the height of the tidal wave, therefore increases inwards across the shelf. Where there are embayments along the coast, tidal waves become focused, so that the tidal range increases towards the middle of the bay. This is true of the east coast of the USA, off Georgia, and of the German Bight in Europe.

If the shelf becomes even wider, as in an epicontinental sea, tidal power will gradually be exhausted due to the high frictional resistance. This is the case on the present Siberian continental shelf where the tide is very low (<10–20 cm). Wind stress may set up rather large storm tides. When the water is very shallow, as on the Bahamas Bank, friction damping takes place over much shorter distances.

The great epicontinental seas, in Ordovician and Cretaceous times for example, were also characterised by low tidal ranges.

About 1/3 of the world's coasts have tidal ranges greater than 4 m (macrotides), 1/3 have mesotides (2–4 m) and 1/3 microtides (less than 2 m). Small inland seas, like the Mediterranean, the Black Sea and the Baltic Sea, are too small to keep pace with the attraction of the moon and the sun, and have small tidal ranges. This is also true of lakes.

1. *Tidal Channels*. The infills may resemble fluvial channels or submarine channels in that they form fining-upwards sequences. Channels formed in estuaries in fact are often connected to fluvial channel systems. Tidal channels which are not part of a river delta, however, tend to be filled with sandy sediments from the surrounding tidal flat, because they have no supply from land. Channels will often erode their banks and cause them to collapse, and this may result in the formation of intraformational conglomerates if the sediments are slightly lithified.

Lateral migration of tidal channels may produce typical epsilon cross-bedding which is the result of lateral accretion of point bars in the tidal channel.

Tidal channels often contain a bed of marine fossils at the base, and marine trace fossils. Channels on tidal flats which are not associated with deltas (i.e. those in estuaries) will not receive much clastic material from land. Conglomerates and breccias in these tidal channels will therefore typically be of the intraformational type, derived by local reworking of tidal flat sediment. Because of their early lithification, carbonate beds in particular can be reworked to form intra-formational conglomerates and breccias.

Tidal cycle duration is about 12 h and 25 min, with currents switching direction every 6 h, and we sometimes find good examples of cross-bedding with opposite current directions. This is not always the case in tidal environments, however. Some tidal channels are dominated by ebb flow and others by flood currents. This is because the ebb and flood often find different dominant pathways. Bipolar cross-bedding is therefore not an essential feature of tidal channels. Storms with opposing wind directions may also produce some form of bipolar cross-bedding.

In a regressive sequence tidal channels will be overlain by lagoonal sediments (Fig. 2.42).

2. *Flaser Bedding*. Consists of clay laminae in a matrix of sandstone with ripple cross-lamination. The clay occurs mainly as infill in ripple troughs, but may also drape over the ripple crests as well. Flaser bedding forms as a result of alternating periods of currents or wave activity, and slack water. The clay settles out in the slack water periods in a tidal environment, though this type of bedding may also form in other environments where there is rhythmic sedimentation, such as in certain fluvial environments. On tidal flats, clay and silt will settle out at high tide to be deposited between ripples formed by the ebb and flood currents. Here the fine sediment may consist partly of clay pellets (faecal pellets from marine organisms) which settle out faster than clay-sized particles. Flaser bedding belongs to a type of structure which we get with mixtures of sand and clay. *Lenticular bedding* represents isolated laminae or lenses of sand in mud.

The inner parts of a tidal shelf often have embayments consisting of very muddy sediments, usually with abundant bioturbation. Mollusc shells in the mud (Fig. 2.43) are often eroded and deposited as shell lag.

3. *Tidal Bundles*. The best identifying feature for tidal environments is regular lamination consisting of fine sand and mud, making up tidal couplets that each represent a tidal cycle. Both modern and ancient tidal sediments show a regular variation in the thickness of such couplets, reflecting the energy levels of spring and neap cycles. Regular laminations reflecting tidal cycles are often called tidal bundles (Fig. 2.44).

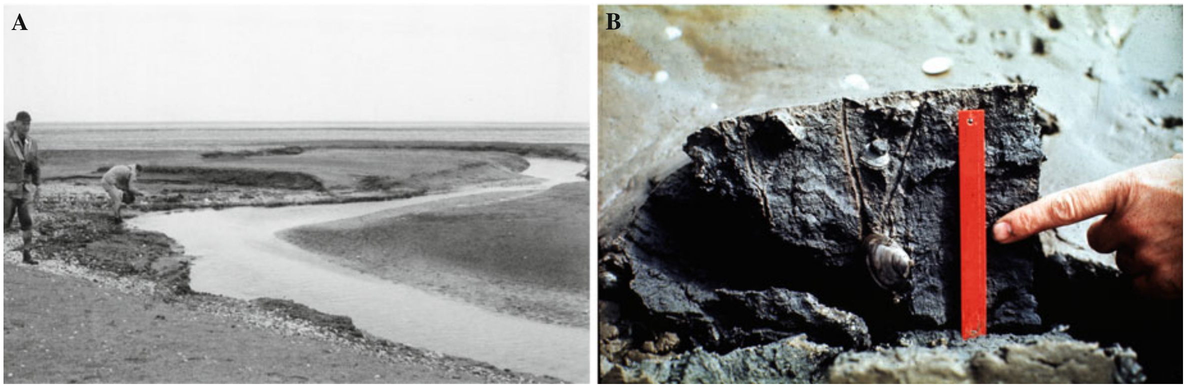

Fig. 2.43 (**a**) Tidal channel on a tidal flat in the muddy facies of the inner part of a tidal flat (near Wilhelmshaven, Germany). Erosion by tidal channels in this mud produces a lag of mollusc shells. (**b**) Mollusc in living position

Fig. 2.44 Tidal bundles. These are bundles of laminae which reflect the tidal cycles between spring tides. From the Late Precambrian Wonoka Formation, Patsy Spring, Flinders Ranges, Australia

2.40 Shallow Marine Shelves

The shelf extends from the nearshore environment to the shelf edge, where there is a rather abrupt increase in slope, usually at a depth of 200–500 m. The width of the shelf varies considerably, and may exceed 1,000 km. Continental shelves are generally very flat areas which may be cut by deeper channels transporting sediments across the shelf from nearshore or deltaic environments.

The study of sedimentation on modern continental shelves is complicated by the fact that sea level was more than 100 m lower only 10,000 years ago. This means that most shelf areas have not yet reached an equilibrium with respect to the modern environment. Sandy shelf deposits are much more difficult to core than muddy sediments and the commonly used gravity corer has to be replaced by vibro-core equipment.

Most shelf areas are below the wave base for normal waves (fair-weather wave base) and sedimentation is governed largely by tidal currents and storms.

When the wind is landward, waves will usually approach the beach obliquely, resulting in wave refraction effects, particularly on relatively steep beaches with high wave energy. This will produce *rip currents*, where the water piling up against the beach has its return flow seawards. Rip currents are often rich in suspended material and transport material from the beach towards the shelf. Another component of wave energy is transmitted parallel with the beach as longshore currents, contributing to the longshore drift of sediment transport.

During onshore storms the sea level near the coast may be raised by several metres due to the combined effect of wind stress, tides and the low barometric pressure associated with storms. The increased potential (elevation) of the coastal water due to these *storm surges* will result in strong bottom currents which are capable of transporting sediment further out onto the shelf. Storm surges may transport fine sand and mud in suspension, but not as true turbidity currents. In this case the increased potential of the coastal water is the driving force and not the density difference between the current and the surrounding water, as with turbidites. Hummocky cross-bedding is a characteristic sedimentary structure produced by the deposition of coarse particles from suspension during storms, and rounded, undulating sand surfaces tend to form. Finer particles will be transported further, into areas with lower energy.

In many modern shelf areas tidal currents are important transport mechanisms, in combination with rare strong storm currents. Sediments deposited in this environment are characterised by rather abrupt transitions from well-sorted sand to mud and from bioturbated to non-bioturbated strata.

The main characteristic of tidal shelves is the mobility of the clastic sediments, on scales ranging from the diurnal tidal cycle to annual (storm augmented) cycles and gradual long-term movement of the largest bedforms. In shelf areas with relatively strong tidal currents (> 150 cm/s) we may get *furrows* and *gravel waves*. At velocities of less than 1 m/s *sand ribbons* may be deposited – longitudinal bedforms developed parallel to the currents. Sandwaves are large-scale transverse bedforms, generally 2–15 m high, with a wavelength of 150–500 m. Sandwaves require current velocities exceeding 60–70 cm/s. Cross-stratification may be symmetrical or asymmetrical on both sides of the sandwave, depending on the relative strength of the opposing currents.

Relatively low-energy environments (< 50 cm/s) are characterised by *sand patches* and mud. The sand forming the patches probably only moves during storms.

Shelf sediments characteristically accumulate at relatively low overall sedimentation rates (1–10 mm/1,000 years).

2.41 Continental Slopes

Continental slopes are the areas between the edge of the mostly very flat continental shelf which commonly lies at a depth of 200–500 m and the *continental rise*, where the ocean deep begins at a depth of 2–4,000 m. The continental slope gradient is typically 2–6°, and it is 20–100 km broad. The gradient is a function of a number of different factors, but the stability of the shelf edge constitutes a major control factor. The steepest slopes are therefore to be found off carbonate banks with well-cemented coral reefs and carbonate beds which have high shear strength. In areas with rapid sedimentation, loose sediments have little shear strength and submarine slides, slumping and formation of turbidites occur, maintaining relatively gentle slopes (1–2°). Where sedimentation is slower the sediments have more time to consolidate, and will be more stable. The steepest submarine slopes in clastic sediments (greater than 10°) are therefore found in submarine canyons, where erosion cuts into older, well-consolidated sedimentary strata.

Along passive continental margins the continental slope is associated with the transition from continental crust to oceanic crust. In areas with a large supply of sediment, the shelf may have prograded beyond this boundary.

2.42 Organic Sedimentation on the Slope of the Continental Shelf

The continetal slopes are enriched in organic matter compared to the shelf and the deep ocean.

This is because the slope is where we have the greatest upwelling of nutrients from the deep. We also find low oxygen content in the water column on continental slopes, allowing much of the organic matter produced to be retained in the sediments. There will also be high productivity on shallower slopes in front of deltas because of the large nutrient supply from river water, but the organic matter may be greatly diluted by rapid clastic sedimentation. On the continental shelf the supply of nutrients is small and the prevalence of stronger currents and turbulence means that most of the organic production there will be oxidised.

Out in the open ocean basin organic production is relatively low, due to a limited supply of nutrients. Much of the planktonic organic matter is oxidised near the seafloor by deep currents of cold, oxygenated water from the polar areas.

Sediments deposited on the continental slope are therefore more promising as source rocks for oil than shelf and deep-water facies.

2.43 Sediment Transport on Submarine Slopes

Gravitational processes are naturally important on submarine slopes.

Gravity forces can be represented by forces (vectors) normal to, and parallel to, the slope. The component parallel to the slope consists of shear forces which may overcome the shear strength of the sediment, causing slumping. Sliding of large volumes of sediments downslope produces extensional faulting in the upper part of the slope and compression in the lower part. Gravitational instability on the slopes may also develop into debris flows and turbidity currents. Collapse of the sediment grain framework may cause sudden compaction and liquefaction of the slope sediments.

Traction currents may, however, also play a part on the submarine slopes, particularly in canyons but also near the toe of the slope, where we may have *contourites* – deposited by currents flowing parallel to the slope contours.

2.44 Submarine Canyons

These are valley-shaped depressions which extend from the top to the bottom of the slopes, down to 2,000–4,000 m. In some cases they may start in shallow water near the beach, in others close to the edge of the shelf. The height from the bottom of the canyon to the top of the slope on each side may be up to 2,000 m. We are dealing with enormous topographical features, which would have been very impressive indeed if they had been on land, towering structures on the scale of the Grand Canyon.

Shepard et al. (1979) systematically gathered data on currents and sediment transport in submarine canyons and found that tidal currents are of great importance *also* at great depths in submarine canyons. Current meters have shown that currents flow *both up and down* the submarine canyons, and that they switch every 6 h like tidal currents in shallow water. Current velocity is often only 10–20 cm/s, but in many canyons velocities of up to 40 cm/s occur sometimes, powerful enough to transport fine to medium-grained sand. The flow velocity tends to be greatest in the upper part of the canyon and diminish downvalley. Most sediment transport takes place during these episodic and unusually high flow velocities which may be linked with storms which create wind-induced shear forces which sweep the water up against the coast *(storm tides)* and may cause currents to develop along the bottom and down the submarine canyons. However, high flow velocities have also been measured without it being possible to associate them with storms or wind stress. The most powerful flow velocity is most commonly directed downvalley, but upvalley-directed streams have been observed with velocities of up to 90 cm/s.

Detailed measurements in the submarine canyons off the coast of California reveal that the highest flow rates are oriented up the canyon in a way which seems to indicate that they are generated by internal waves from the ocean basin, and not by gravitational forces. Currents may then develop when the waves “break” against the coast or the continental shelf.

Powerful currents in submarine canyons are capable of transporting sand, sometimes in rare instances even coarser material. These are frequently not turbidity currents, but *traction currents*, which transport and deposit better-sorted material. Turbidity currents have been observed in submarine canyons as well, but definite (observed) examples were only low-velocity, low-density turbidity currents with a maximum velocity of 70–100 cm/s. In submarine canyons we thus have both traction currents, which are controlled partly by tidal forces, and turbidity currents. Downward-moving currents driven by tidal forces may, if they contain much suspended material, turn into turbidity currents. The downward-moving currents have both a component of traction and gravitation.

The relief of submarine canyons is due partly to erosion down into the underlying sediments, and partly to lack of deposition in the canyon while the adjacent beds were being deposited. During low sea level stands rivers may prograde closer to the shelf edge and hence supply more sediment to the submarine canyons, thus feeding submarine fans. At sea level highstands, currents in the submarine canyons and the shelf may erode shelf and slope sediments and deposit pure sand onlapping an erosional unconformity at the toe of the canyon.

Most of the canyon itself is an area of sediment transport and erosion. Deposition takes place where there is a change of slope near the basin floor. Here the channel defined by the canyons splits up into several channels which build depositional lobes called *suprafan lobes* (Fig. 2.45).