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Résumé de l'article

La mise au point d'instruments d'arpen-tage de haute précision au cours destrente dernières années a permis la car-tographie systématique des bassinsocéaniques, ce qui a complètement mo-difié notre vision des fonds océaniques.Alors que l'on croyait jadis qu'il s'agis-sait d'une morne plaine sans traits dis-tinctifs recouverte de sédiments, on saitmaintenant que sa configuration et sesstructures sont complexes. Contraire-ment aux surfaces terrestres dont lesformes sont largement tributaires desprocessus d'érosion, les fonds océa-niques conservent un registre beaucoupplus continu des processus volcaniqueset tectoniques qui l'ont modelé. Lesbassins océaniques sont composés engros de marges continentales, de fondsabyssaux et de crêtes médio-océa-niques. Les marges océaniques passives, comme celles qui constituent lepourtour de l'océan Atlantique, présen-tent généralement un plateau continen-tal étendu surmontant de fortes épais-seurs de roches sédimentaires, alors quele plateau continental des marges activesest plutôt étroit, et comportent des zonesde subduction et des arcs volcaniques.Alors que les dépôts de la plaine abys-sale des bassins océaniques qui jouxtentles marges passives sont des turbiditeset que sa topographie est typiquementplane, la bathymétrie des plaines abys-sales des marges actives est beaucoupplus irrégulière et parsemée de nom-breuses collines abyssales. La morpho-logie des crêtes médio-océaniques quiceinturent la planète varie en fonction dutaux d'expansion, et sa conformationsegmentée est reliée de près aux proces-sus volcaniques et tectoniques de l'ex-pansion des fonds océaniques.

SERIES

OCEANIC LITHOSPHERE 2. The Origin and Evolution of Oceanic Lithosphere: Bathymetry and Morphology of the Ocean Basins

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SUMMARY
The development of high-precision surveying instruments over the last 30 years has made possible the systematic mapping of the ocean basins, completely changing our view of the sea floor. What was once thought to be a featureless plain covered with sediment is now known to contain a variety of complex and intricate structures. Unlike terrestrial surfaces, which owe their shape largely to erosion, the sea floor preserves a much more continuous record of the volcanic and tectonic processes by which it was formed. The ocean basins are divided broadly into continental margins, abyssal depths, and mid-ocean ridges. Passive margins, such as those surrounding most of the Atlantic Ocean, typically have broad continental shelves underlain by great thicknesses of sedimentary rock, whereas active margins have narrow shelves and are characterized by subduction zones and volcanic arcs. Ocean basins adjacent to passive margins typically have flat abyssal plains underlain by turbidites, whereas those adjacent to active margins typically have a more irregular bathymetry broken by numerous abyssal hills. The morphology of the world-encircling mid-ocean ridges varies with spreading rate, and reveals a pattern of segmentation closely linked to the volcanic and tectonic processes associated with sea-floor spreading.

RÉSUMÉ
La mise au point d'instruments d'arpentage de haute précision au cours des trente dernières années a permis la cartographie systématique des bassins océaniques, ce qui a complètement modifié notre vision des fonds océaniques. Alors que l'on croyait jadis qu'il s'agissait d'une morne plaine sans traits distinctifs recouverte de sédiments, on sait maintenant que sa configuration et ses structures sont complexes. Contrairement aux surfaces terrestres dont les formes sont largement tributaires des processus d'érosion, les fonds océaniques conservent un registre beaucoup plus continu des processus volcaniques et tectoniques qui l'ont modelé. Les bassins océaniques sont composés en gros de marges continentales, de fonds abyssaux et de crêtes médio-océaniques. Les marges océaniques passives, comme celles qui constituent le pourtour de l'océan Atlantique, présentent généralement un plateau continental étendu surmontant de fortes épaisseurs de roches sédimentaires, alors que le plateau continental des marges actives est plutôt étroit, et comportent des zones de subduction et des arcs volcaniques. Alors que les dépôts de la plaine abyssale des bassins océaniques qui jouxtent les marges passives sont des turbidites et que sa topographie est typiquement plane, la bathymétrie des plaines abyssales des marges actives est beaucoup plus irrégulière et parsemée de nombreuses collines abyssales. La morphologie des crêtes médio-océaniques qui ceinturent la planète varie en fonction du taux d'expansion, et sa conformation segmentée est reliée de près aux processus volcaniques et tectoniques de l'expansion des fonds océaniques.

INTRODUCTION
Man has been exploring the continental landmasses for millennia. During the Renaissance, as a result of the explosion in trade and commerce, the continental shorelines were explored and later exploited over a period of 300 years. In contrast, exploration and mapping of the ocean floor has been accomplished largely over the last 50 years.

Although some features of the deep sea floor were discovered as early as 1872 during the expedition of HMS Challenger, systematic mapping of the ocean basins did not begin until after World War II. By the late 1950s, some reasonably well-informed ideas on the evolution of sea-floor structures had been formulated and enough bathymetric data had been collected to produce maps of a world-encircling system of underwater volcanic mountains known as ocean ridges (Heezen *et al.*, 1959: Heezen and Tharp, 1977). Per-

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haps the greatest advances were made through the development of the echo-sounder, which measures the time taken for a sound pulse generated at the sea surface to return to the surface after reflecting off the sea floor. This travel time is converted to the depth of water using a standard velocity/depth curve for seawater. The earliest echo sounders generated a wide beam, which had poor spatial and depth resolution, and were used for locating and charting features rather than detailing their morphologies. However, by 1950 precision depth recorders were available which could measure water depths to an accuracy of 1-2 m. In the 1970s a further advance in instrumentation took place with the introduction of multi-beam echo-sounders, particularly Sea Beam. Sea Beam is a narrow beam instrument capable of focusing in on a 400 m-wide swath in water depths of 4000 m. Such multi (narrow) beam instruments, coupled with further developments of the wide beam type, revolutionized ocean floor mapping by allowing for the production of bathymetric swath maps with a resolution of 1 part in 5000. These high-resolution, two-dimensional maps now provide us with the shapes of sea-floor structures and a previously impossible ability to distinguish among many different small-scale features. Two further methods of investigating sea-floor topography have also been developed. The first is the deep tow sidescan system, which is lowered to within 100 m of the sea floor and "flown" over the sea bed, producing a highly detailed relief image of areas on either side of the flight path. The second is the use of highly sophisticated submersibles to make direct observations of the sea floor. Modern machines, whether manned or robotic, can dive to deep ocean depths and have the advantage of being able to take specifically targeted samples from accurately known positions. Lastly, it should be pointed out that many of these advances would have been much less effective had it not been for the development of global positioning and navigation using Earth orbiting and geostationary satellites.

In spite of the huge amount of data collected since the 1950s, there are still parts of the ocean floor that are poorly surveyed, particularly in the Indian and Southwest Pacific oceans. Since the mid 1970s measurements of the marine geoid, *i.e.*, the shape of the surface of equal gravitational potential energy around the Earth, which equates with the shape of mean sea level, have allowed for refinement of many bathymetric maps of deep water areas because undulations in the geoid correspond roughly with water depth. In shallower regions, satellites such as Seasat can actually reveal the nature of the sea floor through imaging radar.

Our view of the sea floor has therefore changed rapidly as our ability to image it has improved. What was once thought of as a featureless expanse covered in sediments is now seen to contain complex and intricate structures. Together with geochemical studies, seismic reflection, refraction and tomography experiments, and studies of magnetization and bottom photography, mapping of the sea floor provides important clues to the volcanic, tectonic and sedimentological processes acting in the ocean basins.

MORPHOLOGY OF THE OCEAN BASINS

Three major morphological provinces are recognized in the oceans: continental margins, abyssal depths, and ocean ridges (Fig. 1). Continental margins account for about 25%, abyssal depths about 41% and ocean ridges about 33% of the total area covered by the oceans (Wyllie, 1971). Other features, such as trenches, rises and oceanic plateaus, although of geological importance, are areally insignificant.

Continental Margins

Although variable in character, continental margins can be divided into two categories: passive margins such as those surrounding much of the Atlantic Ocean and active margins such as those bordering the Pacific Ocean.

Passive margins are characterized by continental shelves of widths up to 400 km, which are connected to the abyssal depths by the continental slope and rise. The continental shelf is that part of the sea floor lying between the coastline and the continental edge or shelf break, the point at which the sea-floor gradient steepens abruptly from a few degrees to slopes in excess of 10° (Fig. 2). Although variable, the shelf break averages about 130 m water depth. The continental shelf is a remarkably flat expanse, typically with less than 20 m of relief. Seaward of the shelf break is the continental slope, a relatively narrow zone with an average gradient of about 1:40, underlain generally by continental or transitional crust capped by a thick accumulation of terrigenous sediments. The continental rise typically is underlain by a gently sloping wedge of sedi-

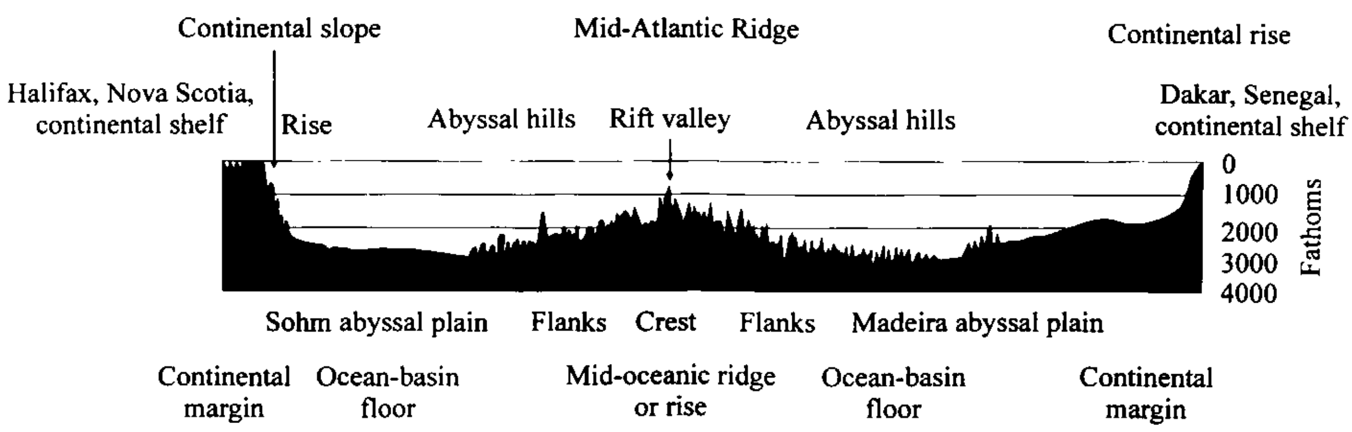

Figure 1 Cross-section of the Atlantic Ocean showing the major physiographic features of the ocean basin-continental margins, abyssal plains, and mid-ocean ridge. Not to scale (after Holcombe, 1977).

ments several kilometres thick, lying between the steeper continental slope and the almost horizontal abyssal plain. Relief is low and the average gradient is between 1:100 and 1:700. Both the continental slope and rise are cut by numerous submarine canyons, which typically head in the continental shelf well shoreward of the shelf break. These act as conduits through which sediment is transported from shallow regions directly to the deep ocean floor as density flows.

Active margins are so called because of the tectonic activity that characterizes them, resulting from their association with the convergence of lithospheric plates. They are marked by narrow continental shelves which pass through a steep slope directly into a deep trench (Chilean type) or by a marginal basin and island arc complex separating the continent from the deep ocean (Mariana type) (Fig. 3). Both types are bounded seaward by long, narrow trenches, rarely more than 100 km wide, that reach depths in excess of 11 km and form the surface expression of the subduction zones in which lithospheric slabs descend into the mantle.

Recent sea-floor swath bathymetry and sidescan acoustic imagery of parts of the western Pacific margin have shown that there are many more submarine volcanoes than originally thought, both in the back-arc and fore-arc regions. Volcanoes in back-arc regions erupt basaltic pillow lavas and appear related to weaknesses in the lithosphere associated with arc stretching. Those in fore-arc regions may erupt highly magnesian, boninitic lavas. Many fore-arcs are also characterized by diapirs composed of serpentinite mud and serpentinized peridotite, in places containing inclusions of crustal rocks. The formation of these diapirs indicates the presence of mantle, and therefore a very thin crustal carapace, in the outer fore-arc above the subducting slab.

Abyssal Depths

The abyssal depths, comprising more than 40% of the ocean basins, lie between the continental margins and the ocean ridges. The morphology of these areas is much more complex than originally thought and includes such features as abyssal plains, abyssal hills, seamounts, oceanic rises, and oceanic plateaus. The abyssal plains are extremely flat: at first glance, featureless portions of the sea floor with gradients of less than 1:1000. They lie in water depths of between about 3 km and 6 km and extend laterally from anywhere between 200 km and 2000 km. In ocean basins adjacent to passive margins, they are underlain by vast thicknesses of clastic sediment, transported to the sea floor through the submarine canyons. In ocean basins adjacent to active margins, most of the continent-derived sediment is trapped in marginal basins or in the trenches, and never reaches the deep ocean floor. Pelagic deposits with much lower sedimentation rates are dominant in these basins. Thus, the floor of the Pacific Ocean is much more irregular than that of the Atlantic and is characterized by numerous abyssal hills, which are relatively small: steep-sided structures that rise less than 1000 m from the sea bottom. They are considered to be small volcanoes generated along the axes or on the flanks of oceanic ridges. In the Atlantic and In-

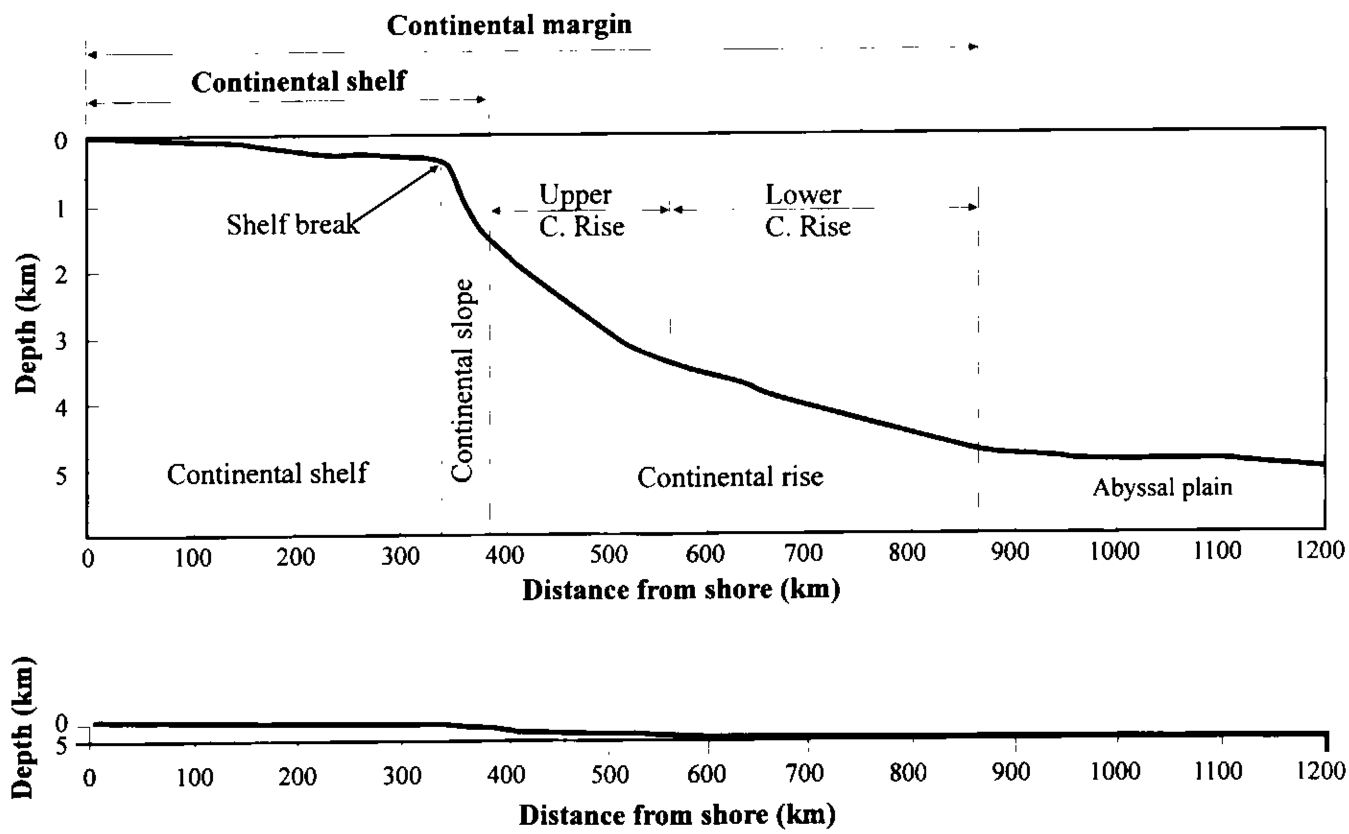

Figure 2 Morphologic features of a typical passive continental margin, including the continental shelf, continental slope and continental rise (after Anikouchine and Sternberg, 1973): (A) Vertical exaggeration 65:1; (B) Vertical exaggeration 4:1.

dian oceans, abyssal hills are common only along the boundary between abyssal plains and ocean ridges, occurring in clusters and groups. Whereas many abyssal hills must be covered by sediment in these basins, in the Pacific they visibly occupy as much as 80% of the deep ocean floor.

Seamounts are submarine volcanoes that rise more than 1000 m above the sea bottom, and are distinguished from abyssal hills by their size and distribution. They may be either conical or shield-shaped, with slopes that rarely exceed 15°. Many seamounts form linear chains (*e.g.*, Emperor Seamounts) and are associated with volcanic islands. Where they rise above the sea surface they are eventually eroded and subside to form flat-topped seamounts or guyots. These may be capped by

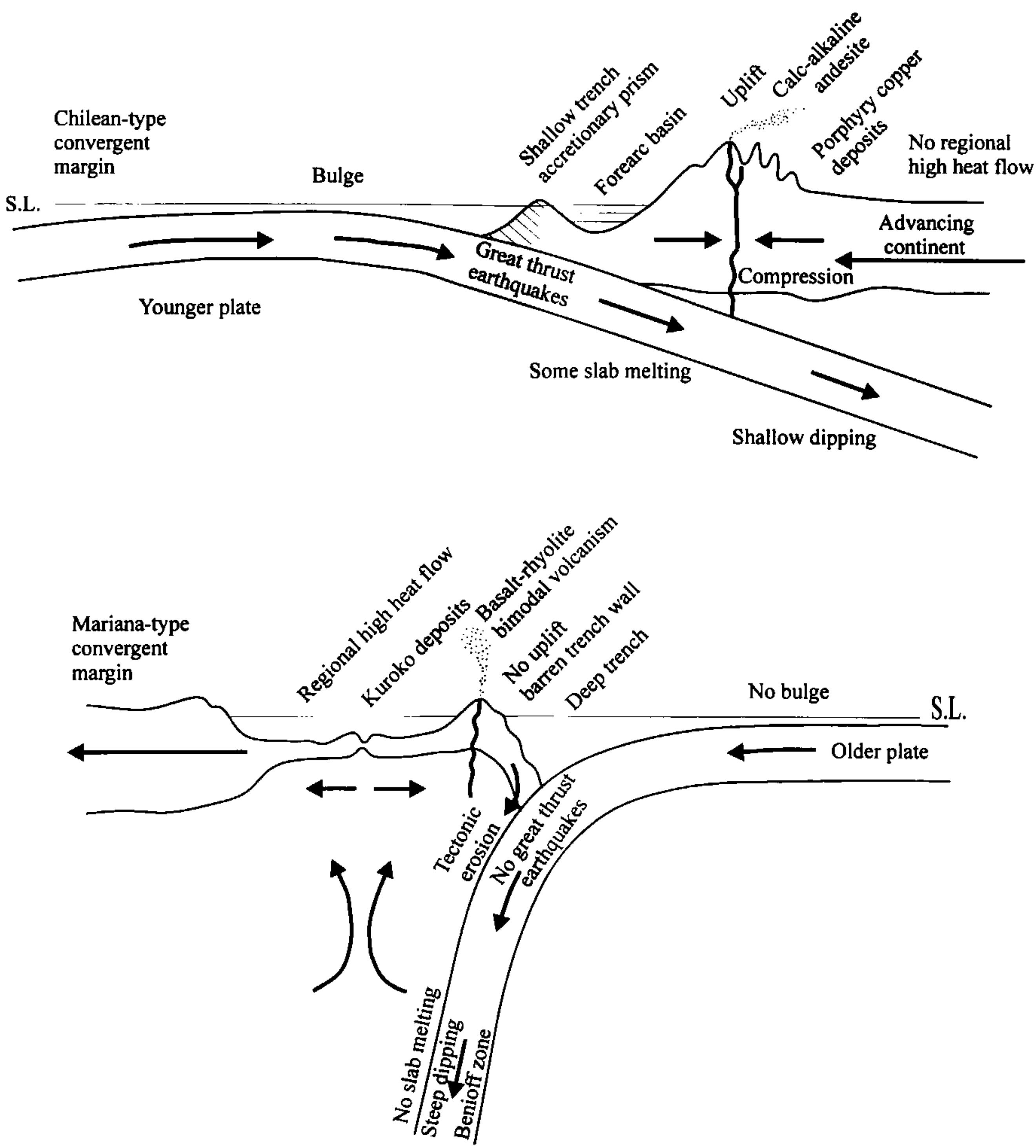

Figure 3 Morphological features of typical active continental margins (after Uyeda, 1978): (**A**) Chilean-type margin characterized by shallowly dipping subduction zone, shallow trench and calc-alkaline volcanism on the continental margin. (**B**) Mariana-type margin characterized by steeply dipping subduction zone, deep trench and bimodal volcanism in an intraoceanic arc.

shallow water sediments and surrounded by fringing reefs, but drilling has shown them to be volcanic in origin. Ocean islands, seamounts and guyots are all expressions of intraplate volcanism, typically ascribed to hot spot magmatic activity in the mantle.

Areally extensive, flat-topped features on the sea floor are oceanic plateaus. Examples include the Kerguelan Plateau in the Indian Ocean and the Ontong-Java Plateau in the Pacific. These rise some 2-3 km above the surrounding sea floor, have low relief, and are capped by sedimentary rocks. Drilling on both Kerguelan and Ontong-Java has revealed a volcanic substrate beneath the sedimentary cover formed by huge outpourings of basaltic lava onto the ocean floor. Thus, they are also interpreted as being related to mantle plumes emanating from hot spots and may be considered as large igneous provinces, akin to continental flood basalts.

Aseismic ridges are elevated, linear features distinguished from the major ocean ridges by a lack of volcanic and tectonic activity. Examples include the Ninety East, Broken and Chagos-Laccadive ridges in the Indian Ocean. The origin of these structures is not well understood but it has been suggested that they represent paleotransform faults. Broader elongate elevations of the ocean floor are known as oceanic rises or swells (*e.g.*, the Bermuda Rise), although some confusion might arise from the use of the term "rise" for part of the ocean ridge system (*e.g.*, East Pacific Rise). However, oceanic rises are more akin to aseismic ridges with smoother profiles than the ocean ridges.

Ocean Ridges

The ocean ridges are one of the most important physiographic features of the ocean basins and the sites where new oceanic lithosphere is formed, *i.e.*, divergent plate margins. Depending upon where these boundaries are drawn, the ridges occupy as much as 33% of the ocean floor and a large volume of the ocean basins. The ridge system is global and represents the largest and most volcanically active mountain chain on Earth, extending some 50,000 km and rising to an average water depth of 2500 m. Because of its importance in the generation of lithosphere, the ridge system has been the focus of numerous observations and experiments in the last 10 years, highlighted by efforts to investigate its structure by deep sea drilling and particularly through the international collaborative project, INTERRIDGE. The fundamental aspect of this programme has been to focus on oceanic ridges at a variety of scales to allow for sufficient spatial and temporal definition of the global system to test numerical models of the interaction of its various components. As a consequence, certain parts of the system are now known very well.

It is now evident that the physiographic characteristics of ocean ridges are related to the rate of spreading and crustal accretion (*e.g.*, Phipps-Morgan and Chen, 1993). This relationship is clearly seen in the classic relief maps of Heezen *et al.* (1959) and Heezen and Tharp (1977): slow-spreading ridges such as the Mid-Atlantic Ridge, where about 3 cm of crust are created on average annually, rise abruptly from the surrounding abyssal plains and display rugged topography, central rift valleys, and numerous small axial volcanoes; fast-spreading ridges such as the East Pacific Rise, which is generating lithosphere at some 6-17 cm·a^{-1}, have smoother, rounded profiles and generally lack a central rift valley. Local relief on slow-spreading ridges may reach up to 2000 m, whereas on faster spreading segments it is commonly less than 200 m. As expected, ridges with intermediate spreading rates exhibit intermediate topography and relief (Fig. 4).

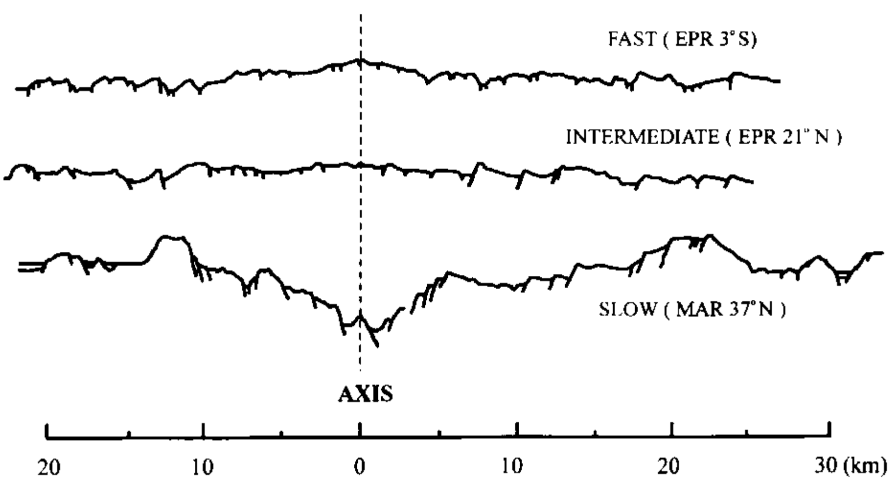

Figure 4 Typical profiles across fast-, intermediate- and slow-spreading ridges. Fast-spreading ridges have smooth profiles, an elevated crestal ridge and a narrow axial graben. Slow-spreading ridges are characterized by rugged topography, an axial valley 1-2 km deep, and small volcanoes at the ridge axis. Ridges with intermediate spreading rates have intermediate topography (after Macdonald, 1982). EPR, East Pacific Rise; MAR, Mid-Atlantic Ridge.

Many models have been proposed to explain the observed ridge axis topography. Tapponnier and Francheteau (1978) suggested that the rift valley at slow-spreading centres could be explained by steady state necking of the lithosphere, whereas Madsen *et al.* (1984) proposed an isostatic model for the axial topographic high and ridge crest gravity anomaly of fast-spreading ridges. More recently, Chen and Morgan (1990 a,b) have argued that the differences between the topographic profiles of fast- and slow-spreading ridges can be explained by the width of the decoupling zone between the brittle crust and ductile mantle. What is clearly apparent is that amagmatic, structural spreading processes are much more common in the case of slow-spreading centres where the supply of magma is restricted.

Regardless of the spreading rate, the depth of the sea floor increases exponentially away from the ridge crests as the newly formed lithosphere cools, thickens and ages (Fig. 5), and the close relationship between depth and age allows for the approximate dating of the ocean crust from bathymetry alone. This relationship breaks down in crust older than about 80 m.y. because little subsidence takes place beyond this age.

The ocean ridge system is repeatedly offset along a series of linear fractures, identified by Wilson (1965) as transform faults. These are long, narrow, strike-slip zones perpendicular to the ridge axis, seismically active between the ridge segments, but with aseismic extensions marked by deep valleys that extend hundreds of kilometres across the ridge flanks. The fracture zones that mark these transform faults are steep-sided troughs with depths up to several thousands of metres, floored with talus deposits derived from the steep walls. Considerable amounts of peridotite and gabbro have been dredged from fracture zone walls and floors, suggesting that the lower crust and upper mantle are exposed on the sea floor in these localities. Most fracture zones have a characteristic morphology. As the ridge crest approaches a transform fault or fracture zone, it bends toward the active segment and deepens significantly so that at the junction between the ridge crest and the transform valley there is a nodal basin that marks the deepest part of the fracture zone. Most of the plutonic rocks found in transform fault valleys come from the inside corner high, that portion of the wall lying between the active ridge segments (Bonatti and Michael, 1989; Dick, 1989) (Fig. 6). In this area, low-angle detachment faults dipping toward the ridge crest may expose lower crust and upper mantle by tectonically removing upper crustal units (Dick *et al.*, 1981) (Fig. 7). Such a detachment fault has been drilled on the inside corner high of the Kane Fracture Zone on the Mid-Atlantic Ridge (Karson and Lawrence, 1997) and similar faults have been inferred for the Atlantis Fracture Zone on the Mid-Atlantic Ridge (Cann *et al.*, 1997) and the Atlantis II Fracture Zone of the

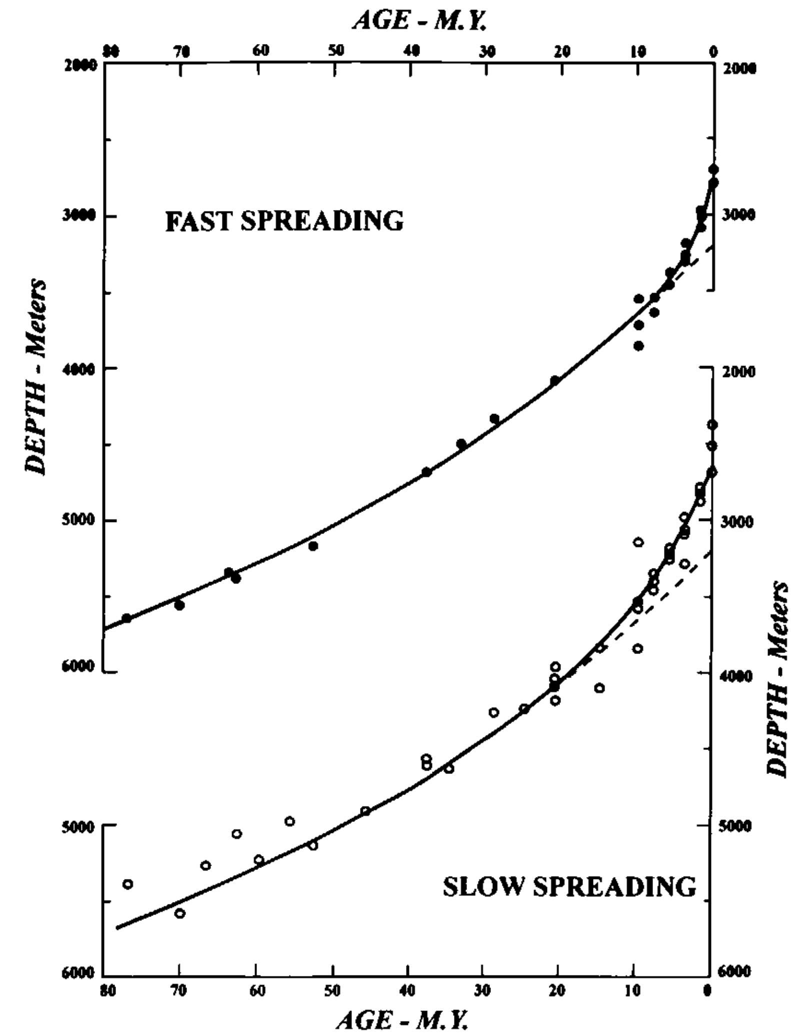

Figure 5 Plots of sea-floor depth *versus* age for fast- and slow-spreading ridges showing exponential subsidence as the crust cools (after Le Pichon *et al.*, 1973).

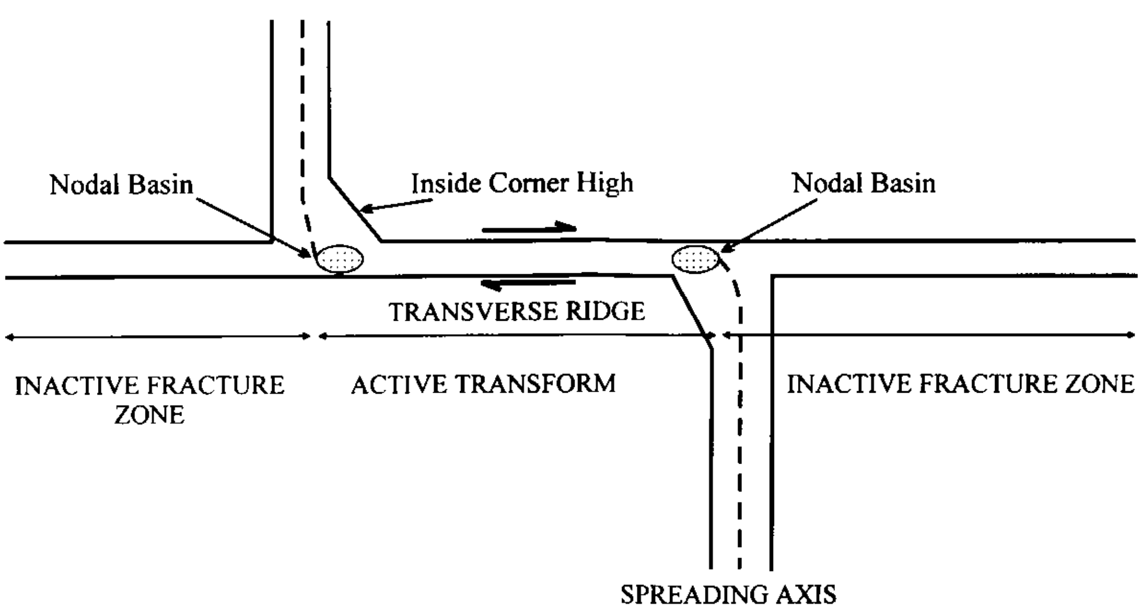

Figure 6 Geologic sketch map showing the major features of an active transform fault and associated fracture zone. Note curvature of ridge axis as it approaches the transform fault and the nodal basins at the ridge-transform intersection. The inside corners are characterized by topographic highs which form transverse ridges parallel to the transform as they migrate away from the spreading axis (modified from Karson and Dick, 1983).

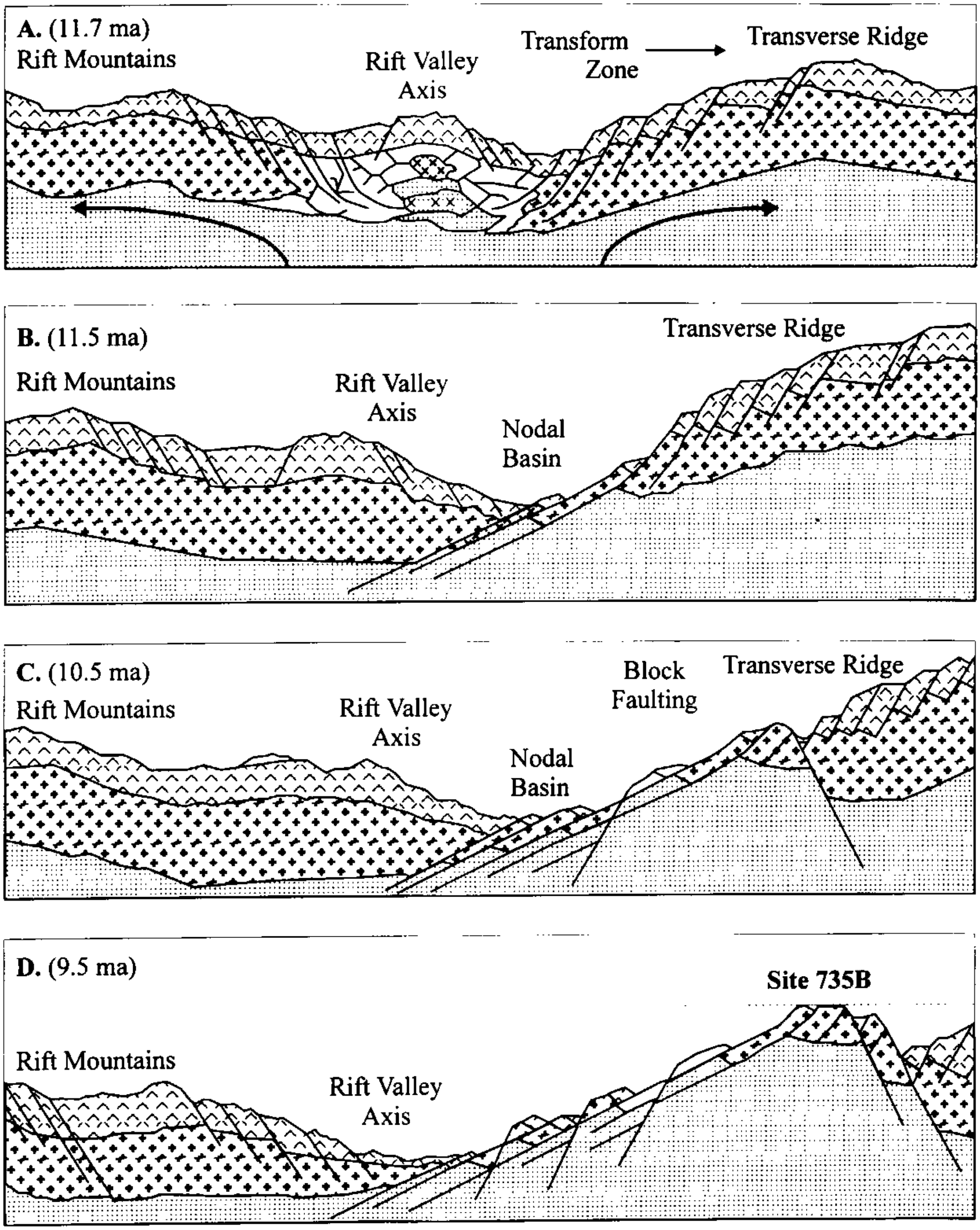

Figure 7 Diagram illustrating how a low-angle detachment can unroof the lower crust and upper mantle on the seafloor (after Dick *et al.*, 1991) (**A**) Initial symmetrical spreading at ridge axis with lithospheric necking; (**B**) Part of the new crust is welded to the underlying lithosphere forming a detachment fault during amagmatic spreading, (**C**) Extension of the crust is terminated by block faulting, (**D**) Crustal section is elevated to form an inside corner high which reaches sea level and is planed by wave action. The upper layer (inverted with ^ pattern) is layer 2 composed of pillow lavas and sheeted dykes, the middle layer (solid plus pattern) is layer 3 composed of gabbroic rocks and the lower layer (stipple pattern) is upper mantle.

Southwest Indian Ridge (Dick *et al.*, 1991). As the upper crustal rocks are tectonically thinned and removed, the lower crust rises to shallow levels to form a series of fossil inside corner highs, which eventually form a transverse ridge parallel to the fracture zone. The transverse ridge of the Atlantis II Fracture Zone on which ODP Site 735 is located, is believed to have formed in exactly this way (Dick *et al.*, 1991). What causes these lower crustal segments to rise is not clear but may be related in part to serpentinization of the underlying mantle peridotite.

Segmentation of the Ocean Floor

Almost any map of ocean floor relief will show at a glance how the sea floor is segmented on a variety of scales (Table 1). This segmentation is best recognized in the depth and continuity of the ocean ridges but in actual fact extends for hundreds of kilometres away from the ridge axes. Transform faults define first order segments (Fig. 8) and display offsets of the order of tens to hundreds of kilometres that juxtapose portions of oceanic crust of different ages (0-10 m.y.). The Mid-Atlantic Ridge is cut every 50 km or so by such faults, whereas the East Pacific Rise is much more continuous, and transform faults separate segments of the order of 100-km length. Nevertheless, these first order segments in the Pacific are themselves clearly broken by second and third order ridge axis discontinuities with offsets less than a few tens of kilometres, or indeed no apparent offset at all. A range of second and third order discontinuities has been described including overlapping spreading centres (OSC), non-offset transforms (NOT), propagating ridges (PR), oblique relay zones (OBZ), and simple bends in the ridge axis or deviations from the general alignment of the ridge (DEVALS) (Macdonald *et al.*, 1993a; 1993b). The difference between second and third order terminators is one of magnitude, but neither appear to be sustainable for long periods of time since ridge topography seems unaffected by them, once away from the axis. Third order segmentation seems to be the shortest lived, with no expression in crust more than 10^5 years old.

Second order segmentation is marked by "discordant zones" flanking the ridge. There is a clear difference between these features as developed at fast- and slow-spreading ridges. Macdonald *et al.* (1988) and Carbotte and Macdonald (1992) have shown that at fast-spreading ridges the axial discontinuities appear to migrate over a period of a few million years, along the strike of the ridge, leaving behind a series of inactive propagators and nodal basins with a distinctive topography. At slow-spreading ridges (Sempéré *et al.*, 1990), second order discontinuities appear to be longer lived and remain either in fixed position or migrate along strike to create a series of obliquely oriented ridges and basins.

Regardless of the scale of the segmentation, the along axis relief of each segment is characteristically arched (Fig. 9). The shallowest water depths are found above the midpoints of the segments and depths increase as the segment terminators are approached. The differences in water depths vary according to the scale of segmentation and the nature of the spreading centre. Thus first order segments on slow-spreading ridges may show variations up to hundreds of metres, whereas at fast-spreading ridges, the axial high may not vary greatly in depth along the length of the segment (Scheirer and Macdonald, 1993).

Segmentation of the ocean ridge system undoubtedly has its origin in magmatic processes (Whitehead *et al.*, 1984), although the different segmentation scales suggest a variety of causes (Forsyth, 1992). The undulating topography illustrated in Figure 10, can be used to define magmatic centres, or spreading units, that coincide with provinces reflecting the petrological and geochemical characteristics of basalts erupted at the axis. This is not a perfect correlation but it has led to the belief that both topography and magma composition are the result of physical (e.g. temperature) and chemical properties of the underlying mantle. However, tec-

Table 1 Characteristics of ridge segments (after Macdonald *et al.*, 1991).

	Order 1	Order 2	Order 3	Order 4
Segments				
Length (km)*	600±300 (400± 200)	140±90 (50±30)	50±30 (15±10?)	14±8 (7±5?)
Longevity (a)	>5 × 10^6	0.5 × 10^6 (0.5-10 × $10^{6)}$	~10^4-10^5 (?)	~10^2-10^4 (?)
Discontinuities				
Offset (km)	>30	2-30	0.5-2	<1
Age (a)**	>0.5 × 10^6 (>2 × 10^6)	<0.5 × 10^6 (<2 × 10^6)	~0	~0
Off-axis trace	fracture zone	V-shaped discordant zone	none	none

Note: Information in parentheses is for fast-spreading ridges (>6 cm·a^{-1}) if it differs from slow-spreading ridges
* Errors are 1 σ
** Age of seafloor juxtaposed to the spreading axis at discontinuity

tonic processes (Lonsdale, 1994) and crustal structure (Bell and Buck, 1992) also must play a vital role in ridge segmentation. The exact processes are still not well understood.

Along-axis variations in crustal thickness can also be correlated with bathymetry. The average thickness of the ocean crust is about 6 km, but may be as high as 7-8 km at the shallowest midpoint of a crustal segment. The crust thins rapidly in the vicinity of major transform faults to as little as 3 km on slow-spreading ridges (Fig. 10) and to about 5 km on ridges spreading at rates greater than $3 cm \cdot a^{-1}$ (Chen, 1992). These variations are thought to reflect a high magma supply rate at or near the midpoints of segments and a lower supply rate at the cold edges adjacent to transform faults.

There have been a number of attempts to model the segmentation of the ocean ridge systems in terms of the magmatic activity beneath the ridge axis (see summary in Schouten and Whitehead, 1992). The models are, however, too crude to explain the details now available through ridge studies. Nevertheless, it has been possible to show that the segmentation is not simply a surface phenomenon, but must extend for tens of kilometres beneath the ridges. Although questions still remain unanswered, including the controls on spreading rate and the major differences between fast- and slow-spreading ridges, it is clear that these processes involve a close interplay between magmatism, tectonic activity and hydrothermal circulation. These processes will be considered further in forthcoming articles in this series.

SUMMARY

Over the last 40 years highly detailed bathymetric maps have been produced for large areas of the ocean basins. These maps reveal a complex sea-floor topography which can be correlated with the petrologic, tectonic and hydrothermal processes acting in the ocean basins. The depth of the ocean floor reflects the density of the underlying rocks, which in turn is related to the age, composition and thermal characteristics of the crust and upper mantle.

Mid-ocean ridges are relatively high-standing regions because of the high heat flow associated with rising asthenosphere and the creation of new crust. Segmentation of the ridges occurs on a variety of scales, reflecting the underlying magmatic and hydrothermal processes. As the newly formed crust spreads away from the ridge axis,

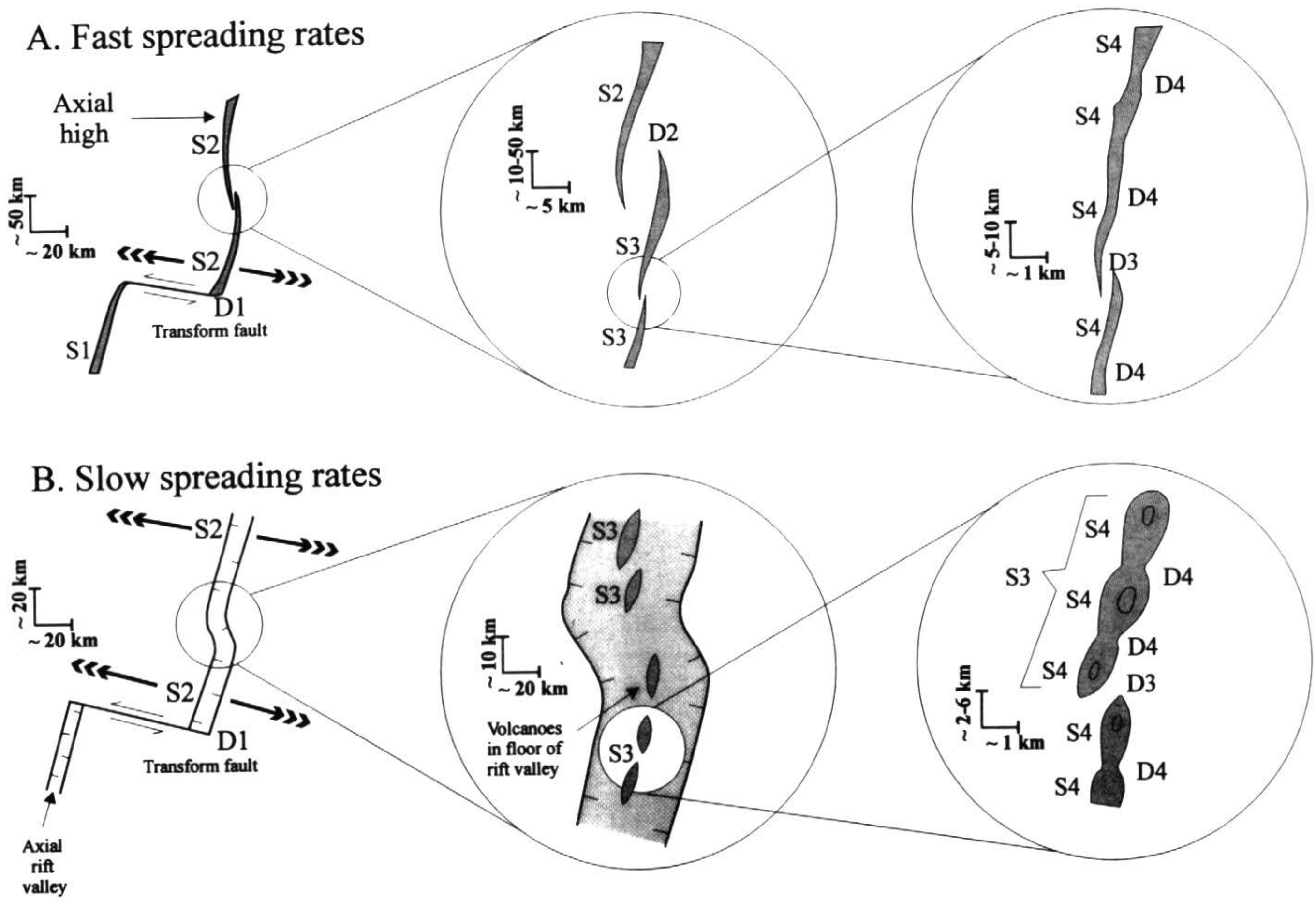

Figure 8 A hierarchy of ridge-axis discontinuities of order 1 through 4 for fast (**A**) and slow (**B**) spreading centres (after Macdonald *et al.*, 1991; Macdonald *et al.*, 1993a). S1, S2, S3, and S4 are ridge segments of order 1, 2, 3 and 4 and D1, D2, D3 and D4 are ridge-axis discontinuities of order 1, 2, 3 and 4. A ridge segment is first order if it is bounded at both ends by first-order discontinuities and second-, third- or fourth-order if it is bounded at one (or both) end(s) by second-, third- or fourth-order discontinuities. First-order discontinuities are always transform faults, whereas second-order discontinuities are overlapping spreading centres on fast-spreading ridges and oblique shear zones on slow-spreading ridges. Third-order discontinuities are small overlapping spreading centres on fast-spreading ridges and intervolcanic gaps on slow-spreading ridges. Fourth-order discontinuities are deviations from axial linearity (devals) on fast-spreading ridges and intervolcanic gaps on slow-spreading ridges.

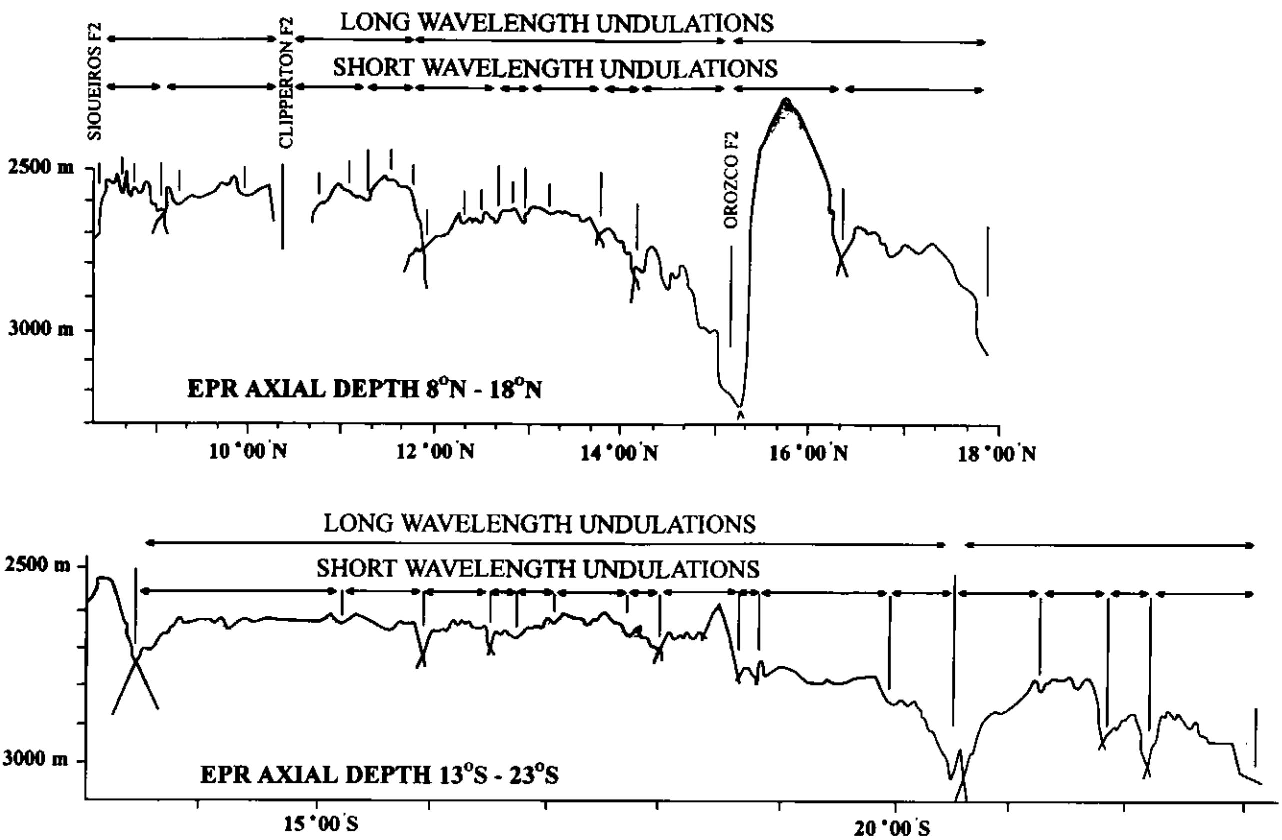

Figure 9 Axial depth profiles for the East Pacific Rise showing location of ridge-axis discontinuities of different orders: (**A**) East Pacific Rise 8°N - 18°N (after Macdonald *et al.*, 1993a,b); (**B**) East Pacific Rise 13°S - 23°S (after Sinton *et al.*, 1991). Long wavelength undulations are terminated at both ends by transform faults, whereas short wavelength undulations are terminated by a variety of ridge-axis discontinuities.

it cools and subsides, eventually reaching abyssal depths.

To a large extent, the morphology of the deep ocean floors reflects the nature of sedimentation in a particular ocean basin. Where submarine canyons transport large quantities of terrestrial sediment to the abyssal depths, the sea floor is typically a flat and featureless plain. In ocean basins dominated by slower pelagic sedimentation, the sea floor is much more rugged and irregular, broken by numerous abyssal hills. Locally, intraplate volcanism produces such features as oceanic plateaus, volcanic islands, and seamounts.

Landward of the abyssal depths, the ocean basins pass into the continental margins whose morphology reflects the tectonic processes by which they are formed. Passive margins are tectonically stable regions, formed by earlier continental rifting and are now marked by thick accumulations of sedimentary rock. Active margins are marked by deep trenches, which are the surface expression of subduction zones along which lithospheric plates are carried back into the mantle. Beneath continental margins, subduction zones have a relatively shallow dip and the trench marks the boundary between continent and ocean basin. Intra-oceanic subduction zones, like those in the western Pacific Ocean, are much steeper and are separated from the continent by a volcanic arc and marginal basin. The differences in active margins are reflected both in the character of the associated volcanism and the nature of earthquake activity.

Large portions of the ocean basins still lack detailed bathymetric maps and there is still much to learn from marine surveying. Equally important are technological advances in our ability to image and sample the sea floor. Ultimately, it will be the integration of petrological, geochemical, geophysical and geochronological data with detailed maps of the ocean floors that will elucidate the processes by which the oceanic lithosphere is created, modified and eventually recycled back into the mantle.

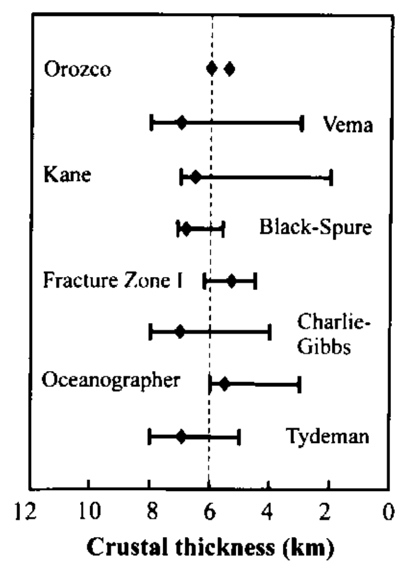

Figure 10 Crustal thickness variations at major fracture zones in the Atlantic Ocean. The solid diamond represents the average crustal thickness about 20 km from the fracture zone. The low end of the error bar represents the thinnest crust beneath the fracture zone and the high end of the bar represents the crustal thickness near the centre of the ridge segment, at its shallowest point (after Chen, 1992).

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