuous belt along the coast. Water has to circulate between the lagoons and the ocean through gaps between barrier islands. The gaps are called *tidal inlets*, and the distance between them will be a function of the tidal range. Both seaward and landward of the inlets small sandy deltas may develop in response to ebb and flood currents respectively (Fig. 2.41). Flow at ebb tide will normally be stronger than that at flood tide. This is due to the profile of the lagoons. At high water in the lagoon the volume of water which must flow out to compensate for a specific lowering of sea level is greater than the volume needed to raise the water level in the lagoon correspondingly at low tide when the area of the lagoon is smaller. The tidal inlets are therefore capable of transporting more sediment out during the ebb, and structures indicating this flow direction may predominate (Fig. 2.41). Inlets are characterised by an erosional base, and lateral migration of inlets produces a characteristic fining-upwards sequence. The strong currents in tidal inlets often generate sand waves which tend to migrate in the ebb direction, but they may also be modified by flood currents.

Ebb-tidal deltas consist of a channel dominated by ebb currents with smaller flood-tide channels on the sides. At the ocean end of the channel sediment is deposited in a sand ridge which is similar to a channel mouth bar in an ordinary delta. This sand ridge, which is called *a terminal lobe, is* subject to wave erosion, and smaller *swash bars* may form, which reach above sea level. In areas with strong wave power, ebb-tidal deltas will be less obvious because of erosion and further transport along the barrier ridges. Ebb-tidal deltas will be characterised by greater water depths than flood tidal deltas.

Flood tidal deltas form inside the lagoon and are well protected from wave erosion. Here the water flows into flow channels which branch inwards in a flood-tidal delta, where the sediments are deposited on a tidal flat. Ebb currents move back along the edges of the outer side of this delta and may form small *spillover lobes* when ebb-currents penetrate over the edge of the flood-tidal delta. Flood-tidal deltas are associated with shallower channels than ebb-tidal deltas and are not eroded very much by waves.

Tidal channels fill with sand which forms an upward-fining sequence overlain by tidal flat sediments (Fig. 2.42). In areas with carbonate sediments or cohesive clays, erosion due to lateral migration of tidal channels results in intraformational breccias.

Barrier island deposits thus consist of a long, thin body of sand. The thickness of the sand layer will correspond to the depth of the wave base plus a few

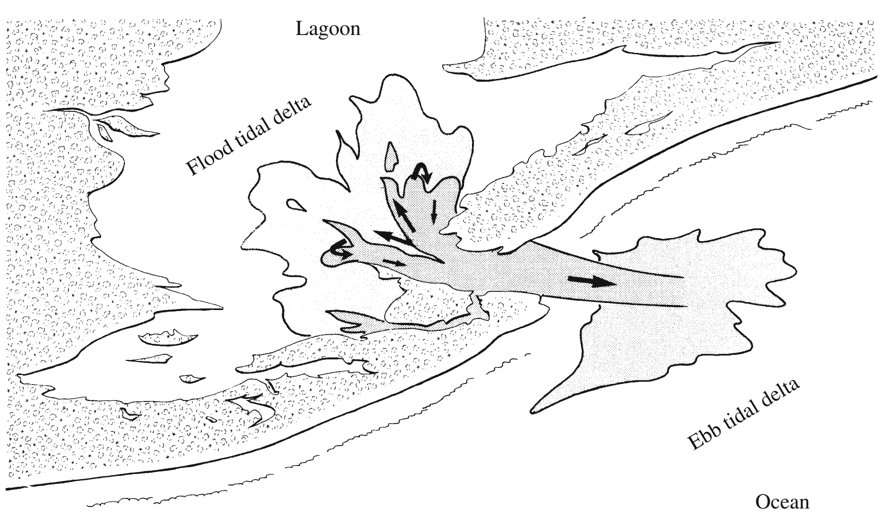

Fig. 2.41 Tidal channels with tidal deltas forming between barrier islands. The barrier islands and the channels will migrate laterally and deposit channel facies sediments by lateral accretion. Note that the ebb-tidal delta outside the barrier is much more exposed to waves than the flood-tidal delta in the lagoon

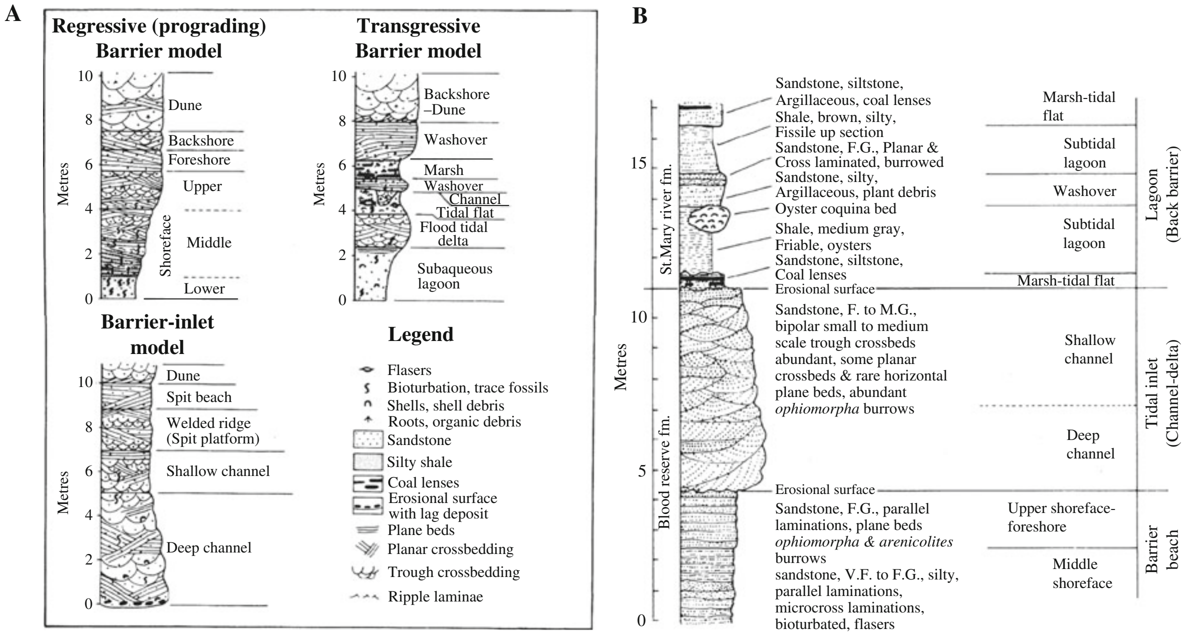

Fig. 2.42 (**a**) Interpretation of a vertical section in a tidal environment (From Walker 1979). (**b**) Section through a barrier beach cut by a tidal inlet channel into the lagoonal facies

metres which correspond to the height it builds up to above sea level.

In areas with a larger tidal range, this lateral migration will be rather pronounced, and fining-upwards sequences will also be common.

If the barrier islands are drowned by a transgression, a carpet of clay and silt will be deposited over these sandstone deposits. This represents the ideal stratigraphic trap for oil and gas. Compaction or tectonic tilting will cause the sandstone deposits to interfinger with mud from the lagoon deposits, which are a good source rock. Oil will be able to collect in the top of the barrier ridge sand or in flood-tidal delta deposits (or *washover fans)* which represent *pinch-outs* in the muddy lagoon sediments.

2.39 Tidal Sedimentation

Tidal range is an important factor in coastal sedimentation. We distinguish between:

1. Microtidal environment (tidal range less than 2 m).
2. Mesotidal environment (tidal range 2–4 m).
3. Macrotidal environment (tidal range greater than 4 m).

The average tidal range in the open sea is only about 50 cm. Along the coasts, however, we often get increased interference by tidal currents. This is particularly true around large islands where tidal waves converge on the lee side and can build up, and also in bays along the coasts. In long, narrow bays we may get a high degree of *resonance*. This occurs if the bay has a length and depth which cause tidal waves which are on the rebound to reinforce the next incoming tidal wave. The highest tidal range which has been measured is 16.3 m in the Bay of Fundy in Canada. In some areas around the British Isles the tidal range may be up to about 12 m, and there are also large ranges in the German Bight and adjacent parts of the North Sea. Tidal waves also move anticlockwise in N hemisphere around centres with zero tidal range (amphidromic points), with the tidal range increasing radially with the distance from the centre. This is typical of the tidal pattern in the North Sea.

The width of the continental shelf plays a major role in determining tidal ranges. When tidal waves enter shallow water, their velocity is reduced due to friction against the bottom. When the water depth and the velocity decline, the height of the tidal wave will increase, so that the total energy flux is maintained. The tidal range, which is thus