1.4.1 Plate tectonics

Plate tectonics is the theory, supported by large amounts of empirical data, which explains the evolution of the Earth’s outer shell, or lithosphere. The lithosphere is fragmented into a series of smaller segments, known as plates, which move relative to one another. The term plate tectonics also refers to that branch of the geological sciences which studies the morphology and movements of plates, and the various phenomena affecting them. This fragmentation and movement, responsible for the current configuration of the Earth’s crust, generate seismic phenomena and cause the formation of the sedimentary basins which contain almost all the source and reservoir rocks where hydrocarbons accumulate. Life on Earth is made possible by the atmosphere and hydrosphere which were produced and continue to be sourced primarily by the degassing of the mantle through volcanism and other phenomena that are a direct consequence of plate tectonics. As such, the movements of the plates can be said to represent the basis for life on Earth.

In this article, we will outline the essential aspects of plate tectonics. We will first describe the structure of the lithosphere and analyse the data currently available on the movements of the plates (in other words, their kinematics) on the basis of geological and seismological methodologies, and on space geodesy. We will then describe the three main types of plate margins: divergent (or rifting), transform (laterally moving) and convergent (or subductions). We will deal summarily with sedimentary basins, as a function of their geodynamic environment, and then with their nature and origin. Finally, we will examine hypotheses regarding the dynamics and energy sources governing the movement of the plates.

Lithosphere

The lithosphere is made up of the crust and the lithospheric mantle. Since we differentiate between oceanic and continental crust (Fig. 1), the lithosphere is subdivided in the same way. The crust and the lithospheric mantle are separated by the Moho discontinuity. Beneath this, in the mantle, the propagation velocity of seismic P-waves (longitudinal) increases abruptly from about 6.8-7 km/s to about 8-8.2 km/s, and that of S-waves (transverse) from 3.9 km/s to 4.5 km/s. We do not have sufficient data to determine the extent to which oceanic lithospheric mantle differs from continental lithospheric mantle; we therefore assume a peridotitic composition for

Fig. 1. Schematic stratigraphy of the crust and continental and oceanic lithosphere.
both, with a density of about 3.3 g/cm³. The lithosphere thus starts at the Earth’s surface and reaches down to the isotherm of about 1,300°C; above this temperature, the mantle begins to melt partially. This marks the boundary with the zone known as the asthenosphere (from the Greek ἄσθενος for “weak”), or ‘low velocity channel’ where, as a result of the partial melting of the mantle, P-waves and S-waves slow respectively to velocities of 7.9 km/s and 4.4 km/s. The base of the lithosphere is thus interpreted mainly as a phase transition, rather than as a chemical variation (Fig. 2).

The oceanic lithosphere is thinnest next to mid-ocean ridges (about 10 km) and thickens as it moves away up to about 100 km. This distance corresponds to an increase in ocean depth. The older the oceanic crust, the deeper the seafloor. We therefore believe that the 1,300°C isotherm, marking the base of the oceanic lithosphere, sinks as the lithosphere cools and spreads away from the mid-ocean ridge. Consequently the seafloor also sinks, due to the higher density of the lithosphere. During the first 10 My (million years) after its formation, the seafloor subsides by about 1,000 m as it spreads away from the mid-ocean ridge; during the following 26 My it subsides by a further 1,000 m. This variation is described by the simple formula \( z = k \cdot E \) where \( z \) is the difference in depth between the mid-ocean ridge and the seafloor.

![Fig. 2. Model of the Earth where the various shells correspond to physical discontinuities allowing them to slide past one another, similar to that between the molten outer core and the inner core, whose differential rotation generates the Earth’s magnetic field. The lithosphere behaves in an elastic manner, whereas the mantle exhibits visco-elastic behaviour, giving it the ability to flow if subjected to stress over a long period of time. The convective motions postulated for the mantle therefore take place in the solid state. There are two large areas in the lower mantle which show relatively low seismic velocities beneath the central Pacific and Africa.](image-url)
expressed in metres, \( k \) is a constant equal to about 320 and \( E \) is the age of the oceanic crust expressed in My. This important relation, known as the Sclater curve, allows us to calculate the depth of the sea below a mid-ocean ridge up to the age of about 60-80 million years. Over this age, seafloors no longer appear to sink as a consequence of thermal effects. Heat flow diminishes with distance from the mid-ocean ridge (Stein, 1995) and the velocity of seismic S-waves increases; these facts indicate a decrease in ‘melt’ in the underlying mantle. 

The oceanic crust is about 5-8 km thick and has an average density of 2.9-3 g/cm\(^3\). It is made up of three layers, not all of which are necessarily present, forming a sequence from bottom to top of: a layer of gabbro, a layer of dikes, and an upper layer of lavas, pillow lavas and oceanic sediments. 

The continental crust, being less dense (about 2.7-2.8 g/cm\(^3\)), is thicker than the oceanic crust, with the Moho at an average depth of about 30-40 km. The crust thickens beneath cratons and orogens to about 15 km. Continental crust consists of a sequence from bottom to top of: a lower felsic crust, generally stratified by magmatic and metamorphic processes; an upper crust, mainly consisting of rocks of varying metamorphic grade and granite intrusions due to earlier orogenies; a sedimentary cover whose thickness varies from 0 to 15 km. The sedimentary cover consists of sediments deposited during eustatic raises or epeirogenic subsidence within cratons, or of intraplate or passive continental margin syn-rift sediments. In the proximity of orogens, the upper layer is composed of foreland basin sediments (flysch and molasse).

The age of the oceanic crust ranges from 0 to 180 My (Fig. 3), whereas the continental crust may reach ages of over 3,900 My. This is a consequence of the extreme mobility of the oceanic crust, which forms rapidly in mid-ocean ridge zones and disappears equally rapidly in subduction zones due to its higher density. The lighter continental crust, on the other hand, subducts less easily into the mantle and thus remains floating on the surface, growing slowly to increase the area dimensions of the continental lithosphere, which has an average thickness of about 100-150 km, up to a maximum of about 200-250 km beneath the major cratons (Windley, 1995; Gung et al., 2003).

The lithosphere is subdivided into plates; a plate is a segment of lithosphere characterized by its independent motion relative to the adjacent lithosphere. The major plates are: the North American, South American, European, African, Arabian, Indian, Australian, Antarctic, Pacific and Nazca plates. There are other smaller plates such as the Philippine, Cocos and Juan de Fuca plates. Plate tectonics is generated by the different velocities among the plates. The movement of the plates towards or away from one another is governed by the relationship, or the degree of coupling, between the lithosphere and the underlying mantle. The Earth’s seismicity is manifested only within the lithosphere, and disappears at a depth of 670 km. This is the maximum depth at which subduction zones can be detected, at the transition between the upper and lower mantle.

**Plate kinematics**

One fundamental aim of tectonics is to determine the depth of decollements (or decoupling zones). Along decollements, upper and lower zones slide past one another. In plate tectonics, the main decollement lies at the base of the lithospheric mantle, coincident with the asthenosphere. This part of the mantle has the lowest average viscosity, generally between 10\(^{17}\) and 10\(^{19}\) Pa·s, and locally as little as 10\(^{15}\) Pa·s where the asthenosphere is hydrated. There are various structures within the asthenospheric decollement which may
explain the differences in velocity between the plates above, in other words their relative motion.

Faults are surfaces of fracture and movement of the brittle part of the crust, which behaves in a mainly elastic way. These may be horizontal (i.e. decollements) or inclined up to 90°. The rock above a fault is known as the hangingwall; that below a fault is known as the footwall. When the hangingwall moves up relative to the footwall, the fault is known as a reverse fault; it is described as a thrust fault if it has an average inclination of about 30°. If the hangingwall moves down relative to the footwall, the fault is known as an extensional or normal fault, and has an average inclination of 60°. When hangingwall and footwall are indistinguishable because the fault is vertical and movement purely horizontal, we speak of a transcurrent or strike-slip fault. At the crustal level, the depth of the decollement determines the spacing between faults: the more superficial the decollement, the closer together the faults, and vice versa.

In subduction zones, where one plate sinks beneath an immediately adjacent plate, accretionary prisms are formed. These are mainly a combination of thrusts and folds, which pile up and warp the rocks of the plates above and below the subduction zone (overriding and downgoing plates). Accretionary prisms thicken in the direction of subduction, taking on a wedge shape; they are therefore also known as accretionary wedges. The deeper the basal decollement, the greater the volume of the accretionary prism. The term accretion refers to the transfer of rock from the downgoing to the overriding plate, where the accretionary prism itself is located. The expression tectonic erosion, on the other hand, refers to the transferral of the decollement into the overriding plate, thus temporarily subducting fragments of the downgoing plate. In this case there is no accretion; this type of mechanism has been suggested for some sectors of the Andean subduction zone.

In areas where plates are moving apart (or rifting zones) the asthenosphere also seems to be the main basal decollement.

Plate movements

The movement of the plates is obvious both from tectonic structures (Fig. 4), and from seismicity and geodetic measurements (Fig. 5). Space geodesy has confirmed that the relative movement of the plates is often distributed in a zone at their margins, with a width varying from 10 km to several hundred km, through numerous active faults which absorb the deformation. Generally speaking, transform margins are narrower than convergent margins. Past movements are recorded by the formation of orogens along subduction zones, indicating that the plates have converged, and by the specularity of magnetic anomalies in oceanic rifts. The movements of the plates can be analysed in relative terms, between pairs of plates; however we can also attempt to examine
them in terms of absolute movements, using independent reference systems such as hotspots, the fixed stars, or the Earth’s centre of mass.

The movement between two plates may occur at any angle, creating a full range of tectonic environments (i.e., compressional, transcurrent or extensional), and all situations in between where plates converge with a lateral or transcurrent component (transpressional environments), or diverge with a transcurrent component (transtensional environments). Movements currently measured using space geodesy are of the same order of magnitude as those which can be deduced for the geological past from the study of magnetic anomalies in the oceanic crust. Consequently, despite slight oscillations in long wavelength velocities, the movements of the plates can be considered stable over time. It remains to be said that plate margins are born and die, modifying or abolishing velocity gradients.

Since the plates are moving over a sphere, the relative motion of two plates can be described using Euler’s fixed point theorem (according to which the motion of a portion of a spherical surface can be represented as a single rotation around a fixed point). Specifically, by identifying the rotation pole of this relative motion, we can calculate the increase in linear velocity as distance from the pole increases (Fowler, 1990). However, in nature two plates may have a rotation pole which is not fixed, especially where one of the two plates also has an independent sub-rotation.

If we consider the movements of the plates, which can be deduced for at least the last 50 My from structural data such as rift zones, transform zones and orogens, we can conclude that the plates do not move randomly, but follow a global flow. This flow has a general undulation (see again Fig. 4) describing a sort of tectonic equator, although this seems to represent a sinusoid rather than a great circle. The lines of flow represent the mean direction of plate movements. Along oblique plate margins (transtensional or transpressional environments) the stress field is deviated and is not parallel to either the relative or absolute movement of the plates. For example, the Arabian Plate is moving NE-SW, the Red Sea Rift is a sinistral transtension and the Gulf of Aden a dextral transtension.

The flow is characterized by a gradual change of direction in plate movements from WNW-ESE in the Pacific to E-W in the Atlantic, subsequently turning SW-NE across Africa, India and Europe. It then turns back towards the Pacific direction of flow. The largest area of continental lithosphere (Eurasia) is concentrated where this flow tends to bend towards the Pacific. The flow of plates deduced from tectonic data is confirmed by space geodesy in the summary map of GPS (Global Positioning System) stations developed by NASA (National Aeronautics and Space Administration; see again Fig. 5). Vectors in particular have confirmed the SW-NE movement of both Africa and Europe. On this map, plate movements are referred to the Earth’s centre of mass, conventionally considered to be in line with the constellation of GPS satellites. This is the reference system known as ITRF (International Terrestrial Reference Frame), in which it is assumed that there is no net rotation of the

Fig. 5. Current plate movements based on space geodesy, postulating the absence of a differential rotation of the lithosphere relative to the mantle. Satellite data largely confirms this undulating flow, interpreted on a tectonic basis.
lithosphere relative to the Earth’s underlying interior (no-net-rotation).

In fact, if we analyse plate movements using other reference systems, such as hotspots or the Antarctic, the lithosphere does have a net rotation relative to the mantle, with a mean westward direction. This is particularly evident if we consider the velocity at which the Pacific travels WNW, so high that the sum of the movements of all the other plates cannot compensate for it, thus determining a residual westward movement.

The movement of the plates is faster in the equatorial and tropical belts, as indicated by space geodesy, earthquakes and magnetic anomalies for past movements. The flow of the plates, its westward polarization and the greater velocity of plates at low latitudes suggest that plate tectonics is influenced by the Earth’s rotation. The concentration of the mantle also seems to support this hypothesis; it is colder and heavier in the equatorial belt. The drift to the west, or more accurately along the tectonic equator, is also evident from surface geology, such as the asymmetries of mountain belts on the western and eastern Pacific margins (see again Fig. 4), the arcs of westward-dipping subductions indicating the presence of an obstacle to flow in the opposite direction, and the asymmetry of rifting zones.

**Hotspots**

Hotspots are important for an understanding of the internal dynamics of the Earth, and are particularly useful to measure plate movements with respect to the reference frame which they themselves represent. There are areas of enormous lava emission on both the continental and oceanic lithosphere, where several million cubic metres of basalt are erupted over the space of a few million years, such as the basalt traps of Paraná in Brazil, the Deccan Traps in India, or the Ontong-Java Plateau in the south-western Pacific (LIPs, Large Igneous Provinces). Their origin is not clear, either as regards the depth of their source or the dynamics of the process. There are also magmatic events which describe linear tracks over the Earth’s surface, both subsea and subaerial, which become more recent in a given direction. These lines are known as hotspots, and can be found both within plates and at plate margins. The most typical examples of intraplate hotspots are the Hawaiian-Emperor chain – whose age ranges from over 70 My to the current active volcanism of Mauna Loa, with an intermediate bend in the migration at about 47 My – or the islands of Luisville and MacDonald, also in the Pacific Plate. Other typical examples of hotspots which have created chains of progressively younger volcanoes located near plate margins are Iceland, the islands of Ascension and Tristan da Cunha along the Mid-Atlantic Ridge, or Easter Island near the East Pacific Rise. There are various schools of thought as to the origin of hotspots: among which they are sourced from the lower mantle, or from the upper mantle. Whatever the depth of their source, hotspots indicate that the lithosphere and asthenosphere are moving relative to one another. According to other studies, hotspots originate as a result of excess heat produced by radioactive decay or the heat of the Earth’s core welling upwards along a path of least resistance. Another possibility is the presence of more fluids, lowering the melting temperature and thus generating greater magmatism at lower temperatures. In the latter case, hotspots are also known as wetspots since the mantle is not hotter than normal, but simply has a higher water content. This model may provide a concrete explanation for the existence of hotspots along mid-ocean ridges. One interpretation of intraplate hotspots suggests that magmatism is generated by the heat of viscous friction in the decollement of the astenosphere between the lithosphere and the sub-asthenospheric mantle.

Hotspots thus represent an important reference system for the study of plate movements. Specifically, hotspots within the Pacific Plate have remained fixed relative to one another for at least 5 My. This gives us a reference point in the mantle to study the relative movement of the lithosphere; the relative movements of plates can be recalculated using this reference system, which conventionally does not assume the absence of a differential rotation between lithosphere and mantle. Using the hotspots reference system, Gripp and Gordon (2002) have observed that the lithosphere has a net westward rotation of about 50 mm/y relative to the mantle, with a pole of rotation at 56°S 70°E. However, this calculation also takes into account hotspots located at plate margins, postulating that these are sourced from the lower mantle. If we consider only the hotspots within the Pacific Plate, and assume that the source of magmatism is located in the decollement due to the heat of friction, the westward drift of the lithosphere relative to the mantle is much greater, roughly double. This means that the flow of plates in Fig. 4 has a mean westward orientation, in other words all the plates move along this sinusoidal flow, but at different velocities. Velocity gradients, determined by the degree of decoupling from the mantle, generate the various types of plate margin and plate tectonics. The lower the viscosity of the asthenosphere, the faster the overlying plate moves westward. The viscosity of the asthenosphere is lowest beneath the Pacific (5×10¹⁷ Pa·s) and the Pacific Plate moves most rapidly WNW (>100 mm/y). Therefore lateral variations of the asthenosphere viscosity, and variations of the asthenospheric and lithospheric thicknesses, should control the different plate velocities. When a plate
moves westward faster than the plate to its east the plate margin is divergent; if the former moves more slowly the margin is convergent.

**Rifting zones**

Rifting zones are areas where the lithosphere separates into two plates that diverge from one another. The continental rifting stage is extremely slow, with rates of horizontal extension in the order of 0.1-0.3 mm/y, and may last for long periods (30-50 My or more). The process of extension (rifting) initially involves a lengthening and flattening of the continental lithosphere. This process can be quantified by dividing the initial thickness of the lithosphere by its thickness when flattened a ratio known as the $\beta$ factor (McKenzie, 1978). For example, an area of lithosphere 100 km thick, subjected to tension and reduced to a thickness of 20 km, has a $\beta$ factor of 5. This means that the higher the value of $\beta$, the greater the thinning and the rise of isotherms, and consequently the heat flow.

The continental rifting stage is accompanied by growth sedimentation, with a typical tripartite sequence from bottom to top of: fluvial sandstones, evaporite deposits and carbonate sediments. This sequence is evidence of the gradual entry of the sea into the thinned zone of continental lithosphere;

![Fig. 6. Comparative models of rifting:](image)

**Fig. 6.** Comparative models of rifting:
A, pure shear (McKenzie, 1978); B, simple shear (Wernicke, 1985); C, detachment (Lister *et al.*, 1986).

subidence is thus generated by the simultaneous rise of the denser asthenosphere from below.

Models of lithospheric extension can be divided into pure shear, simple shear and detachment models (Fig. 6). In pure shear, the lithosphere thins instantaneously and symmetrically, subsequently undergoing thermal cooling with accompanying subsidence (McKenzie, 1978). In simple shear, the lithosphere is cut by a large-scale low-angle fault, with one plate overlying and another underlying the extension, giving the rift a strongly asymmetrical component (Wernicke, 1985). Isostatic uplift of the underlying plate, and an axial disalignment between the superficial extension and the uplift of the underlying mantle have been suggested. Other models combine the two described above (Buck *et al.*, 1988), or involve detachment (Lister *et al.*, 1986) where the shear zone presents decollements between the brittle upper crust and the ductile lower crust, and between the latter and the lithospheric mantle.

Rifts do not always evolve into oceanic rifts; in other words, they may abort or even become recompressed, giving rise to tectonic inversion structures (as in the North Sea). Alternatively they may lead to the complete fracture of the continental lithosphere, allowing the formation of new oceanic crust; for this reason divergent margins are also known as constructive margins. In this case, passive continental margins are formed; these may develop in the presence of extensive magmatism or grow in an almost complete absence of volcanism; we therefore speak of volcanic and non-volcanic continental margins. For example, the Atlantic margins of Brazil and Greenland are classic volcanic margins, since during the Cretaceous and Cenozoic rifting they were accompanied by large-scale emissions of magma. Variations in syn-rift magmatism may be due to chemical and thermal heterogeneities in the mantle, or to the variable presence of water, an abundance of which causes a decrease in the melting temperature of mantle rocks, and thus an increase in lava production.

At the point where two plates are separating, the underlying mantle rises to compensate isostatically for the mass deficit (Fig. 7). This upwelling, considered adiabatic, decompresses the mantle and allows it to melt. The magmas of rift zones have characteristics ranging from alkaline to tholeiitic.

The transition from continental to oceanic rifting is also known as *breakup*. Sedimentation within the passive continental margin is marked by the so-called *breakup unconformity*, an unconformity which buries the main extensional growth structures, and which signals and dates not only the birth of the new ocean, but also the passive continental margin’s transition from tectonic subsidence to thermal subsidence, and
Thus from a state of rifting to one of drifting. The tectonic and thermal subsidence of the margin takes place at a low rate (0.1 mm/y). The transition from continental to oceanic rifting causes an enormous acceleration (100-1,000 times) in extensional velocity, which passes from rates of continental extension of 0.1 mm/y to rates of oceanic extension of 10-100 mm/y.

The development of new oceanic crust takes the form of a sort of ‘new skin’ generated by the mantle as it nears the surface. Mid-ocean ridges can be divided into three types, depending on their velocity: slow (Mid-Atlantic Ridge, 20 mm/y); intermediate (Indian Ridge, 30-50 mm/y); and fast (East Pacific Rise, >100 mm/y). Slow mid-ocean ridges generate a rift valley and have a more elevated and jagged topography, whereas fast mid-ocean ridges have no rift valley, are less elevated and have a gentler morphology. The Mid-Atlantic Rift Valley is also more irregular in morphology and is characterized by the presence of numerous extensional faults.

Various ocean basins have opened along thickenings in the lithosphere generated by previous orogens. For example, the central and north Atlantic appeared where the Palaeozoic Appalachian mountain belt had previously developed. The oceans then closed, completing the Wilson cycle, which postulates that rifts are created at the location of earlier subduction zones and that orogenic belts close earlier rift zones. This indicates that rift zones are caused by heterogeneities in the lithosphere and their interactions with the underlying asthenosphere, apparently independent of lower mantle processes.

We can distinguish between various types of rifting on Earth, alongside the linear rifts which produce the major ocean basins. These include the back-arc basins which form over W-directed subductions, characterized by high rates of subsidence (0.6 mm/y); these are associated with the eastwards retreat of the subducting slab. Examples are the Caribbean, the western Mediterranean, the Pannonian Basin and the Japan Sea.

Extensional tectonic episodes may also occur in accretionary prisms when the critical angle of repose is exceeded. However, these extensional faults have a superficial decollement (in the upper few km), whereas extensional faults in classic rifts have decollements in the brittle regime of the upper crust, and the ductile regime of the lower crust, reaching the base of the lithosphere at the boundary with the asthenosphere.

In continental margins and back-arc basins major faults seem to be evenly spaced, with two peaks in average spacing at 25-30 km and 4-6 km. Rifts may be concentrated in a few km (for example the East African Rift Valley which crosses the entire length of East Africa, but has an average width of a few tens of km), or may be several hundred km wide (such as the Basin and Range in the western United States).

Studies of ophiolites, remnants of oceanic crust embedded in orogens, and the polarization of seismic S-waves in the mantle indicate that olivine crystals tend to lengthen parallel to the direction of extension. This confirms the hypothesis that there is a significant decollement between lithosphere and asthenosphere which causes the iso-orientation of the crystals, as is also shown by the deformed xenoliths of asthenospheric mantle found in lava.

Asymmetry ascribable to geographical polarity can also be seen in rift zones, where the eastern flank is on average 100-300 m higher than the opposite flank, both in subsea and subaerial environments. The explanation advanced for this asymmetry is that the

Fig. 7. Model of an oceanic rift.
The left-hand plate is more strongly decoupled from the asthenosphere, so that it moves westward faster than the right-hand plate, thus creating a rift. The mid-ocean ridge moves relatively westward. The upwelling of the asthenosphere compensates for the separation of the two plates. As it wells up and becomes decompressed the asthenosphere melts, producing new oceanic crust-lithosphere. The residual asthenosphere is lighter and in its westward motion generates a mass deficit responsible for the lower depth of the eastern side of the ridge, and later for an uplift of the continental lithosphere to the right (Doglioni et al., 2003).
mantle which melts beneath a mid-ocean ridge loses iron and other elements which melt more quickly. The residual mantle thus becomes lighter by about 20-60 kg/m³, passing for example from 3,400 kg/m³ to 3,360 kg/m³, and moving eastwards underneath the lithosphere. The presence of lower density mantle underneath the eastern sides of rifts indicates a mass deficit offset by a corresponding uplift which, in the flank of a mid-ocean ridge, slightly decreases thermal subsidence. Asthenospheric mantle lightened by partial melting beneath a mid-ocean ridge causes isostatic uplift when it passes underneath a continent and replaces denser asthenosphere. This mechanism may explain the uplift of Africa, France or India as a result of the passage beneath the continental lithosphere of lighter asthenosphere, depleted beneath the Mid-Atlantic Ridge or the Indian Ridge (see again Fig. 7).

**Transform zones**

Plate margins moving roughly parallel to the relative movement between two plates are described as transform boundaries, where the prevalent tectonics is transcurrent. These margins probably have a decollement at the base of the lithosphere. Transform faults, also known as transcurrent or conservative margins, may develop in both continental and oceanic lithosphere. A typical continental example is the sinistral transcurrent fault of the Dead Sea which separates the Arabian Plate from the African Plate. Oceanic examples include the Romanche and Vema fracture zones in the central Atlantic, with dextral transcurrent, separating the African Plate to the north from the South American Plate to the south. Oceanic transform faults are among the Earth’s longest tectonic structures; they may be several thousands of km long. As a consequence of the convergence of lithosphere of different ages, and thus with different thermal states and bathymetry, bathymetric differences of 2-4 km between the two sides of the fault may develop along its length. Complete sections of oceanic crust may be exposed along these submarine escarpments, with their corresponding basal Moho and transition to the underlying mantle (Bonatti et al., 2003).

In some cases, oceanic transform faults result from the irregular propagation of continental rifting, which follows the weakest zones of the lithosphere. This is the case, for example, in the Romanche transform fault which reflects a large undulation in the Mid-Atlantic Ridge, exemplified by the large promontory of north-west Africa. Other smaller transform faults form near mid-ocean ridges, without corresponding undulations on continental margins; their origin seems to be linked mainly to the intrinsic dynamics of oceanic rifts.

Undulations along transcurrent faults create local transtensional depressions such as pull-apart basins, or uplifts in transpression zones, such as push-up structures. It has been noted that rates of magma production in rift zones are proportional to the velocity of expansion. As the angle of a mid-ocean ridge with respect to the movement of the plates gradually decreases until it becomes inserted into a transform zone, magmatism gradually disappears because the rate of expansion in a pure transform margin is zero.

In terms of energy, transform faults are passive structures, which apparently do not contribute actively to plate tectonics, unlike the phenomena of ridge push for mid-ocean ridges and slab pull in subduction zones.

The San Andreas Fault in California is frequently cited as an archetypical example of transform and transcurrent faults. However, this fault has unique and unusual geodynamic characteristics when compared to typical transform faults, and cannot be considered a classic example of a transcurrent zone. This fault, with its associated fault system, forms the belt where the North American Plate interacts with the Pacific Plate, along the zone where the Pacific Ridge transfers from the Juan de Fuca Ridge to the northwest (Mendocino transform fault) to the East Pacific Rise to the south-east.

As is well-known, this plate margin is a dextral transpression zone, where transcurrent dextral movements occur alongside thrusts parallel to transcurrent, as indicated by geological data and the focal mechanisms of earthquakes.

The Pacific Plate is moving in a direction of 300°, forming an angle of about 25° with the San Andreas Fault, which has a direction of 325°. Since the Pacific Plate is moving WNW faster than the North American Plate, the angle between the fault and the direction of the Pacific should generate a dextral transtension rather than a transpression. However, the zone where the Pacific Ridge transfers from the Juan de Fuca Ridge to the East Pacific Rise in the Gulf of California is moving WNW more slowly than the North American Plate, which is thus able to overthrust obliquely towards the west onto the Pacific Plate, with a sinistral tranpressional component.

The dextral tranpressional tectonics of the San Andreas system can therefore be subdivided into two components: sinistral transtension along the oblique western margin of the North American Plate, responsible for most compressional earthquakes, and overthrusting of the North American plate onto the dextral transtensional transfer zone of the Pacific Ridge. Since the dextral transtension is faster than the sinistral tranpression, the dominant movement is
dextral. This unusual situation is due to the oblique directions of the margins of the Pacific and North American plates relative to their absolute motion, and the different velocities of the three elements in play: Pacific Plate, the transfer zone of the Pacific Ridge, and North American plate.

Californian geodynamics is thus characterized by an unusual subduction in which, in contrast to normal subduction zones, in E-W section the downgoing plate of the subduction diverges from the overriding plate, whilst overriding and downgoing plates converge, albeit more slowly, in a NE-SW direction. The E-W divergence is absorbed by the extension in the Basin and Range, whereas the NE-SW compressional component is expressed mainly in the overthrusts and transpression of the Coast Ranges and the Californian offshore. This suggests that the compression perpendicular to the San Andreas Fault is not the natural consequence of a transcurrent movement, but rather an independent geodynamic causes, may coexist in a single area; in this case, the sinistral transpression and the faster dextral transtension.

**Subduction zones and orogens**

Convergent, or destructive, margins are created when a plate sinks, or is subducted, into the mantle. The subducting lithosphere is known as a slab. Orogenic or accretionary prisms are formed in association with subduction zones (Bally, 1983); these are characterized by a series of parameters such as the dimensions of the mountain belt, rates of uplift and shortening, the extent of erosion, etc. An example of the front of a mountain belt is the accretionary prism of the Apennines, located on the hinge of the subduction zone of the same name (Fig. 8). Generally, subductions form when two plates converge, and the heavier of the two, usually an oceanic plate, begins to penetrate the asthenosphere (Fig. 9). According to the classification drawn up by Bally et al. (1985), we speak of B-subduction for oceanic lithosphere (named after its discoverers H. Benioff and K. Wadati), and A-subduction for continental lithosphere (named after its discoverer O. Ampferer). Most of the Earth’s seismic energy (>90%) is released along subduction zones; for example, the ten largest earthquakes of the 20th century occurred in the circum-Pacific subduction zones (eight) and in the Himalayan and Indonesian subductions (two). The most powerful earthquake ever recorded took place along the Chilean subduction zone in 1960 with a magnitude of 9.5. This is because fracturing rocks under compression requires much more energy than fracturing rocks under tension. Furthermore, subduction zones, in contrast to rifts, are cold zones, where the lithosphere exhibits more brittle behaviour, and thus higher resistance to deformation.

Current subductions have convergence velocities ranging from 1 to 120 mm/y. However, there are also active subductions in the absence of convergence; this means that the slab still retreats, but only in W-directed subductions (such as the Apennines and Carpathians).
The deep zones where the subduction is interrupted or absent are known as slab windows. These may form as a result of the lengthening of the slab as the subduction becomes arched, or of the different subduction velocities of two underlying plates. An alternative interpretation explains slab detachment on the grounds of weight.

Subductions have a retreating hinge, whose velocity may be greater or lower than the convergence velocity between the two plates above and below the subduction. If the overlying slab has a lower velocity of convergence than the retreat of the slab, a back-arc basin is formed (for example, the Japan Sea is the back-arc basin of the subduction of the same name, the Tyrrhenian Sea and the entire western Mediterranean are the back-arc basin of the Apennine-Maghrebid mountain belt, and the Pannonian Basin is the back-arc basin of the Carpathians). This, too, is a situation which appears to occur only in W-directed subductions. By contrast, if the rate of convergence is greater than the retreat of the slab, as is frequently the case, an extremely high two-sided orogen is formed (such as the Alps). In the first case an accretionary prism forms, bringing with it a wave of extensional tectonics which
causes the rifting of the back-arc (Fig. 10). The pairing compression-extension in W-directed subductions is replaced by compression-compression in mountain belts created by E- or NE-directed subductions, creating typical two-sided orogens. Extensional tectonics may shape the upper part of these mountain belts when the critical angle of repose is exceeded.

If the overlying plate is continental, and two plates are converging, the transition from oceanic to continental subduction is known as the collision phase. Magma in subduction zones varies in character from calc-alkaline to shoshonitic. Magmatism is found in vertical projection from the isobath of about 100-130 km of the subducting slab, and is thought to be generated by the fluids released by the subduction zone, leading to the partial melting of the overlying mantle. The number of volcanoes and the volume of magma erupted are proportional to the velocity of subduction. This suggests that the heat of friction may contribute to the production of magma. Magmatism is conditioned by the composition of the subducting lithosphere, the thermal state of the slab, its angle of dip and thickness.

W-directed subductions are on average more recent than 50 My, whilst E-directed subductions may be older than 100 My. In W-directed subductions the subducing lithosphere is thin (20-40 km), whereas the underlying plate is always thicker (see again Fig. 10). The Moho in the overriding plate is generally newly formed, migrates eastwards and develops during the growth of the back-arc basin. The crust of the overriding plate is thinned and has a depth of 10-25 km. By contrast, the Moho in the underlying plate is pre-existing and of variable age. In mountain belts linked to E- or NE-directed subductions, on the other hand, the pre-existing Mohos in the two plates are superimposed beneath the orogen (see again Fig. 10) and the thickness of the crust reaches its maximum values (55-70 km).

W-directed subductions nucleate along the back-thrust belts of E- or NE-directed subductions when thin oceanic or continental lithosphere is present in the foreland of the back-thrust belt. For example, the Lesser Antilles islands arc began to form along the back-thrust belt of the Central American Cordillera, and migrated eastwards only where the North American and South American continents narrowed; Atlantic oceanic lithosphere was present at the front of the Central American orogen’s back-thrust belt.

A similar interpretation can be proposed for the Apennines, which originated along the back-thrust belt of the Alps; the Alpine foreland contained a relict branch of the Mesozoic Tethys Ocean. These ‘Palaeo-Alps’ are now buried and flattened beneath the western Apennines and the Tyrrhenian Sea, which is the back-arc basin of the Apennine subduction. A similar relationship may apply to the Carpathian subduction, which originated along the back-thrust belt of the Dinarides. In back-arc basins, rapid and irregular thinning takes place with areas where new oceanic crust develops or areas where thicker remnants of continental lithosphere remain. This results in the phenomenon known as boudinage, or a situation where during extension more competent blocks are held within a less viscous matrix, which flows in the thinned areas (necks). The arcs of W-directed subductions are 1,500-2,000 km long.

W-directed subductions are on average deeper, up to 670 km, and more steeply inclined (45°-90°) than E- or NE-directed subductions (see again Fig. 9); in the latter most seismicity generally disappears at 300 km, and the angle of dip is lower (15°-60°). The westward drift of the lithosphere relative to the underlying mantle may explain this difference in dip angle, which in the past was attributed exclusively to the different age of the subducting oceanic lithosphere; in other words, as an effect of the weight of cold oceanic lithosphere. However, there are instances where the same lithosphere subducts in two opposing directions yet maintains this asymmetry; there are also W-directed subductions characterized by a steep dip angle and the features described above which involve both young oceanic lithosphere (for example, the Sandwich Islands arc in the south-west Atlantic) or even continental lithosphere (the central and northern Apennines, the Carpathians, and the Banda Arc). In W-directed subductions, the basal decollement of the overriding plate is folded and subducted, and the accretionary prism involves only the upper surface of the underlying plate. In E-directed subductions, the basal decollement of the overriding plate actively carries upwards elements of both the underlying and overriding plate, thus thickening the crust and the corresponding orogen (see again Fig. 9). The differing behaviour of decollements in these two types of subduction would explain why accretionary prisms in W-directed subductions are composed mainly of sedimentary cover whereas in the orogens of E-directed subductions the whole crust is warped, causing a higher structural elevation of the mountain belt, and substantial outcrops of crystalline basement (see again Fig. 10). The different behaviour of decollements in these two types of subduction also leads to variations in the pressure and temperature to which accretionary prism rocks are subjected, generating unusual metamorphisms. For example, high pressure and low temperature metamorphism is more frequent in mountain belts associated with E- or NE-directed subductions, whereas high temperature and low pressure metamorphism is more frequently found above W-directed subductions, where the asthenosphere replaces the slab at a shallow depth in the back-arc basin.
The strongest evidence that the lithosphere is drifting westward, and therefore that the underlying mantle is rotating in the opposite direction, comes from the persistent asymmetries between W-directed subduction zones and those directed E or NE. Orogens associated with W-directed subduction zones have a lower topographical and structural profile than mountain belts associated with E-directed subductions. This is clear if we compare subductions in the western Pacific with those in the eastern Pacific, for example the Marianas and the Andes (see again Fig. 4).

In the first case, a back-arc basin is formed, and the subduction trench is extremely deep, on average over 4,000 m; the accretionary prism involves the upper layers of the subducting crust, in general the sedimentary cover. On average, prisms in this type of subduction are below sea-level, as in the islands of Fiji, the Marianas and Barbados. The highest mountain belts in this type of subductions are the Apennines, the Carpathians and the mountains of Japan, where the accretionary prisms have deeper basal decollements, and the volumes involved above the subduction are greater. Gravimetric anomalies in W-directed subductions are much more pronounced than those in E-directed subductions, with a negative maximum in the foreland basin and a positive maximum in the back-arc basin, where the asthenosphere reaches layers very close to the surface. A similar pattern characterizes variations in heat flow, lowest in the foreland basin and highest in the back-arc basin.

In E- or NE-directed subductions such as the Andes or the Himalayas, there is no back-arc basin; the mountain belt is double-sided and therefore has two foreland basins, in front of the frontal belt and one in front of the back-thrust belt (see again Fig. 10). On average, these mountain belts are above sea level; the foreland basins have an average depth of about 3,000 m in oceanic subductions, and are often above sea-level in continental subductions, on both sides of the orogen. The mountain belt has decollements which enter the mantle, the entire crust is involved in accretion, and the surface rocks thus cover the whole spectrum of metamorphic and intrusive basement rocks.

Topography and free-air anomalies across subduction zones confirm the presence of two distinct signatures (Fig. 11). Low average topography (−1,250 m) and marked gravimetric anomalies characterize the mountain belts of W-directed subductions. A higher average topography (1,200 m) and less marked gravimetric anomalies are typical of orogens in E- and NE-directed subductions. This contrast is particularly obvious along the Pacific margins, but can also be seen along other subduction zones in the world; in the Atlantic, the Mediterranean, the Himalayas and in Indonesia. Thus topography and gravimetry confirm the existence of two separate classes of subduction zone, largely independent of the age and nature of the subducting lithosphere.

**Foreland basins**

Foreland basins are the sedimentary basins located at the front of mountain belts or accretionary prisms. The characteristics of foreland basins also confirm the differences between subduction zones. W-directed
Subductions have extremely deep foreland basins, which migrate rapidly eastward, and rates of subsidence of >1.2 mm/y. Subsidence is so strong that the accretionary prism’s thrust anticlines may have negative rates of uplift; as such, the anticlines may be subsiding more as they rise (Fig. 12). Examples can be found at the front of the Apennine prism, in the Carpathians and the Banda Arc. This strong subsidence appears to be caused by slab retreat, and prevails to such an extent that the accretionary prism may even find itself below the foreland (see again Fig. 8).

In foreland basins located in front of mountain belts above W-directed subductions, the section of the accretionary prism has an area which is, on average, less than that of the foreland basin itself, in other words a ratio of less than 1 (see again Fig. 10). Examples are the prism and corresponding trench of the Marianas, or the Apennine mountain belt and the Po-Adriatic foreland basin, where in some places more than 8 km of sediments have collected over 5 My. In this type of foreland basin, rates of subsidence are so high and the adjacent mountain belt so low (i.e. with limited erosion) that the foreland basin is underfilled (see again Fig. 12).

By contrast, mountain belts linked to E- or NE-directed subductions have two foreland basins: in front of and along the back-thrust belt of the orogen. Rates of subsidence are relatively low (<0.2 mm/y) and the thickness of sediments is about 3 km deposited over 20 My, as for example at the front of the northern Alps. The anticlines and the accretionary prism are always higher than the foreland (see again Fig. 12). The ratio of the area in section of the mountain belt to the total area of the two foreland basins is paradoxically always greater than 1: although the mountain belt is extremely high, the two foreland basins are smaller in size (see again Fig. 10). In this type of mountain belt (Rocky Mountains, Alps, Himalayas), erosion is so high and the space in the two basins so limited that the foreland basins are overfilled, and rapidly pass from the flysch facies to the molasse facies until they fill up and sediments from the orogens by-pass the basins and are transported to remote deltas. Examples are the large Ganges and Brahmaputra deltas where the material eroded from the Himalayan chain accumulates, no longer finding space to deposit in the foreland basin.

If we accept the westwards drift of the lithosphere, W-directed subductions are generated primarily by the sinking caused by the mantle, which has a relative eastward motion. In this case, the foreland basin is located on the subduction hinge and its subsidence coincides with the retreat of the slab. In E- or NE-directed subductions, which follow the direction of mantle flow, the latter supports the lithosphere from below, thus in part counterbalancing the load of the mountain belt, which in this type of geodynamic environment is largely responsible for the sinking of the foreland basin. When the subsidence of the foreland basin is greater than the uplift of the prism,
the total uplift of the anticlines is negative; otherwise it is always positive (see again Fig. 12).

These asymmetries are in line with the supposed impact of the westward drift of the lithosphere relative to the mantle; the movement of the latter towards the east inclines the subductions westward, making them retreat, and generating the arcuate shapes typical of the Lesser Antilles, Sandwich, the Apennines, Carpathians, Marianas, Japan, Banda, etc. In these subductions the lithosphere is mainly dispersed within the mantle (see again Fig. 9). In E- or NE-directed subductions, which dip in the direction of movement of the underlying mantle, the lithosphere is held up by the flow and thickens.

There are orogens which do not follow the flow shown in Fig. 4, such as for example the northern part of South American and the Pyrenees. These orogens are linked to subductions generated by the sub-rotation of the South American and Iberian plates, and have characteristics resembling those of orogens associated with E-directed subductions; in other words, twosidedness, absence of back-arc extension, high structural and morphological elevation, and foreland basins with low rates of subsidence.

**Sedimentary basins**

The sedimentary basins in which organic substances that may generate hydrocarbons accumulate are a direct consequence of plate tectonics. These form both within plates and at their margins as a result of three main processes of subsidence: thinning of the lithosphere, in other words extensional or transtensional tectonics; thermal cooling of oceanic and continental lithosphere at passive margins; folding of the lithosphere at subduction zone hinges due to slab retreat, or to sinking generated by the load of a mountain belt or a delta on a continental margin (Fig. 13).

Sedimentary basins form where the crust subsides or where there is a pre-existing empty basin which can be filled with sediments. The weight of the sediments usually generates a further load which causes the lithosphere to sink. The compacting of sediments caused by lithostatic stress (equal to \( r g z \), where \( r \) is the density of the column of rock, \( g \) the gravitational acceleration, and \( z \) the thickness of the column of rock) leads to a decrease in the porosity of the rock, and an expulsion of fluid from its pores and thus further subsidence. Lithostatic stress also leads to a decrease in volume caused by pressure-solution, and thus further subsidence.

Subsidence in an extensional zone is a function of the rate of extension and the inclination of the extensional faults. Given identical rates of extension, more steeply angled faults allow for more rapid subsidence.

Intraplate extensional basins cause the crust and the lithosphere to weaken; as a result, if there is a modification in the stress field these are the first areas to undergo tectonic inversion. A classic example is the Atlas mountain chain, generated by a sinistral transtension and en échelon (stepped) extension during the Mesozoic, and later inverted to form a dextral transpression.

The thermal subsidence of the oceanic crust discussed above also takes place at passive continental margins if the adjacent oceanic crust is no older than 60 My. Foreland basins are typical basins linked to the folding or sinking of the lithosphere, and form as a result either of the load of a mountain belt and its sediments, or the retreat of the subduction. The slope of the basement beneath the foreland basin, towards the interior of the mountain belt, is known as the foreland regional monocline; it is less steeply angled (2-5°) in mountain belts where load is responsible for subsidence, and more steeply angled (4-10°) in foreland basins where subduction hinges of west-directed subductions are retreating ‘eastwards’ (see again Fig. 10).

In line with subsidence values for other major tectonic environments, foreland basins linked to W-directed subductions are those with the highest values.

There are areas of the Earth where several geodynamic factors governing the evolution of a basin may simultaneously coexist. For example, in the Sicilian Channel an active extension with extensional faults oriented NW-SE is separating Sicily from Africa; simultaneously the overthrusts of the Apennine-Maghrebid chain, oriented roughly E-W, are advancing towards the southeast, cutting across the normal faults;
these in turn cut across the overthrusts. The plain of north-eastern Italy is the foreland of the Alpine retrobelt, of the front of the Dinaric belt and of the Apennine chain; we thus have the combined effect of three different mountain belts generating subsidence in the same area with different mechanisms, velocities and in different directions. The San Andreas Fault is another example of a sinistral transpression superimposed on a faster dextral transtension oriented WNW-ESE.

**Plate dynamics**

Despite the enormous progress made in the Earth sciences, we still have no complete theory regarding the mechanisms which move plates, able to reconcile surface kinematics with supposed movements inside the planet. The forces acting on the lithosphere are of different types: the pull exerted by convective motions in the underlying mantle; ridge push, in other words the weight of mid-ocean ridges; slab pull, in other words the weight of subducting slabs; forces external to the planet such as those responsible for the tides (Bostrom, 2000). Plate movements are so slow that the corresponding forces of inertia are negligible.

**Mantle convection**

The upwelling of the mantle in rift zones and the sinking of lithosphere in subduction zones are in themselves evidence of the convection taking place in the mantle. On the geologic time-scale, the Earth’s mantle, though apparently solid, can be considered an extremely viscous fluid (with a viscosity above 10^{22} Pa s). A fluid heated from below and cooled from above may transfer heat through its thickness in two ways: conduction or convection. The mantle has an internal temperature gradient of less than 1 °C/km. The Rayleigh number (Ra) measures the ability of a fluid to transmit heat by convection. The lithosphere transmits heat both by conduction and by convective motions in the fluids which cross it.

The Rayleigh number for a layer of thickness \( h \) with constant temperatures \( T_0 \) and \( T_1 \) above and below, is given by:

\[
Ra = \frac{\rho c_p \alpha (T_1 - T_0) h^3}{\mu \kappa}
\]

where \( \rho \) is the density, \( g \) the gravitational acceleration, \( c_p \) the specific heat, \( \alpha \) the thermal expansion coefficient, \( \mu \) the viscosity and \( \kappa \) the thermal diffusivity (given by the ratio \( k/\rho c_p \), where \( k \) is the thermal conductivity). Quantities in the numerator favour convection, whereas quantities in the denominator indicate diffusivity and thus thermal conductivity, as well as the viscosity which slows convection. Thus in the presence of a high Rayleigh number convection is prevalent, whereas a low value indicates a predominance of conduction. The so-called critical Rayleigh number marks the transition between these two regimes. It is thought that about 90% of the mantle’s heat derives from radioactive decay occurring within it, whilst only 10% derives from the underlying core. The value of the Rayleigh number required to render a spherical mantle convective is equal to about 3 \( \times 10^6 \) but in fact, if we assume the values estimated by PREM (Preliminary Reference Earth Model; Anderson, 1989), the value of Ra calculated for the mantle is equal to about 9 \( \times 10^6 \). This means that convective motions must be occurring in the mantle; however, we do not know either their kinematics (pattern of the lines of flow and velocity), or how these internal movements can be reconciled with plate kinematics, which is much simpler than that of the convection cells which can be deduced from models.

The part of the mantle where convective phenomena could be expected to occur is the upper mantle. Here the Rayleigh number is higher because viscosity is lower, thermal conduction is lower because this zone contains less iron than the lower mantle, and the thermal gradient is higher than in the lower mantle. In the latter, the temperature increases by less than a degree per km, whereas in the upper mantle it may increase by several degrees per km.

There are two large areas where we can assume an upwelling of the lower mantle, identified by seismic tomography as volumes characterized by the lower propagation velocity of seismic waves: one in the central Pacific, and one in central-southern Africa (Romanowicz and Gung, 2002). Models linked to convection often conflict with evidence gathered at the surface: for example, the composition of the mantle is assumed to be homogeneous, although it is well-known that the whole Earth is intensely stratified. Were the mantle homogeneous, and its motions guided only by thermal gradients, we would expect portions of lithospheric mantle to become detached and sink into the underlying mantle. However, this phenomenon is currently unknown; were it to occur, it would generate an uplift of the surrounding residual lithosphere.

In convection models, upwellings of mantle are associated with lateral descending currents, but the Atlantic, East African and Indian rifts developed without any intermediate subduction. There are also cases of neighbouring pairs of subductions without rifting in between. In convection models, rising and sinking currents are stationary, whereas in nature all plate margins, rifts, subductions and transform zones migrate. Convection cells in models have polygonal shapes, whereas real plate margins, such as the Mid-Atlantic Ridge, are linear.
Given the obvious incompatibility between convection and surface kinematics, convection in the mantle cannot be considered the ‘conveyor belt’ which moves the plates (mantle drag). Furthermore, the lithosphere is decoupled from the mantle, as evidenced for example by the Hawaii hotspot, whose source within the mantle is moving ESE relative to the overlying lithosphere. The Mid-Atlantic and Indian ridges have moved away from Africa during their development, and are therefore moving relative to one another. This means that an active upwelling of stable mantle beneath the two mid-ocean ridges is not compatible with plate kinematics, and that rifts are passive structures, decoupled from and moving relative to the mantle. If mid-ocean rifts are moving laterally, this may explain why these are always sourced from still-productive mantle; were they statically positioned on the mantle, their sources would gradually become depleted. Seismic tomography has confirmed the presence of low propagation velocities for seismic waves only to a depth of up to 100-200 km beneath mid-ocean ridges. This probably indicates partial melting, whilst the underlying mantle often has relatively higher seismic velocities, suggesting the presence of cold mantle, and thus the absence of a deep source for mid-ocean ridges.

**Ridge push.** The rise of a mid-ocean ridge causes an increase in potential gravitational energy, in other words ‘ridge push’. Since this push is not linked to the upwelling of magma along the ridge, only the increased weight determined by the greater height of the ridge is taken into account. A simple expression of ridge push \( F_{rp} \) per unit of length (of the ridge) is:

\[
F_{rp} = g \rho_w \int h \, dx - \rho_s \int w \, dx
\]

where \( g \) is the gravitational acceleration, \( \rho_w \) the density of the water, \( h \) the elevation of the ridge above the seafloor, \( x \) the horizontal width of the flanks of the ridge, \( w \) the depth of the seafloor relative to the ridge and \( \rho_s \) the density of the water. The value obtained for ridge push, also taking into account the effect of the cooling of the lithosphere and the weight of water, is equal to about \( 3.9 \times 10^{13} \text{Nm}^{-1} \) (Turcotte and Schubert, 2002).

**Slab pull.** Slab pull (downwards pull of the subduction) is the mechanical action which can be ascribed to the lower temperature of the subducting slab relative to the warmer surrounding mantle. As they subduct, oceanic basalts may be transformed into eclogites, high density rocks, due to the extremely high pressure; the subducting slab thus has a negative density gradient relative to the surrounding upper mantle. The simplest expression of slab pull \( F_{sp} \) per unit of length, assuming that lithosphere and mantle are of identical composition, and that there is only one thermal boundary is:

\[
F_{sp} = g \rho_l (\theta_l - \theta_m) \, d
\]

where \( g \) is the gravitational acceleration, \( \rho_l \) the density of the lithosphere, and \( \theta_m \) the density of the mantle. Assuming values of \( 10 \text{ ms}^{-2} \) for \( g \), 100 km for \( d \), 660 km for the depth of the slab \( z \), and 3,300 kg/m³ and 3,220 kg/m³ for the density of the lithosphere and the mantle respectively, we obtain a slab pull of about \( 5.2 \times 10^{13} \text{Nm}^{-1} \). However, the subducting lithosphere is often thinner than this, and more importantly the upper mantle has densities far greater than 3,220 kg/m³, partly because it probably has chemical and mineralogical stratifications, with a gradual increase in density from top to bottom. Turcotte and Schubert (2002) calculate a slab pull of about \( 3.3 \times 10^{13} \text{Nm}^{-1} \). PREM, for example, suggests a density of 3,970 kg/m³ for the mantle at a depth of 600 km.

The olivine in the mantle, in addition to the olivine/spinel transformation at a depth of about 400 km which increases its density, may transform from magnesium-rich olivine (forsterite) to iron-rich olivine (fayalite), thus causing a further increase in density and decrease in volume. The value of slab pull is, therefore, probably an overestimate. Another argument against slab pull is the fact that the focal mechanisms of earthquakes mainly indicate that subducting slabs are subjected to internal compression parallel to the slab; were slab pull taking place, the slab would be undergoing traction. Nevertheless, slab pull is currently considered the greatest force affecting the lithosphere, being of a greater order of magnitude than ridge push.

There is geological and tomographic evidence that continental lithosphere is also subducted. Otto Ampferer, an early 20th century Austrian geologist, suggested the existence of a continental subduction beneath the Alps, based on the stacking of the Alpine nappes. The accretionary prisms where passive continental margin sediments can be seen piled up, indicate that the lithosphere on which they once rested has disappeared due to subduction. We have no data on the depth to which continental lithosphere, despite its lower density, can subduct with the help of transformations which increase its weight. In the central-northern Apennines there is a continental subduction reaching a depth of at least 100-150 km. This indicates that it is not only the weight of cold oceanic lithosphere which moves plates by slab pull, since in this case continental lithosphere would be unable to subduct. An eastward flow of the mantle, on the other hand, would contribute to the retreat and subduction of the lithosphere.

Another force which may act on the lithosphere is trench suction. As it retreats, a subduction zone sucks the overlying plate towards the hinge zone of the slab, moving it towards the subduction itself and/or causing
its margin to thin. This mechanism also becomes secondary if slab pull is not the driving force behind plate dynamics.

**Effects of the Earth's rotation**

Plate tectonics has hitherto been attributed exclusively to endogenous phenomena: the cooling of the planet and thermal convection. However, it has been shown that the movements of the mantle and plates disturb the Earth's rotation, provoking oscillations in the rotational axis. The westward drift of the lithosphere relative to the mantle, and all its tectonic implications, in turn indicate that the Earth's rotation contributes to plate dynamics, both in terms of the direction of movement, and above all in terms of energy.

The gravitational attraction of the Moon and Sun generates both fluid and solid tides on Earth, creating a permanent westward drift of the lithosphere, and simultaneously acting as a brake on the velocity of the Earth's rotation. An increase in the duration of a day of about 1.79 ms/century has been measured. For example, thanks to studies of stromatolites and tidal deposits, it has been established that 700 million years ago a year consisted of about 400-430 days; in other words the length of a day was about 21-20 hours, due to the greater velocity of the Earth's rotation (Denis et al., 2002). This greater velocity of rotation also caused a greater flattening of the Earth's poles; between 2.5 billion years ago and today the flattening of the poles relative to the equator has decreased from 0.005 to 0.003.

The centre of gravity of the Earth-Moon system lies within the mantle; Bostrom (2000) has shown that if we consider this system a double planet, gravity at the Earth's surface is slightly angled (0.38°) as an effect of the Moon's gravity. This inclination also causes asymmetry in mantle convection.

The Earth's solid inner core did not exist until 2 billion years ago, and according to some authors only began to solidify in the last 500 My. The lower mantle also shows an accumulation of denser material in its inner parts; this material can no longer rise due to the immense pressures at a depth of about 2,800-2,900 km. This means that the densest elements are slowly accumulating in the lower parts of both the core and the mantle, provoking a decrease in the Earth's moment of inertia and a corresponding increase in the velocity of rotation; however, this is not sufficient to compensate for the slowing due to tides. The combination of tidal effects and the sinking of the denser parts of the Earth towards the core represent a pair of forces acting on the asthenosphere, the layer of least resistance, which might explain the mean westward movement of the lithosphere. In this model, plate tectonics is a combination of rotational effects and convective motions in the mantle (Scoppola et al., 2003).

Were it to be confirmed that OIB (Ocean Island Basalt) magmas at hotspots are sourced from the asthenosphere, as are MORBs (Middle Oceanic Ridge Basalts) at mid-ocean ridges, and IABs (Island Arc Basalts) in subduction zones, sourced from depths of 100-150 km, there would be strong evidence that most of the Earth's magma comes from the superficial layer of the upper mantle. As such, given the lack of concrete petrological data on the composition of the lower mantle, the latter may be richer in iron, and thus denser, than hitherto thought. In this case, the effect of slab pull would be even less than estimated above, and no longer able to drive plate dynamics. A combination of astronomical effects and convection is therefore better able to explain the Earth's geodynamics.

**References**


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### 1.4.2 Sedimentary basins

Sedimentary basins are structural depressions of the Earth's crust that are filled with sediments more than 1 km thick. They are, in general, underlain by a peneplaned highly deformed igneous and/or metamorphic basement, which is of little interest to petroleum geologists, and is therefore called **economic basement**.

Many basin classifications have been proposed and most related papers can be found in Foster and Beaumont (1987) or are summarized in Busby and Ingersoll (1998). All such classifications are simplified mental constructs that merely provide an overview of very complex and variable geology. It is true that only a few basin types (e.g. foredeeps) contain the lion's share of the ultimate hydrocarbon reserves of our Earth. However, basin classifications do not allow the realistic prediction of analogue-based future hydrocarbon reserves. For each hydrocarbon-rich basin of a given type there will always be a similar, but hydrocarbon-poor, basin type elsewhere.

Plate tectonic concepts (see Section 1.4.1) offer a valuable background that allows the classification of basins (Bally and Snelson, 1980; Busby and Ingersoll, 1998). A greatly simplified plate tectonic map of the world (**Fig. 1**) shows that, since the Early Jurassic, ocean-spreading formed the presently preserved, relatively rigid, oceanic crust that underlies two thirds of the Earth's surface. Orogenic belts are here called megasutures. These record complex processes occurring at overall compressional diffuse plate

![Simplified plate tectonic map of the world](image-url)

**Fig. 1.** Simplified plate tectonic map of the world. Note that strike-slip boundaries like the San Andreas of California, or the Alpine fault of New Zealand are difficult to show at the scale of the map.
boundaries that sutured more stable continental lithospheric elements. Bally and Snelson (1980) differentiate four boundaries as follows:

- **B-subduction boundary** associated with subduction oceanic lithosphere.
- **A-subduction boundary** associated with more limited subduction of continental lithosphere.
- **Transform or strike-slip dominated boundaries**.
- **A boundary in Central Asia** characterized by a diffuse envelope around igneous Mesozoic and Cenozoic intrusives; this boundary is also characterized by prominent intra-palate deformation.

Cenozoic-Mesozoic megasutures display all of these boundaries. Paleozoic megasutures represent a series of continental collisions that concluded with the formation of the Pangean supercontinent and therefore are dominated by A-subduction boundaries. Several complex pre-Cambrian megasutures are responsible for the assembly of thick Precambrian lithospheric blocks. Fig. 1 also serves as a first approximation of the age of the economic basement of all Phanerozoic sedimentary basins.

Artemieva and Mooney (2002) recognize a lithospheric thickness distribution centering about 350 and 220 km for Archean lithosphere, about 200 km for Early Proterozoic, around 140 km for Middle and late Proterozoic and about 100 km for Paleozoic lithosphere. Thicker and older continental lithosphere is more likely to be preserved and thus provides relatively more stable platforms for sedimentary basins. The Paleozoic and Mesozoic continental lithosphere is the result of more recent orogenic processes and is relatively less stable, thus allowing the formation of younger sedimentary basins.

As defined here, sedimentary basins have undergone only limited tectonic deformation and are structurally relatively intact. This definition contrasts with the geosynclines of earlier authors. These were hypothetical, large, often elongate basins characterized by substantial subsidence. Geosynclines were simplified, conjectural geological reconstructions of folded belts that were based on good geologic fieldwork but inadequate geophysical background. Today the old geosynclinal nomenclature is obsolete; however, in the following, some of the early nomenclature will be briefly mentioned only to indicate a crude and approximate equivalency. This permits a better appreciation of some earlier, often very detailed, observations in the context of more modern basin studies. Sedimentary sequences that are strongly deformed occur within folded (orogenic) belts such as accretionary wedges associated with the subduction of oceanic
lithosphere, and foreland folded belts (see below) associated with limited subduction of continental lithosphere.

The formation of sedimentary basins typically involves a variety of processes and stages. Hence, assigning a basin to a given class is often arbitrary. However, when studying hydrocarbon systems, it often makes sense to use the younger stages in the evolution of a basin as the key classifying criterion, as hydrocarbon systems in any given basin tend to be more developed during the later stages of its evolution. Detailed modern sequence stratigraphic analysis is a key tool for the economic evaluation of basins. Some authors define sequence-stratigraphic boundaries as a series of unconformities with their conformable continuation, while others prefer to focus on transgressive/regressive cycles. However, the tectono-stratigraphic megasequences of this review are more comprehensive subdivisions that sum up the stratigraphic response to the structural evolution of a given basin (Sharland et al., 2001). Commonly, the unconformable boundaries of such megasequences coincide with the change from one global plate tectonic regime to another and thus they may also correspond to unconformities of second order cycles of the sequence stratigraphers.

### Sedimentary basin types

#### Dominantly extensional basins on rigid lithosphere

**Rift systems.** These elongate fault-bounded basins are mainly characterized by half-graben systems segmented by various types of transfer zones. A single half-graben may dominate and/or be part of a trilete system, i.e. a starlike arrangement often called a “Triple Junction” (e.g. the northern termination of the Rhine-graben). The underlying basement is always involved in the formation of rifts, which are commonly, but not always, associated with stretched, i.e. attenuated continental lithosphere (see Section 1.4.1).

Sengör and Natal’ in’s (2001) elaborate rift inventory and classification are here greatly simplified to differentiate hot spot-related intraplate rifts from transtensional basins related to strike-slip plate boundaries and other rifts associated with diffuse compressional orogenic plate boundaries and their foreland. Thus rifts occur in a wide variety of plate tectonic settings and basin types. Active rifts are characterized by high seismicity, high heat flow and volcanism. Many Tertiary rifts maintain their individuality; however, older rifts from Precambrian to the Mesozoic and a few Tertiary rifts have been affected by post-rifting events. They are often buried

![Fig. 3. Basins on rigid lithosphere. This map shows passive margins as well as cratonic basins. Note that rift systems have been omitted because they would not show well on a world map of this scale.](image-url)
under a substantial thickness of sediments deposited in the course of the evolution of a variety of basins types.

The internal structure of the crystalline basement of rifted regions and the structures affecting the pre-rift megasequence, are of relevance, as older structures may be re-activated during or after the rifting process. Pre-rift megasequences are deposited discordantly on the basement. In turn they are overlain by one or more syn-rift megasequences and several post-rift megasequences. Each of these may have its own hydrocarbon reservoirs and source beds forming hydrocarbon systems that are limited to a single megasequence or else hydrocarbon systems that are shared with overlying and underlying megasequences.

The syn-rift basin fill includes either continental or marine strata as well as volcanic ones.

Continental syn-rift sediments are commonly fluvial clastics, but are also lacustrine lake beds (e.g. the Reconcavo and Tucano basins of Brazil) that are prolific source beds for accumulations in adjacent and overlying reservoirs. Marine rifts systems with their marine source beds may be flanked by reefs located on the structural highs occurring either in the hanging-wall or footwall of normal faults, such as uplifts generated by domino structures, rotational faults or by isostatic uplifts of the footwall. On occasion, syn-rift volcanics include significant reservoirs. Finally syn-rift evaporitic deposits are associated with trap-forming diapiric structures that affect both the syn-rift and the post-rift formations.

Syn-rift deposits often display syn-tectonic growth, i.e. updip divergence and thickening of the strata, towards the fault plane in the hanging-wall, and the reduction, or absence, of the same sediments in the footwall (Fig. 2 A). However, with rapid extension, sub-horizontal strata onlap the downthrown hanging-wall tilted block as well as the relative fault scarp of the footwall (Fig. 2 B). Strictly speaking, such infill could be lumped with the post-rift stratigraphy but the sedimentary records mitigates in favour of inclusion in the syn-rift megasequence.

The post-rift evolution varies considerably, ranging from late syn-rift to post-rift uplift of the rift shoulders, to uplift and partial erosion of the entire rift system but, more importantly, also leading to a large variety of basin types described later on. Basins immediately overlying rifts are also called sag basins, while others informally call the combined rift and sag basins steer’s head type basins. Older rifts initiating the formation of more complex basins will be mentioned later. More elaborate discussion of specific rifts can be found in Landon (1994).

**Passive margins.** Passive margins are also called Divergent or Atlantic-type margins. They are typically conjugated and/or directly associated with spreading oceans. They overlie a continental landward basement and an oceanic basement on the seaward side as they straddle a geologically often ill-defined ocean continent boundary. Fig. 3 shows the distribution of passive margins and cratonic basins, while Fig. 4 illustrates the development of passive margins. They generally overlie a coast-parallel rift system, but in some important cases they also overlie near-perpendicular or oblique failed arms of triple junctions (e.g. Benue trough of Nigeria). All passive margins are associated with the post-Permian break-up of Pangea.

Beginning with the Proterozoic, former passive margins of all ages were involved in the deformation...
of folded (orogenic) belts and particularly foreland folded belts and their associated foredeeps. It is fair to compare and to roughly equate the old term miogeosyncline (or miocline) of earlier authors with today’s passive margins, always keeping in mind that the older terms were conceptual and based on inadequate reconstructions of folded belts.

In recent years passive margins have been subdivided into:

• Rifted margins, underlain by a highly extended crust and associated rift systems. The syn-rift fill may be continental and/or marine.

• Volcanic margins, underlain by very thick wedges of volcanics that are characterized by Seaward Dipping Reflectors (SDR) on seismic profiles (Fig. 5). Note that occasionally explorationists mistook SDR’s for syn-rift sediments, leading to the drilling of some dry holes.

• Transform margins, divided into: transtensional transform margins, characterized by transtensional half-grabens (e.g. the south coast of southern Africa); and transpressional transform margins, characterized by transpressional folds (e.g. offshore Ghana).

The development of all passive margin types can be summarized by their shared megasequence development, modified only for the specific differences of each margin type. Thus, the syn-rift megasequence on rifted margins is replaced and/or covered by a thick seaward-dipping volcanic wedge on volcanic margins (see again Fig. 5). Numerical models suggest that passive margin subsidence is driven by cooling of the rifted /volcanic margin and the oceanic crust that moves away from the hotter mid-ocean ridge combined with the effects of sediment loading. Transtensional half-graben systems commonly form during the early phases of a transform margin evolution. However, transpression occurs mostly later in the evolution of transform margins and is characterized by reverse faults and minor associated flexural basin sequences. Most margins exhibit a more or less obvious unconformity that separates the underlying syn-rift and/or volcanic megasequence from the overlying post-rift or post-volcanic megasequences. This is the break-up unconformity of some authors, which is considered to mark the onset of oceanic spreading and the associated passive thermal subsidence of the continent/ocean boundary (see Section 1.4.1). Its age, to a first approximation, is the same as the oldest adjacent oceanic basement. On volcanic margins it is sometimes difficult to differentiate volcanic basement from regular ocean floor.

The presence or absence of evaporites and particularly of salt is important for the economic evaluation of passive margins. Salt may form part of the syn-rift fill, but very commonly salt is deposited in larger post-rift sag basins. The original distribution of salt determines the scope of salt tectonics. The larger the salt basin, the more complex the salt tectonics and the larger the opportunity for salt-related hydrocarbon traps.

Based on the dominant post-rift sedimentary regimes, passive margins are classified as mixed carbonates/clastic or dominantly clastic margins. Production from structurally relatively undisturbed passive carbonate margins is rather limited, while dominantly clastic margins are major hydrocarbon producers. Note that outlying carbonate platforms such as the Bahamas and the Maldives are mostly underlain by oceanic crust and therefore are not included in conventional passive margins.

Fig. 5. Emplacement of two opposite diverging volcanics (heavy lines) just before the breakup of an ocean. Contrast with rifted margin on the far side of the transform fault.

Megadeltas and their corresponding deep-sea fans form the end member of clastic margins that contain some of the world’s most prolific hydrocarbon provinces including the Gulf of Mexico, the Niger Delta and the Nile Delta. Other megadeltas such as the Amazon, the Zambesi and the Bengal remain
underexplored. The attractiveness of megadeltas is underscored by the common presence of marine source beds, and by widespread syn-sedimentary growth-fault systems that are due to gravitational spreading associated with a shifting load of deltaic depocenters. Deltaic and various shallow and deepwater sands offer good reservoirs separated by adequate seals. The importance of salt and shale tectonics associated with megadeltas is discussed in a series of papers by Edwards and Santogrossi (1990), Jackson et al. (1995), Cameron et al., (1999), Mohriak and Talwani (2000), and Arthur et al. (2003).

Ocean basins. Strictly speaking, the great ocean basins of the world ought to be included in any discussion of basins. Their origin has been briefly summarized in the introduction. Outward from the passive margins, oceanic basins are of little economic interest to hydrocarbon explorationists. The oceanic crust is typically overlain by a relatively thin cover of mudstones that may well include some, mostly immature, source beds and fewer significant reservoirs with increasing distance away from continental margins. Cratonic basins. Cratonic or intracratonic basins (the sinueclises of Russian authors) form on continental lithosphere or cratons (see again Fig. 3). They are deceptively simple, but views regarding their origin vary widely, reflecting the differing backgrounds of various authors. The term craton implies stability of large continental platforms. Particularly stable Precambrian provinces are thought to be associated with stable, deep and buoyant cratonic roots, yet cratonic basins and their adjacent arches (the anticlises of Russian authors) still record a significant degree of instability. Some of the factors influencing such instability are thought to include asthenospheric upwelling, intraplate extension/rifting associated with lithospheric attenuation and intraplate compression. Paleozoic and Mesozoic lithosphere is thinner and weaker than Precambrian lithosphere (see above), which tends to distinguish cratonic basins that overlie Precambrian lithosphere from those that overlie Paleozoic and younger lithosphere. Precambrian shields are large outcrops of the cratonic basement exposing highly deformed, often metamorphic and igneous, rocks (see again Fig. 1). In their subsurface continuation they are separated from all the overlying sediments by widespread regional unconformities. On occasion some Proterozoic basin remnants underlie these unconformities. The overlying Paleozoic and Mesozoic megasequences are of exploration interest as they often contain source beds, reservoirs and seals. The concept of worldwide correlatable cratonic megasequences was originally developed by Sloss (1963, 1988), who gave them the names of Indian tribes. While the reality of these worldwide correlatable megasequences is not questioned, there remains an ongoing debate as to whether they are related to worldwide tectonic phases and/or local changes in structural regimes or else to eustatic sea-level changes or perhaps more plausibly to a combination of all of these factors. The cratonic megasequences in general correspond to second order cycles recognized by sequence stratigraphers and it is likely that their boundaries correspond to major worldwide plate re-organizations.

Paleozoic cratonic basins should be viewed in the context of a Permo-Triassic Pangea assembly, which shows that, with the exception of the margin that extends from North Africa to northwestern Australia, most of the Pangean supercontinent was surrounded by Paleozoic active margins thus weakening the margins of the Precambrian cratons adjacent to these folded belts (Bally and Snelson, 1980). Particularly for North America and South America it may be plausible to infer that, in addition to Proterozoic rifting processes, intraplate compression may have made a significant contribution to the formation of Paleozoic cratonic basins and intervening arches. Compare this with Mesozoic Africa, which is surrounded by spreading ocean ridges associated with the Mesozoic break-up of Pangea leading to widespread cratonic rift systems (Arthur et al., 2003). Many cratonic basins of Western, Central Europe (Ziegler, 1990; Baldschuhn et al., 2001) and West Siberia overlie a relatively thin and unstable Paleozoic lithosphere, facilitating both extensional and compressional reactivations of the underlying basement structures. Finally it is noteworthy that a number of cratonic basins are characterized by widespread flood-basalts perhaps related to the presence of underlying hot spots (e.g. the Siberian Platform and the West Siberian basin). Depending on their timing of emplacement and distribution, these may influence the thermal evolution of the basin. The apparent simplicity of cratonic basins hides a great deal of complexity due to the interplay of local tectonics with distant tectonics and their impact on the stratigraphic evolution of these basins. Thus it is not sensible to develop a single idealized simple prototype for cratonic basins or else for their counterparts, the cratonic arches. The differentiation of megasequences and/or Sloss-type cratonic sequences (see above) is useful to describe hydrocarbon systems of cratonic basins. However, it should be noted that some cratonic basins share the same source beds with neighbouring foreland basins (e.g. some Paleozoic source beds of North America), while other cratonic basins develop their very own source beds (e.g. the Neocomian Bazhenov source bed of West Siberia).
Three particularly well documented and explored cratonic basins are the Illinois basin (Leighton et al., 1990) the Paris basin (Mégnien, 1980) and the northwestern Germany basin (Baldschuhn et al., 2001).

**Perisutural basins**

Deep-sea trenches. Deep-sea trenches (Fig. 6) are elongate depressions immediately adjacent to accretionary wedges associated with the subduction of oceanic lithosphere. They are partially filled with deep-sea mudstones and turbidites ready to be incorporated in the active adjacent accretionary wedge. Deep sea trenches are of no interest to hydrocarbon explorers, however, they have been considered as long-term repositories for radioactive waste.

Foredeeps or foreland basins located on rigid lithosphere. The transition from oceanic subduction to continental subduction occurs in stages. The thinned continental crust of passive margins is first subducted, heralding the inception of continental collision. This is followed by a progressive evolution from deep-sea trench to a sediment-filled remnant ocean basin, then a foreland basin or foredeep, which in turn may, at least in part, get incorporated into the adjacent foreland folded belt or else be broken up in smaller basins by basement involved uplifts of the foreland craton.

Foreland basins are roughly equivalent to the exogeosynclines of earlier authors, whereby most of these authors would limit the term exclusively to the clastic wedge that overlies platform sequences of the former passive margin. However, in this article, we lump the underlying platform megasequences and the overlying foreland clastic wedge megasequences as one unit. This is preferable because hydrocarbon systems in foreland basin involve the complete sedimentary section of the basin. The distribution of foreland basins is shown in Fig. 6.

Remnant ocean basins. These are transitional basins mostly underlain by oceanic crust adjacent to folded belts. A good example is the onshore and offshore Ganges Delta, which is still underexplored and may have a very substantial hydrocarbon potential. The Black Sea may be an additional example.

Foredeeps or foreland basins. These asymmetrical flexural basins are due in part to loading by the adjacent folded belts and/or to slab-pull associated with partially subducted foreland platforms (Fig. 7). An idealized drawing of a foreland basin or foredeep (Fig. 8) illustrates some of the significant megasequences that characterize these basins. The basement may be the peneplaned remnant of a Precambrian or Paleozoic folded belt now acting as a rigid, but flexed, craton. It typically consists of
highly deformed, often metamorphic sediments and intrusive rocks. The basement may be rifted due to earlier rifting that initiated a former passive margin, which now forms one or more platform megasequences.

The platform megasequences of most foredeeps are commonly the preserved remnants of proximal passive margin shelves and include both carbonates and clastics. They often form stratigraphic and combined stratigraphic/structural traps. Platform isopachs and facies trends also commonly run obliquely to the strike of the adjacent folded belts, which permits to observe stratigraphic variations of the platform on outcrops in the adjacent mountain range. Platform carbonates and associated reefs in foredeeps often contain prolific hydrocarbon accumulations. Due to differential entrapment, oil fields tend to occur updip and gas fields tend to be downdip in the foreland basin. Clastics within the platform sequences are generally derived from the adjacent cratons as displayed by progradation towards the mountains. Typically, platform megasequences and their bounding unconformities are coeval with their neighbouring cratonic megasequence boundaries.

The important basal foredeep unconformity forms first when deep-sea trench sediments onlap on the subducted and flexed outer continental margin. As continental subduction proceeds a flexural bulge forms farther updip. A minor migrating uplift associated with that bulge creates a dynamic unconformity. The unconformity truncates underlying platform strata in an updip direction and, given a good seal at the base of the overlying clastic wedge, it will form excellent palaeomorphological/subcrop traps.

The siliciclastics of the overlying clastic wedge were transported by river systems that primarily originated in the mountains. After reaching the foreland, the river system often gets reorganized into longitudinal river systems that redistribute the clastics along the axis of the foredeep (e.g. the Ganges River...
system of northern India). In the schematic diagram (see again Fig. 8), foreland sedimentation is shown to begin with a shaly deep-water sequence with turbidites that onlap on the basal foredeep unconformity. These are the Flysch deposits of earlier authors. As the foredeep evolves, shallow water sequences will take over including delta, prodelta sands and coarse clastics, which correspond to the Molasse deposits of earlier authors.

Foredeeps are mostly filled with lithic sands derived from the nearby rising mountains and hydrocarbon traps are often related to updip pinchouts of these sands. However, there also may be some stratigraphic traps with a craton-derived source for clean quartzose sands.

Foreland folded belts adjacent to the foredeep involve former foredeep megasequences and their underlying platform sequences and, occasionally, the basement (see again Fig. 7). Thus, foreland folded belts cannibalize their adjacent foredeep sequences. Alternatively, however, parts of the foredeep clastic sequence may also onlap on the adjacent folded belt (e.g. the Veracruz Basin of Mexico). In this context, De Celles and Giles (1996) offer a more differentiated view of a wider foreland basin system that includes wedgetop, foredeep and backbulge depocenters. There are different configurations for the foreland basin folded belt boundary. Wedgetop (also named piggy-back or satellite basins) form on top of an actively deforming folded belt and are connected with the adjacent foredeep. Because these smaller allochthonous basins form part of the foreland folded belt hydrocarbon system, they are not discussed here. On the craton-side of the peripheral bulge a widespread backbulge depocenter may also form. Depending on its location such depocenter may either be part of the foreland basin hydrocarbon systems or else of an adjacent cratonic basin hydrocarbon system.

Foredeeps do include the most prolific hydrocarbon accumulations of the world, including many of the Middle East basins. Rich source beds occur both in the underlying platform megasequences and the overlying foredeep sequences. While in the Middle East hydrocarbon trap domains are dominated by Mesozoic and Cenozoic carbonates, it should be pointed out that huge reserves of Middle East size are also contained in the Tar Sands and Heavy Oil traps of the distal foredeeps of Venezuela and Canada. Hydrocarbon systems in foredeeps may be limited to specific megasequences of the platform and the foredeep, but are often shared by both systems with hydrocarbons migrating from the underlying platform across the foredeep unconformity into the overlying clastics, as is the case for the above-mentioned Tar Sands.

The hydrocarbon richness of foredeeps is easily explained by the asymmetry and size of these basins, which, given good source beds, provide large perennial fetch areas from mature source beds. In addition to these conventional hydrocarbons, foreland basins and their equivalents in foreland folded belts contain most of the world’s coal reserves and their associated potential for coal-generated natural gas.

Well-documented overviews of foredeeps are available for the Middle East (Sharland et al., 2001), the European foreland basins (Masce et al., 1998) and the Western Canada Basin (McQueen and Leckie, 1992; Mossop and Shetsen, 1994).

Foredeeps or foreland basins disrupted by basement uplifts. Some foredeeps are disrupted by foreland basement uplifts associated with orogenic processes that impinge on the foreland. The platform sequence and the overlying clastics foredeep wedge will then only be preserved in the basins between the uplifts. However, an additional megasequence will be deposited in the residual basins and on their deformed mountain flanks. Most of these deposits are likely to be alluvial, fluvial and lacustrine, often providing a unique hydrocarbon-rich lacustrine source rock such as the Green River Shales of the US Rocky Mountains. This class of basins not only inherits the hydrocarbon endowment of its predecessor basins but now has an additional novel and often prolific hydrocarbon system. Good examples of these basins are the Green River and Uinta basins of the US Rocky Mountains and the Maracaibo basin of Venezuela.

Basins of central Asia. The boundary types for orogenic system were listed earlier showing that in central Asia this boundary is an ill-defined envelope around Mesozoic and Tertiary intrusives. The region is also characterized by compressional and transpressional uplifts such as the Tien Shan and the Kuen Lun Shan and associated flexural sedimentary basins. During the assembly of Pangea much of central Asia was an active margin where a large number of island arcs systems and some minor continental cratons were accreted to form the basement of a number of basins. Accretion continued farther south into the Mesozoic and culminated with the Tertiary collision of India with Eurasia and the rise of the high Tibetan Plateau. Mountains and sedimentary basins formed to the north of this plateau in response to continued compression and the long-distance impact of India on the Eurasian continent. The Mesozoic and Tertiary infill of these basins is entirely continental and includes lacustrine source beds. Clastic reservoirs are mainly derived from the rising adjacent mountains (Li Desheng, 1991).
Basins located within orogenic belts (epi-sutural basins)

Basins associated with oceanic subduction and island arcs. Fig. 9 illustrates the overall setting of some of these basins.

Forearc basins. These basins straddle the accretionary wedge associated with oceanic subduction and the adjacent volcanic island arc (Fig. 10). Most forearc basins have been tectonically somewhat compressed, providing significant anticlinal hydrocarbon traps. Superposed forearc basins contain several separate megasequences with the lower megasequence perhaps providing source beds and the upper one having the reservoirs. Commercial production from forearc basins is known from the Talara Basin (Peru) and the Cook Inlet (Alaska). A few forearc basins may be dominated by extensional structures, and others are affected by strike-slip faulting.

Circum-Pacific backarc basins. A large number of backarc basins are floored by oceanic crust that was either trapped or else formed by backarc spreading (see again Fig. 10). Rare arc-side hydrocarbon production from structures with clastic and volcanic reservoirs is known from Japan, while very significant continent-side production is known from structures offshore Indonesia, Vietnam, southern China and Sakhalin. Of greater exploration interest are the backarc regions of Indonesia, Malaysia and the Gulf of Thailand (Hall and Blundell, 1996). These are formed by the interplay of the opening of the South China Sea and the subduction of the Indian Ocean plate. The generally peneplaned basement of these basins consists of earlier island arc systems. There follows one or more, mostly continental, syn-rift megasequences that may include prolific lacustrine source beds.

Overlying these rifts are one or more marine and/or continental megasequences with good carbonate and clastic reservoirs and occasional source beds. Continued compression may lead to selective inversion of the earlier rift systems and the formation of smaller folded belts that verge towards the backarc basin. Consequently, some backarc basins often end up to be elongate asymmetric flexural basins and it may be difficult to separate subsidence due to cooling following an earlier rifting event from subsidence associated by flexing of the basin towards the arc itself, which may be due to mantle flow in the underlying backarc mantle wedge and/or else to volcanic loading (see Section 1.4.1).

Backarc basins associated with continental collision or post-orogenic collapse basins. These basins range from mostly oceanic to transitional and
continental, depending on the nature of their basement, and the degree of extension these basins have undergone. So far, only continental backarc basins offer some exploration interest. They are straddling orogenic belts and develop late during the orogenic evolution. Their basement consists of highly deformed sedimentary and metamorphic rocks of a buried folded belt. Continental and marine early syn-rift megasequences sometimes may provide source beds. They are overlain by post-rift megasequences that may have some good reservoirs. Late compression occasionally leads to the partial inversion of some of the rifts underlying these basins.

A good example for this class of basins is the complex Pannonian/Transylvanian basin of Hungary/Rumania, which is associated with the Alpine continental collision (Durand et al., 1999). Here the basement is formed mainly by a stack of Alpine basement-involved thrust sheets that extrude along complex strike-slip faults towards the north and the east. The extensional and transtensional structures in these basins are generally related to roll-back of the subducting slab (see Section 1.4.1). Hydrocarbons systems involve source beds, reservoirs and seals limited to the mostly Neogene successor basin fill.

A variation of similar basins (see again Fig. 2 A), but in a Cordilleran setting, is the Great Basin of the Western US which in simple terms can be viewed as a diffuse transtensional rifting system (Basin and Range), located between two regional megashears (the San Andreas fault of California and the Rocky Mountains trench of Canada). Hydrocarbon systems in the Great Basin are very complex because the hydrocarbon source beds form either in the extensional basin fill, or else in the poorly defined subcrop of the underlying foreland folded belt. Reservoirs would be clastics derived from the nearby uplifted ranges.

**Basins associated with major strike-slip faults.** Major strike slip-fault systems are often associated with diffuse transform plate boundaries such as the San Andreas (California), the Alpine (New Zealand) and the El Pilar (Venezuela) fault systems. The associated sedimentary basins are relatively small and often complex, ranging from transtensional pull-apart basins to transpressional basins that include inverted earlier transtensional structures.

The economic basement of these basins often consists of accretionary wedges as well as intruded and volcanic arc terranes. In some cases the basin is underlain by an earlier forearc basin megasequence, that is followed by one or more transtensional and/or transpressional megasequences. Structural deformation in many of these basins is still active today as evidenced by continued earthquake activity in the region and updip- converging of Plio-Pleistocene stratal wedges on the flanks of growing anticlines (Ingersoll and Ernst, 1987; Scholl et al., 1987; Biddle, 1991; Busby and Ingersoll, 1998).

Hydrocarbon systems of strike-slip associated basins may originate with source beds occurring either in earlier forearc basin sequences or else in later transtensional and transpressional basin. In areas of strong coastal upwelling a single source bed is shared by many basins and sub-basins as is the case for the Monterey formation of southern California. Reservoirs in these basins are dominantly clastics derived from the adjacent uplifted island arc terranes. The deformation of many of the structures in these basins promotes fracture enhancement of the reservoirs.
Concluding comments

A reasonable incentive for grouping basins into different classes would be to extract generalizations that are useful for the study of less explored basins, based on the exploration experience gained elsewhere in similar situations. Geologists know that there is some justification for using analogues. However, there are severe limitations to the use of exploration and production statistics from other apparently similar basins to bolster economic forecasts in less explored basins elsewhere. It is easy to demonstrate that the richness (i.e. the ultimate hydrocarbon reserve per area or else volume of sediment per area) of a given basin type ranges from very rich to very lean for individual basins within the same class. Some of the richest basins in the world such as the Los Angeles, the Ardmore, the Maracaibo and the Sumatra basins, present quite unique combinations of source, reservoir, seal and an overall basin evolution that cannot be satisfactorily replicated elsewhere (Bally and Snelson, 1980).

Even so, hydrocarbon explorers, rightly, continue to compare and analyse sedimentary basins to discover and/or understand hydrocarbon systems. In unexplored or underexplored basins, source beds and reservoirs are poorly known, but often analogues from similar basins elsewhere are useful to support an unexplored or underexplored basins, source beds and/or understand hydrocarbon systems. In order to compare and analyse sedimentary basins to discover and/or understand hydrocarbon systems, it is essential to identify and define new types of stratigraphic traps. On the other hand, following the discovery of a new play, it is of course sensible to use reservoir and hydrocarbon parameters from initial discoveries as analogues to reduce exploration risk.

In fact, worldwide, most of today total onshore and near-offshore ultimate hydrocarbon reserves have been found in close proximity (within a radius of about 200 km) of surface oil shows that were already known early in the last century (Höfer, 1909). Over the years increasingly more sophisticated technology improved the definition of primarily structural targets. Only in recent decades has there been an increased effort to understand the context of sedimentary basins in their totality (Mégnien, 1980; Mossop and Shetsen, 1994). For the hydrocarbon explorer basin analysis ultimately will always and primarily be based on the best possible seismic resolution, which will be particularly useful in definition of new types of stratigraphic traps.

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1.4.3 Source rocks: formation and distribution

We can define a petroleum system as a sedimentary basin or a portion of a sedimentary basin combining all the required elements and processes conducive for the formation of oil and gas accumulations (Magoon and Dow, 1994). The elements required include source rocks, migration routes, reservoir rocks, impervious seals, and traps. The processes involved include: the formation of hydrocarbons as a result of an appropriate burial-thermal history of the source rock; the efficient migration of the generated products along carrier beds and permeable migration conduits, such as porous sedimentary units, fractured rocks or faults; the focusing of hydrocarbon flows towards structural or stratigraphic features acting as traps, where they can accumulate; and the eventual preservation-alteration of hydrocarbons in reservoirs over geological time from accumulation to the present day.

As a part of this scenario, the source rock is indeed a crucial factor because it represents the geological object that feeds the oil and gas charge into the system. In this respect, the nature of the source rock is a major concern in the risk analysis of exploration ventures. Consequently, source rock geology and geochemistry have attracted a great deal of interest and research activity in order to provide exploration experts with the best data possible. This information is directed to minimize the uncertainties on occurrence, stratigraphic location, spatial distribution and thickness, as well as determining the petroleum potential of source rock(s) within a prospective area. The resulting knowledge forms an important basis for any legitimate attempt at risk analysis and the economic assessment of exploration plays.

The purpose of this article is to review the notion of source rock through a discussion on its formation, depositional environment, habitat and stratigraphy.

Formation of source rocks

A source rock is a sedimentary unit, hosting substantial amounts of fossilized organic matter, which is incorporated into the sediment at the time of deposition. Upon burial and associated thermal history, this sedimentary organic matter is subsequently thermally cracked to generate oil and gas (Hunt, 1995; Tissot and Welte, 1984). The sedimentary organic material is mainly derived from algal, bacterial and higher plant tissues that together make up the major part of our Planet’s biomass (Tyson, 1995). For a rock to be termed a source rock, its organic matter content should account for at least 1-2% of the weight of the rock (Bordenave, 1993). This kind of rock is far from common and requires some very specific formation conditions. These conditions, which have been actively debated over the last decades, include the local biomass productivity and the preservation of organic residues, strongly favoured under anoxic regimes, as well as the length and type of transport of organic material from the place of biological production to the sediment repository. One of the most controversial questions hinges on the relative importance attributed to primary productivity versus anoxia.

One school of thought has advocated that organic matter accumulation in the marine realm is linked to high organic productivity in the euphotic zone (e.g. upwelling areas), and that anoxia of the bottom water is actually a direct consequence of this productivity (Calvert and Pedersen, 1992). Other authors have considered that the main factor controlling organic accumulation is the presence of anoxic bottom water, which favours the preservation of organic matter, independently from productivity (Demaison and Moore, 1980; Tyson, 1995).

A more consensual vision is currently emerging from this controversy, acknowledging that both situations can be instrumental and that, more importantly, they are often interdependent. Furthermore, other additional factors have been proposed as influencing the process of the formation of organic-rich sediments, such as:

- The role of highly resistant biopolymers derived from algae, i.e. ‘algeanan’ (Largeau et al. 1990), and from higher plants, ‘cutan’ and ‘suberan’ (De Leeuw and Largeau, 1993). The organic material is better preserved when derived from specific populations of bio-organisms containing a large amount of such
substances. Massive organic rocks such as torbanites, solely made of preserved resistant remains of chlorococcale algae (*Botryococcus*), and cannel coals, solely made of spore remains, represent extreme examples of this process.

- The protection of organic compounds by sorption onto clays, leading to steric hindrance that prevents the degradation of the organic material associated with the minerals (Hedges *et al.*, 2001).

**Organic productivity.** For significant quantities of organic matter to accumulate in a sediment, the depositional environment must be associated with an ecosystem that produces a sufficient amount of biomass (Pedersen and Calvert, 1990). As a matter of fact, it is well documented that the present-day distribution of organic-rich surface sediments in the world ocean corresponds to the areas of high plankton productivity (Huc, 1988b). The production of primary organic matter is mainly based on photosynthesis occurring on land and within the euphotic layer of the water masses (= upper 100 m). In general, a small proportion (= 0.5%) of the organic matter produced on the present-day land surface escapes the continental biological cycle and ultimately makes its way into the seas where part of it eventually accumulates in coastal environments. Consequently, accumulations of terrestrial organic matter are more likely to occur at the outlets of rivers. Under specific sedimentary and climatic conditions, delta settings represent a unique type of environment, since a large volume of organic shales and coals can accumulate there as the result of the large volume of organic shales and coals can represent a unique type of environment, since a sedimentary and climatic conditions, delta settings occur at the outlets of rivers. Under specific of terrestrial organic matter are more likely to occur in intracratonic seas and lakes or in near-shore regions where rivers can supply nutrients originating from continental run-off by conveying the products of the chemical weathering of rocks. High productivity also occurs in areas where upwelling of deep ocean water enables the nutrient pool to be returned to the photic zone. For instance, high productivity is stimulated by coastal upwelling in areas where nutrient-poor surface waters are driven offshore by wind and currents, allowing their replacement by subsurface waters rich in nutrients. Modern sediments deposited under very active coastal upwellings (i.e. offshore Namibia or offshore Peru) are well known to be organic-rich. The Monterey Formation (Miocene of California) and Phosphoria Formation (Permian of West-Central USA) are examples of source rocks related to such ancient upwelling settings.

**Preservation of organic matter.** Living tissues are composed of an assemblage of bio-molecules, which are thermodynamically unstable. As soon as these bio-molecules cease to be involved in living processes, i.e. when they are secreted or excreted, or after the death of the organisms, they tend to loose their integrity and can be ultimately transformed into simple more stable components such as $\text{CO}_2$, $\text{H}_2\text{O}$, $\text{CH}_4$, $\text{NH}_4^+$, etc. This degradation can depend on physicochemical processes (oxidation, photolysis, etc.), but is predominantly mediated by biological processes.

Organic matter is actually a basic source of nutrients and energy for heterotrophic living organisms, including consumers (zooplankton, nekton, zoobenthos, land animals, insects and soil dwellers) and decomposers (microbial communities). The processes and efficiency of alteration as well as the resulting end-products of organic matter decomposition are to a large extent controlled by the availability of electron acceptors. The presence of adequate oxygen concentration (atmospheric or dissolved in water) provides a suitable living medium for organisms ranging from aerobic microbes to higher. In such a situation, the overall decomposition process corresponds to oxidation using molecular oxygen as an electron acceptor.

In the absence of molecular oxygen, anaerobic micro-organisms use nitrates followed by sulphates as an oxygen source in order to oxidize organic matter. Ultimately, when the medium is totally devoid of oxidants ($\text{O}_2$, $\text{NO}_3^-$, $\text{SO}_4^{2-}$) fermentative degradation occurs using organic matter itself as an electron acceptor, while methanogenesis takes place, via $\text{CO}_2$ and acetate reduction.

Degradation caused by aerobic organisms is by far the most efficient process for the breakdown of organic matter. It is enhanced by the mechanical and enzymatic breakdown of tissues due to feeding and...
digested by higher organisms. A minimum dissolved oxygen concentration of 0.1ml/l is required to sustain meio- and macro-benthos (Savrda et al., 1984). In oxygenated bottom environments, a significant percentage of the organic matter is consumed by benthic fauna on the seafloor and by burrowing organisms in the near-surface sediment. Moreover, the activity of bottom-dwelling animals results in a mixing of the upper layer of sediment (bioturbation) that significantly increases the time of exposure to decomposition processes. Furthermore, the burrowing activity maintains a circulation of water that replenishes the electron acceptors (dissolved oxygen, sulphates) in the sediment pores, thus fuelling the bacterial oxidative degradation of organic matter (Fig. 1).

These latter processes do not occur in anoxic environments because, as soon as molecular oxygen is no longer available, no organisms higher than bacteria can survive (Savrda et al., 1984). Anoxic conditions are toxic to macro- and meio-benthic fauna, including burrowers, and this leads to the formation of undisturbed laminated sediments in which water circulation is strictly limited (see again Fig. 1). In such an environment, organic preservation is enhanced by the lack of benthic animal scavengers and by the limited supply of electron acceptors to the sediment (Demaison and Moore, 1980). The extent of exposure to an oxygenated environment has been recognized to be of paramount importance for the preservation of organic matter in the sedimentary record, and is defined as the concept of ‘oxygen exposure time’ (Van Mooy et al., 2002).

However, it is important to stress that anoxia is not a depositional environment as such, but rather the result of an imbalance between the consumption and the replenishment of molecular oxygen. Consumption is controlled by the oxidation of organic matter by aerobic organisms, while replenishment is controlled by the efficiency of the transfer of atmospheric oxygen, by diffusion or convection, which is the only source of molecular oxygen, to the environment in question.

The depositional environments liable to anoxia correspond to high productivity areas, where the oxygen demand is high, due to the oxidation of large amounts of organic matter in the process of burial, and situations with restricted circulation of oxygen-rich surface water (in contact with the atmosphere) towards the bottom, due to geomorphologic features, such as silled basins, deep and narrow basins or water stratification. The latter results from the occurrence of different water bodies exhibiting marked density contrasts (i.e. fresh water overlying denser salt water, warm water overlying denser cold water).

Upwelling systems provide an example of anoxic conditions driven by productivity. The high level of organic production promotes the formation of an underlying anoxic core that eventually impinges on the continental platform, resulting in an open shelf setting which is particularly favourable for source rock deposition. In such areas, the input of large amounts of biosynthesized organic matter is associated with their preservation by anoxic bottom waters (Demaison and Moore, 1980).

Intra-cratonic silled seas, depressions on carbonate platforms, elongated and narrow seas, as well as deep elongated rift basins, all represent
examples of geomorphologic settings that lead to sluggish water circulation. This type of situation restricts the delivery of molecular oxygen within the water mass.

The Black Sea is an example of the occurrence of low-density fresh water, coming from rivers, overlying denser saline marine water. This causes the development of a density stratification which hinders oxygen renewal in the deep waters and which triggers anoxia. Due to their climatic regime, low-latitude lakes often exhibit water stratification owing to differences in temperature between the warm surface and cold bottom waters.

On a global scale, specific time intervals known as Oceanic Anoxic Events (OAEs) correspond to episodes remarkable for the deposition of widespread organic-rich sediments (Arthur and Schlanger, 1979). For instance, some OAEs are well-defined and identified during the Cretaceous, i.e. OAE 1a (Early Aptian), OAE 1b (Late Aptian-Early Albian), OAE 1c (Late Albian) and OAE 2 (Cenomano-Turonian boundary). These events are assumed to be associated with the extensive stratification of oceanic waters. They led to reduced ventilation and the development of dysoxic and anoxic conditions in minimum oxygen zones along the continental margins of the tropical Tethys Ocean, in restricted intra-cratonic seas and in basins of the widening Atlantic Ocean. These conditions gave rise to the regional deposition of organic-rich source rocks. The Iabe Formation of offshore Congo and the La Luna Formation in Venezuela, are examples of prolific source rocks associated with these Cretaceous anoxic events.

Transport of organic matter. An important aspect of the formation of organic-bearing sediments is the transport of organic material from the site of bio-production to the site of sedimentation. On the scale of the basin, transport is a determining factor as far as preservation and distribution are concerned. The influence of transport on preservation is directly linked to the extended concept of ‘oxidant exposure time’ (including the exposure to oxygen and to other oxidants such as sulphates).

In the aquatic environment, fresh organic compounds supplied below the base of the photic zone, owing to primary production, are highly reactive and likely to be intensively degraded by heterotrophic organisms. In fact, simple mass balance calculations (organic production versus organic material actually incorporated into the underlying sediments) suggest that degradation of organic matter detritus takes place largely within the water column. This hypothesis is supported by sediment trap experiments showing the almost exponential destruction of organic matter as a function of water depth/residence time (Suess, 1980).

The pure organic detritus which exhibit a low density (1-1.7 g/ml) are unlikely, as such, to play a significant role in vertical mass flux, because of their long residence time in the water column. However, the repacking of organic detritus and small particles by physico-chemical (i.e. flocculation) and biological processes produces large organo-mineral particles (fecal pellets, ‘marine snow’ and aggregates) which settle more rapidly and which can act as efficient carriers for the organic matter. The increased density and sinking rate of these particles have been shown to be related to high primary biological productivity (Dagg and Walser, 1986). It is again noteworthy that high productivity not only delivers a large amount of organic matter but also promotes a better preservation (oxygen demand and improved repacking of organic detritus), by increasing the efficiency of transport towards the underlying sediments. Thus, the enhanced preservation and sinking rates associated with high productivity probably explain the deposition of organic-rich sediments in specific deep-water settings, as seen in offshore Namibia for the sediments deposited since the late Miocene (Huc et al., 2001).

Together with the sinking rate of organic particles, water depth is a crucial parameter controlling the fate of organic matter in sedimentary environments. Under favourable conditions (productivity, anoxia, etc.), shallow water environments probably represent an optimal setting for the accumulation of substantial amounts of organic matter. This would occur as soon as the sediment floor becomes located below the storm wave base. With increasing depth, the longer exposure time of organic particles in the water column favours their continuing degradation. Understanding the role of water depth allows us to acquire some insight into the distribution of the source rocks in the perspective of sequence stratigraphy. In a given basin, sedimentary processes govern the lateral distribution of organic matter. The negative correlation between organic content and sediment grain size is well documented (Hunt, 1995). This phenomenon can be the result of an equivalent hydraulic behaviour for organic particles and fine-grain sediments, and the sorption of organics onto clays (Ransom et al., 1998; Hedges et al., 2001). In any case the organic matter tends to be winnowed, accumulating in depocentres that exhibit lower hydraulic energy. At the regional scale in epicontinental seas these depocentres tend...
to occur in the bathymetric lows of the basin, which produces a concentric pattern with a progressive centripetal increase in the organic content of the sediments (Fig. 2). This ‘bull’s eye’ pattern is documented in recent environments, i.e. Caspian Sea, Black Sea, Lake Bogoria in Kenya, etc., and in the sedimentary record, i.e. Upper Jurassic Bazhenov Formation in Western Siberia, Lias of the Paris Basin, Oligocene of the Dongying depression in China, Lower Jurassic of the northern North Sea, etc. (Huc, 1988a).

**Type of organic matter**

Source rocks are characterized by the nature of the organic matter they contain. The organic matter contained in the source rocks is the result of the fossilization of the organic remains of once-living organisms. According to the usual definition (Durand, 1980), kerogen is the part of this organic matter that is insoluble in organic solvents (such as chloroform or dichloromethane). In thermally immature sediments, the kerogen accounts for almost all the organic matter present. The kerogen is composed of more or less altered organic material directly derived from the biopolymers making up the tissues and products of the living precursors (inheritance). It also contains other products of the random polycondensation of intermediate moieties coming from the decomposition of these biopolymers (neoformation).

As previously mentioned, the main precursors of these organic remains are mostly algae, bacteria and higher plants. The relative contribution of these different precursors and their degree of alteration vary as a function of depositional environment. This is the principal factor controlling the properties of the kerogens.

Living organisms are constituted of biopolymers including proteins, carbohydrates (e.g. cellulose), lipids and lignin, the latter only being present in the tissues of the terrestrial higher plants. Hydrogen is the most abundant atom in petroleum compounds, followed by carbon, with hydrocarbon molecules themselves being made up only from these two specific atoms. In this respect, the most prolific kerogens in terms of petroleum generation are those containing the highest concentration of hydrogen and the least of oxygen. Most of the proteins and many of the carbohydrates are destroyed during the early diagenesis (the first tens or hundred metres of burial). However, regarding the carbohydrates, we should point out that cellulose is a notable exception. Cellulose is less subject to decomposition, and, to some extent, can ‘survive’ diagenesis with relatively little alteration. Broadly speaking, compounds derived from lipids (very rich in hydrogen), lignin (poor in hydrogen, due to its aromatic nature) and cellulose (rich in oxygen), and which are therefore the most resistant, are preferentially preserved and relatively concentrated in the resulting fossilized organic matter. The partly inherited and partly neoformed kerogen are thus exhibits the more or less altered chemical imprint of its precursors.

Within detrital environments (i.e. deltaic), the contribution of terrestrial higher plants implies the preferential occurrence of ligno-cellulosic material (i.e. wood fragments, etc.), exhibiting a low H/C atomic ratio and a high content of oxygen. This results in kerogens displaying a less hydrocarbon-prone character than the kerogens derived from algal (i.e. phytoplankton) or bacterial material: neither contain lignin, algae are poorer in cellulose (mainly occurring in cell membranes) than higher plants, and bacteria are devoid of cellulose.

The H/C and O/C atomic ratios of kerogens are conventionally used to sort the organic matter of sediments into three main practical and classical ‘types’. These types are schematically related to three main depositional environments (Fig. 3). Type I (H/C > 1.6, O/C < 0.1): lacustrine environments; Type II (1.2 < H/C < 1.6, 0.1 < O/C < 0.2): marine
environments; Type III (H/C<1.2, O/C>0.2): continental and marine detrital environments (i.e. deltas).

This is far from being a strict genetic classification; it merely allows an appreciation of the magnitude of the H/C and O/C parameters, which provide some information on the (initial) petroleum potential of the organic matter (Durand, 1980).

Below we give indicative values (% weight of the kerogen) of the amount of petroleum-like compounds that can be potentially released by the different types of kerogens during the thermal evolution. Type I: ~60-70%; Type II: ~40-60%; Type III: ~15-25%.

Historically, these types have been defined according to specific reference series (Tissot and Welte, 1984; Vandenhove and Largeau, in press). These include: Type I: Eocene, Green River Shale Formation (Utah, USA); Type II: Lower Toarcian shale, Western Europe (including the Schistes carton from the Paris Basin, France, and the Posidonian Schieffer from Germany); Type III: Upper Cretaceous from the Douala Basin (Cameroon) and the Miocene of the Mahakam Delta (Kalimantan, Borneo Island, Indonesia).

An important difference between the organic matter accumulated in marine and lacustrine environments is that the anaerobic degradation occurs in the presence of sulphates within marine systems, but generally without sulphates in freshwater lakes. Consequently, the anaerobic degradation of organic matter in marine environments corresponds to an oxidation (sulphates being the electron acceptor) producing H$_2$S. The anaerobic degradation of organic matter in freshwaters corresponds to fermentation, which may eventually be associated with methanogenesis activity resulting in the formation of methane (i.e. marsh gas).

A further aspect of the composition of kerogens is the sulfur content. A determining factor in the quality of the generated hydrocarbons (oils rich in sulphur versus sweet oils), sulphur has an influence on the kinetic behaviour of the kerogen during thermal alteration and is a minor constituent of living tissues. The sulphur content of a given kerogen is actually acquired by incorporation during the very first step of its geological evolution (early diagenesis). A kerogen is likely to be sulphur-rich if it has been deposited in a marine environment (due to the occurrence of sulphates in the medium), under anoxic conditions (anaerobic formation of H$_2$S and polysulphur compounds) and in an iron-depleted environment. Under such conditions, the inorganic sulphur species interact with the organic ones and are incorporated into the kerogen as organic sulfur moieties. In this context sulphur-rich kerogens are often associated with carbonate and pure siliceous environments. When present, iron has the property of preferentially scavenging the sulphur species and to form the precursors of pyrite. In such a situation, generally associated with siliciclastic environments, the formation of a sulphur-poorer organic matter is promoted (see again Fig. 3). A sub-Type has been designated to accommodate sulphur-rich marine kerogens: Type IIS.

**Distribution of source rocks in space and time**

*Main source rock habitat.* Most rocks with high organic content are deposited under specific geological, oceanographic and climatic conditions, for example:

- Intra-cratonic depressions flooded during high sea-level stands, which are often separated from the open sea by sills, and consequently liable to become anoxic. Nutrients fuelling the aquatic productivity are supplied by the surrounding landmasses. The Upper Jurassic Bazhenov Formation of Western Siberia, the Cretaceous Interior Seaway of Western USA and the Early Lias of the Paris/German Basin are documented examples of such settings.

- Marginal basins associated with depressions within carbonate platform complexes. Examples are provided by the source rocks deposited in the Arabian-Persian Gulf during the Upper Jurassic and the Cretaceous: Hanifa Formation, Shilaif Formation, Shuaiba Formation, Kahzdumi Formation etc.

- Continental shelves and continental slopes when associated with upwelling systems. A situation encountered in the Miocene Monterey Formation of California.
Rift environments likely to develop lacustrine source rocks in hot and humid climates. Many examples of lakes with especially organic-rich sediments are reported in recent corresponding environments, i.e. Lake Kivu and Lake Tanganyika, as well as in the geological record (e.g. the Early Cretaceous of the African and South American margins, Bucamazi Formation, Lagoa Feia Formation, and the Eocene-Oligocene Pematang Formation of Central Sumatra).

Elongated and narrow basins related to the early stages of oceanic opening are conducive to the development of anoxia when invaded by marine waters, e.g. the Upper Jurassic Kimmeridge Clay Formation of the North Sea, and Cretaceous source rocks of the South Atlantic.

Deltas containing thick deposits of organic-rich shales and coal, e.g. the Miocene Mahakam Delta in Kalimantan, Indonesia, and the Tertiary Niger Delta.

Stratigraphic distribution of source rocks. The average content of organic matter in the sedimentary record is known to vary considerably, ranging from lean (<0.5% organic matter) to rich, with 5 to 40% in shales and up to nearly 100% in humic and algal coals. At a global scale, the chronostratigraphic distribution is irregular, and major accumulations of sedimentary organic matter, and thus source rocks, seem to be concentrated within a limited number of specific stratigraphic intervals. The abundance of source rocks at other periods of geological time is estimated to be very minor (Bois et al., 1982; Klemme and Ulmishek, 1991).

According to different approaches the relative contributions of the source rocks belonging to the six most important intervals are as follows: Silurian (450-420 My), 18-20%; Upper Devonian-Lower Carboniferous (380-340 My), 14-18%; Upper Carboniferous-Lower Permian (310-280 My), 13-18%; Upper Jurassic (170-150 My), 15-17%; Middle-Upper Cretaceous (110-90 My), 17-24%; and Oligocene-Miocene (40-5 My), 7-14% (Klemme and Ulmishek, 1991; Huc et al., in press).

In Fig. 4, we show a plot of the global organic carbon burial during the Phanerozoic (545-0 My), based on carbon isotope measurements (Berner, 2003), compared with the curve of the accumulation rate of organic matter in source rocks (sediments with TOC, Total Organic Carbon, >3%; Huc et al., in press) and the curve of the intensity of tectonic degassing normalized to the present-day value (Berner and Kothavala, 2001). This diagram shows that the peaks of CO2 degassing are in phase with the global accumulation of organic matter in sediments and the formation of significant regional source rocks.

This relationship can be tentatively rationalised in terms of the currently accepted bio-geochemical carbon cycle (Holland 1978; Westbroek, 1992) and the results of modelling studies by Robert A. Berner (Berner and Kothavala, 2001). These models suggest that the increase of partial pressure of CO2 in the atmosphere (P_{CO2}) promotes enhanced chemical
weathering of rocks. This is mainly mediated by plants which disrupt and chemically destroy the bedrocks by the action of their root systems and associated microorganisms in the rhizosphere. Such micro-organisms produce aggressive acids in order to extract nutrients, metals and oligo-elements needed for their growth from minerals. The chemical weathering on continental masses yields ionic species including Ca$^{2+}$, HCO$_3^-$ and nutrients that feed into surface and ground waters. The dissolved Ca$^{2+}$ and the HCO$_3^-$ are transported to the sea where they are precipitated as carbonates, mainly through biological processes. These carbonate deposits ultimately act as a sink for the atmospheric CO$_2$.

With the notable exception of the Silurian, it can be seen that, at the first order scale, the periods of enhanced organic matter accumulation correspond to periods of favoured carbonate accumulation (Ronov et al. 1980). Both phenomena lead to a natural sequestration of CO$_2$ during periods of increasing atmospheric CO$_2$. The increased PCO$_2$ induces a negative feedback phenomenon that ultimately reduces the atmospheric CO$_2$ by storage in the form of carbonates. Such deposits represent the largest reservoir of carbon in the Earth’s crust (75%), while sedimentary organic matter accounts for the remainder (25%) (Hayes et al. 1999).

The process of photosynthesis controls the intensity of chemical weathering. As photosynthetic activity increases, soil formation intensifies and deepens as land plants take up their need for mineral-derived nutrients. Indeed, while these nutrients are actively recycled by land plants, they are ultimately transported, along with the Ca$^{2+}$ and HCO$_3^-$ ions, to lakes and marine waters, thus enhancing the plankton productivity for which nutrient availability is the main limiting factor (Holland 1978). Consequently, increased P$_{CO_2}$ levels can be considered a potentially major factor in enhancing organic matter accumulation. Keeping all other parameters constant, increasing the P$_{CO_2}$ is actually reported (Mellilo et al., 1993) as substantially enhancing the primary productivity of land plants on continents (CO$_2$ fertilisation). It also promotes chemical weathering and increases the input of nutrients to soils, streams and eventually to the seas and oceans. As a consequence, we may tentatively propose that the secular increase of the atmospheric CO$_2$ (first and second order cycles), although acting in an indirect way, is a key factor in the deposition of source rocks within given stratigraphic intervals on the global scale. The CO$_2$ fertilisation of the land biomass induces an accelerated formation of deeper soils associated with enhanced chemical weathering. This increase of the nutrients pool on the global scale therefore promotes aquatic productivity as well as the accumulation of organic matter in sediments.

This model emphasises the role of primary productivity in the formation of organic-rich sediment at the global scale and in the long term (first and second order cycles). At the same time, it reconciles the observed correlation between geological time intervals hosting large volumes of source rocks, and episodes of increased atmospheric CO$_2$ due to accelerated secular tectonic activity. However, these periods of increased rates of subduction, metamorphism and volcanism, that introduce CO$_2$ into the atmosphere, are also associated with periods of high sea level. At such time, large areas of the continental shelf are flooded, giving rise to widespread intra-cratonic seas that set the stage, at the global scale, for a better preservation of the biologically produced organic matter (Tissot and Welte, 1984). Moreover, most of these periods are characterised by the extensive deposition of carbonates (see above) on continental shelves that often form widespread platforms harbouring shallow intra-shelf basins. These depositional settings, including epicontinental seas and intra-shelf basins, favour the formation of depositional environments in shallow, isolated or silled basins where water bodies are liable to develop anoxic bottom conditions due to a lack of renewal of dissolved oxygen. Moreover, the sinking organic matter has a reduced residence time within a water column of limited thickness. Both factors imply a decreased ‘oxygen exposure time’ for the accumulating organic matter, which enhances its chance of preservation (Van Mooy et al., 2002). The whole process may be referred to as CO$_2$-induced eutrophication.

Although organic-rich deposits, acting as source rocks, are well documented since the latest Proterozoic, the first appearance of widespread source rocks at the global scale corresponds to the rise of land plants during the Silurian. This observation is significant in that land plants are instrumental in soil formation and chemical weathering. Prior to the Middle Silurian the land surface was probably either formed of exposed bedrock or covered by thin microbial protosoils (Algeo et al., 2001).

When considering the distribution at the first-order scale, there is an apparent offset between the times of maximum accumulation of bulk organic matter in source rocks, and the times of maximum occurrence of coal and Type III source rocks (Ronov et al., 1980; Bois et al., 1982; Klemme and Ulmishek, 1991), as illustrated by the change in the (coal + Type III)/total source rocks ratio (Bernier 2003). This pattern characterises both the Palaeozoic megacycle and the Mesozoic megacycle (see again Fig. 4, and Fig. 5). Coal and Type III source rock deposits are subordinate during the organic-
enriched Upper Devonian-Lower Carboniferous interval, and become prolific during the following organic-tending interval of the Upper Carboniferous-Lower Permian. Similarly, coal deposits are limited during the organic-prone Upper Jurassic interval, but increase in abundance during the Middle-Late Cretaceous and Oligo-Miocene organic-tending intervals, towards the end of the Mesozoic megacycle. The biological evolution in the nature of the terrestrial biomass induced by the progressive colonisation of continents by land plants could explain the shifted distribution of coals and Type III source rocks towards the end of the Palaeozoic megacycle.

However, a different model is required to account for this recurrent time shift in the case of the second megacycle. It is generally recognised that the accumulation of coal beds requires an equilibrium between the creation of accommodation space and sedimentary supply. The most favourable scenario for the accumulation of thick coals corresponds to a vertical stacking depositional regime, associated with a low rate of base-level change in a system undergoing continuous and regular subsidence (Diesel, 1992; McCabe and Parrish, 1992; Bohacs and Suter, 1997). On a world-wide scale such a situation can be envisaged at the end of major orogenic phases due to a global relaxation of tectonic stresses (Dewey, 1988). In the continental realm, this model applies to the major foreland basins, that are common settings for coal deposits, such as the Carboniferous coal measures of northern Europe and the Appalachians, the coal deposits associated with the Cretaceous interior seaway of north America and the Tertiary Guadalas coal beds of Colombia. Similarly, in the oceanic realm, the progressive cooling of the ageing oceanic crust on passive margins represents another setting where considerable accommodation space can be created for the accumulation of major delta systems, such as the Tertiary deltas of the South Atlantic margins. In this context the rare occurrences of coal and Type III source rock deposits at the beginning and at the peaks of the first-order megacycles, and their abundance at the aftermath of such periods, could be explained in terms of global tectonic conditions, which are more conducive for coal accumulation at the end of major orogenic phases.

An apparent regional co-occurrence of Type III organic matter (including coal deposits) and lacustrine Type I source rocks is often observed in the six considered time intervals (see again Fig. 5). This pattern may be related to climatic conditions that are conducive to the accumulation of organic matter in lakes, as well as the formation of extensive coal deposits. To some extent this may also be because the water bodies of the related paleo-lakes might have benefited from the proximity of highly productive land vegetation which supplied nutrients triggering a high aquatic productivity.

Fig. 5. Schematic distribution of major source rocks during:
A, Silurian (450-420 My); B, Upper Devonian-Lower Carboniferous (380-340 My); C, Upper Carboniferous-Early Permian (310-280 My); D, Late Jurassic (170-150 My); E, Middle-Upper Cretaceous (110-90 My); F, Oligocene-Miocene (40-5 My).
Source rocks in a sequence stratigraphy perspective. In terms of sequence stratigraphy, certain authors have noted a potential relationship between the main periods of organic accumulation in the rock record and the first and second order increase of sea level caused by tectonic shifting (Tissot and Welte, 1984; Huc, 1991). This has led to the proposal that the most favourable stratigraphic locations for the development of source rocks correspond to global downlap surfaces (termination of basal strata with sigmoidal geometry on an underlying surface) associated with major cycles of marine ingressions onto the continents (Duval et al., 1986). This holds true for the higher third and fourth order cycles. The organic-rich intervals are usually associated with the maximum flooding surface, and more widely with the end of the retrogradation of the depositional systems towards the coastal areas and the beginning of the progradation of the depositional systems towards the open sea, as long as deposition occurs below the storm wave base (Pasley et al., 1991). This stratigraphic position is well documented for the Kimmedridgian/Tithonian in north-western Europe, the Lias of the Paris Basin, the Paradox Formation (Upper Carboniferous) in the western United States and the Natih Formation (Cenomanian/Turonian) of northern Oman.

Such a scenario can be explained by the occurrence or co-occurrence of various conditions which favour some rock sedimentation, and which occur during the development of depositional systems, including:

- The occurrence of widespread intra-cratonic seas, favoured by high sea level stand, in which a high concentration of organic matter could be triggered by the input of nutrients conveyed by rivers draining the chemical weathering products of surrounding continental surfaces, or by the introduction of nutrients from previously exposed soil horizons, as a result of the erosion associated with progressing flooding of coastal areas (Katz, 1994).
- The occurrence of shallow, isolated or silled basins that are conducive to the formation of anoxic bottom conditions due to the lack of renewal of dissolved oxygen (Demaison and Moore, 1980). This also implies a water column of limited thickness reducing the residence time for sinking organic matter.
- The increased concentration of organic matter in the basin mainly due to a reduced dilution with clastic/carbonate build-up trapped in marginal areas. In some cases this leads the source rocks to be expressed as a condensed section (Creaney and Passey, 1993; van Buchem et al., in press).

Conclusions

Source rocks play a central role in the formation of oil and gas accumulations in petroleum systems. The specific conditions of their formation, the factors controlling organic matter content and quality, as well as the rationale underlying their stratigraphic and regional distribution, have been the subject of a considerable amount of work during the last few decades. The resulting concepts are now widely used in conjunction with seismic sections (see Chapter 2.3), wireline logs and data collected in the well, through a series of analytical approaches allowing the determination of source rock attributes (organic content and type) at the sample scale. The resulting models can be used as a guide to assess the occurrence, quality, thickness, stratigraphic distribution and lateral extent of source beds in sedimentary basins, and provide improved input data for basin modelling applications (see Chapter 2.4).

References


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