1

The Geological Background

Geology influences our lives and our communities in many ways: its clearest influence is in the production of wealth from the exploitation of minerals and oil. However, we must not underestimate the effects of rock composition and the availability of water on the productivity of soils and hence on agriculture and stockbreeding. Geology also influences the shape of the land, and thus the distribution of human populations. All these factors are important today, and were important in the past. The aim of this book is not simply to describe the geology of Greece and the Aegean, as that has already been done,[34,127] but also to comment on how geology influenced the development of the ancient civilisations of this region.

Before discussing these topics, a word on geological methods. The classical procedure, as practised in the nineteenth century, was essentially descriptive: a catalogue of what could be seen on the surface of the earth. Some attempt was also made to predict what could be found in areas hidden beneath the surface or concealed beneath water. Although the descriptive method is still an important part of geology, recent work has been more directed towards an understanding of the processes involved, and this has necessitated extensive borrowing from both physics and chemistry. This expertise has also been passed on to archaeology, giving rise to the new discipline of archaeometry.

In this chapter we will give an outline of the geological background, especially as regards the Aegean region. Many of these ideas are treated in more detail in subsequent chapters. We have tried to simplify some of the geological controversies, but geology is an active field of research and ideas and interpretations change with time. There are many good books covering various aspects of geology, such as *Geology in the Field*,[42] *Rocks and Rock Minerals*,[59] and *The Holocene: an environmental history*.[235]

Geological time

The earth is about 4,500 million years old, and was formed at the same time as the sun and the other planets. Geology is distinguished from most other sciences by its study of this immense period of time. However, geology is also concerned with processes on much shorter time-scales, and eventually blends into those of archaeology.[31]

Initially fossils were used to establish the relative age of different rocks. A limited number of geological periods were defined in terms of these fossils and given special names. More recently it has been possible determine the age of rocks in years, using naturally radioactive elements. However, we continue to use the named geological periods for a number of reasons, not least being that it is easier to refer to names instead of numbers. (See Fig. 1.1.)

The structure of the Earth

The outer core of the earth is largely composed of liquid iron and is the source of the magnetic field (Fig. 1.2). Above the core lie the lower mantle and the asthenosphere, which are the largest part of the earth, and are largely made of magnesium and iron silicates. The uppermost part of the solid earth is the lithosphere, comprising the uppermost mantle and the crust (Figs. 1.2, 1.3). The continental crust is 30-80 km thick and is rich in silicon and aluminium. It extends out underwater to the edge

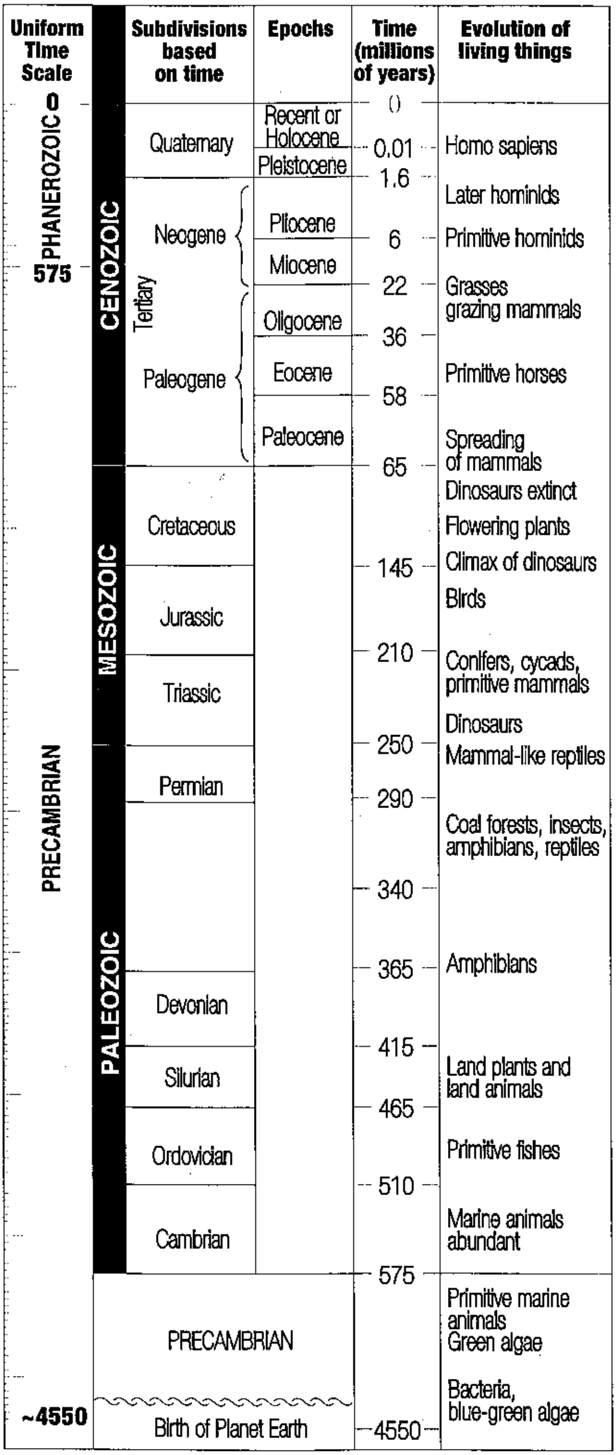


Fig. 1.1. Geological time.

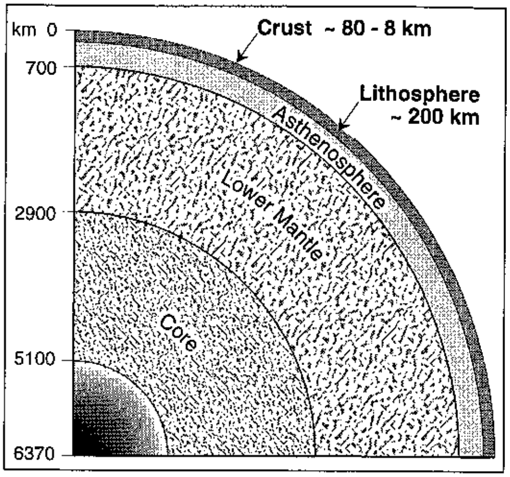


Fig. 1.2. The structure of the earth. The magnetic field is generated in the core. Convection in the mantle drives plate tectonics. The upper part of the earth is divided into the asthenosphere and lithosphere, which includes the crust. The crust is all that we normally see on the earth's surface.

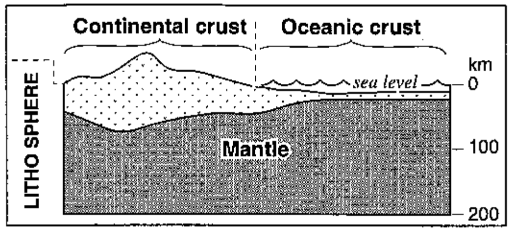


Fig. 1.3. Section through the lithosphere (not to scale). The continental crust is much thicker than the oceanic crust, and has a different composition.

of the continental shelf. The oceanic crust is much thinner, typically 8 km, and is poorer in silicon and aluminium than the continental crust. Another difference between continental and oceanic crust is their age: the continental crust tends to be much older. Some crust, such as much of the eastern Mediterranean sea, is intermediate in properties.

Below the first few metres temperature increases with depth in the earth by about 3°C per 100 m. This is because heat is continually being produced in the mantle and crust by the decay of the naturally-occurring radioactive elements potassium, uranium and thorium. Rocks conduct heat very poorly, so heat escapes from the mantle by convective movements of rock. These movements are transmitted through to the surface of the earth, where indirectly they produce many of the large-scale structures of the earth (see next section).

Plate tectonics

Some time ago it was observed that certain widely-separated coastlines could be fitted together, for example the east coast of South America and the west coast of Africa. These observations were incorporated into the theory of continental drift, according to which various supercontinents had existed in the past, to be split apart and amalgamated again many times. These ideas were not widely accepted, partly because of the lack of a convincing mechanism for the process. Geological and geophysical exploration of the ocean floors in the 1950s and 1960s produced the first evidence that these regions are drastically different from the continents, both in their relative youth and their composition. This proved to be the key to the problem and led to the modern theory of plate tectonics.

In this theory the lithosphere of the earth is divided up into about twenty major rigid 'plates', all in motion with respect to adjoining plates, at speeds of several centimetres per year. The upper part of each plate may include both continental and oceanic crust, but the plate extends downwards into the mantle to a depth of 200-400 km. Here the plates glide over the underlying rock, lubricated by a small amount of molten rock (magma).

This theory contrasts with that of continental drift in that the continents do not move on their own, but are merely carried along by the underlying plates. This suggests that geological processes will be concentrated along plate margins rather than continental margins. The driving force for plate motions is convective movements of the mantle as it tries to rid itself of internally-produced heat.

Although many plates contain both oceanic and continental crust, these two parts are not created and destroyed in the same way: new

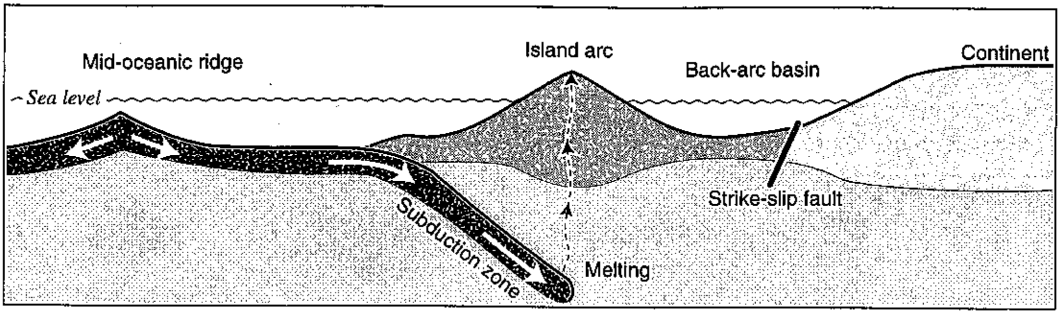


Fig. 1.4. Plate tectonics (not to scale). New oceanic crust is generated at a ridge in the oceans and eventually goes down a subduction zone beneath another plate. Melting of the plate and the overlying mantle produces magmas that rise up to form the volcanos of an island arc. Spreading behind an island arc produces a basin, commonly with some volcanism.

plate is always oceanic (Fig. 1.4). It forms along the margin and moves away in opposite directions at rates of several centimetres per year. Currently no new plate is being formed in the Mediterranean region. Oceanic portions of a plate are consumed in subduction zones. Here the oceanic plate dives underneath the adjacent plate and descends into the earth to a depth of up to 600 km. During the descent parts of the plate and the overlying rocks melt and the liquid rock rises towards the surface. It may solidify before it reaches the surface or it may erupt onto the surface to form volcanos. Where the opposing plate is oceanic the magmatic activity will produce a chain of islands, usually in the form of an arc, such as the Japanese islands, or, less typically, the volcanic islands of the Aegean. Buckling of the crust in front of the volcanic arc may produce a chain of non-volcanic islands, such as Crete, Rhodes etc. Diffuse extension of the crust behind the volcanic chain may produce a basin intermediate between true oceanic and continental crust.

It is not possible for continental crust to descend into a subduction zone as it is too light: any attempt will clog it up and throw up a mountain range. When a small block of continental material, possibly with islands, collides with a subduction zone, a mountain range is formed and the subduction zone shifts to the other side of the former block. During these collisions part of the ocean floor, especially that in the basins behind volcanic arcs, may not go down the subduction zone, but be thrust up onto the land. These slices of sea-floor frequently include peridotite from the uppermost mantle, which is commonly metamorphosed into serpentinite, a dark-green mottled rock named for its resemblance to snake-skin. Other characteristic components of the sea-floor are basalt lavas (see below) that have been erupted underwater as a series of pillow-like blobs, and the sedimentary rock chert. Such isolated remains of former oceanic floor are sufficiently common to have been given a name – ophiolite suite rocks (from the Greek for snake – *ophis*).

Plates may also slide past one another along major faults that can cross both continents and oceans. Here plates are neither created or destroyed. In the Aegean region the North Anatolian fault zone is of this type: the European continent is moving to the right with respect to the Anatolian plate. Plate boundaries may also be transitional between these cases, and their character can change along their length.

Local tectonics

The subject of plate tectonics attempts to explain the large-scale structures on the earth's surface; the detailed structures are the domain of local tectonics. Broadly, two different regimes occur: overall compression and extension (Fig. 1.5).

Compression of the crust must be accommo-

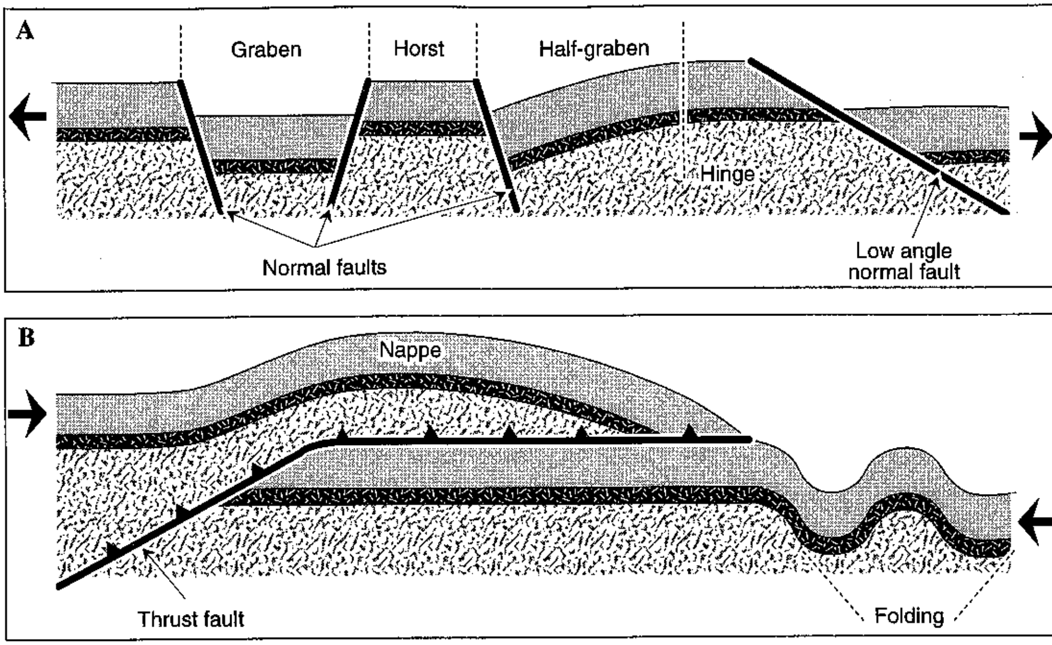


Fig. 1.5. Geological structures produced (A) during extension, (B) during compression. Faults can be reactivated and move in a different direction. Folding tends to occur at greater depths than faulting.

dated by folding of the rock strata or by faulting, i.e. movement along cracks in the rock. Which mechanism dominates depends on the strength of the rock and the depth at which the compression is occurring: at great depths the rocks are weaker and will tend to fold rather than fracture. The faults must accommodate shortening, so the upper block moves over the lower part, thickening the pile of rock. Such faults are commonly almost horizontal, and are called thrust faults. The rock unit bounded by thrust faults is called a nappe.

At the boundary between two colliding continents (see above) the crust is partly folded and partly cut up into nappes, which are piled on top of one another to form a thickened crust. The crust 'floats' on the underlying, denser mantle, hence a thicker crust will ride higher and more will project above the sea as mountains.

Nappes can be very variable in thickness and rock components. For example, some nappes are dominated by ancient sea-floor (ophiolites), whereas others are dominated by limestones deposited in shallow water. Large nappes or groups of nappes with similar rocks and geological histories are called isopic zones, and those of the Aegean are described in Chapter 2.

Extension of the crust can produce a single new plate boundary or it can be distributed over a wide region, as in the Aegean. Extension is accommodated near the surface by faulting. In this case the upper block drops down with respect to the lower block, and the mechanism is termed normal faulting. If the faults are steep then extension will produce tectonic valleys called grabens, separated by mountain blocks called horsts. Very commonly there is a fault on only one side of the graben and the floor has hinged downwards, to create a half-graben. In some areas the normal faults are almost flat-lying, resembling thrust faults, and do not produce grabens and horsts.

Earthquakes

Earthquakes are vibrations of the earth that are produced when rocks juxtaposed across a fault slide past each other. Some faults slide very easily so stress on the fault is dissipated continually. However, many faults are 'sticky' and the stress can build up to much higher levels before the energy is released in an earthquake.

The magnitude of an earthquake (the well-known Richter scale) is not a measure of destructiveness but merely indicates how much energy is released. It varies according to the amount of movement of the rocks (from a few centimetres to several metres) and the surface area of the fault that moved (from a few square metres to tens of square kilometres).

The amount of vibration felt at a particular place on the earth, and hence the local destructiveness of an earthquake, is called the intensity. It is measured on the Modified Mercalli scale and varies with location. Each earthquake therefore has only one magnitude but many intensities. The intensity depends on the distance to the centre of the earthquake, its depth and the local geology. Many sediments, especially those in recently drained lakes, can amplify considerably the earthquake vibrations and increase the intensity.

Joints

Most fractures in rocks do not have any significant movement of the two sides: these fractures are called joints, and several different sets can be seen in most outcrops. Some joints are formed by the same regional stresses that form the faults, others by contraction during cooling of igneous and metamorphic rocks. However, many form as a result of stresses produced during erosion or quarrying: at depth in the earth the stresses are more or less equal in all directions. When material is removed stress in the vertical direction is reduced, but the horizontal stresses are unchanged. Therefore, expansion can only be accommodated by upward movement, and joints parallel to the earth's surface will form. This effect can aid quarrying operations, as large blocks will naturally split off horizontally after a trench has been excavated around the block. It can also cause blocks to split during excavation.

The basic rock cycle

Rocks are naturally occurring materials made up of distinctive components, minerals, that have a limited range of composition and struc-

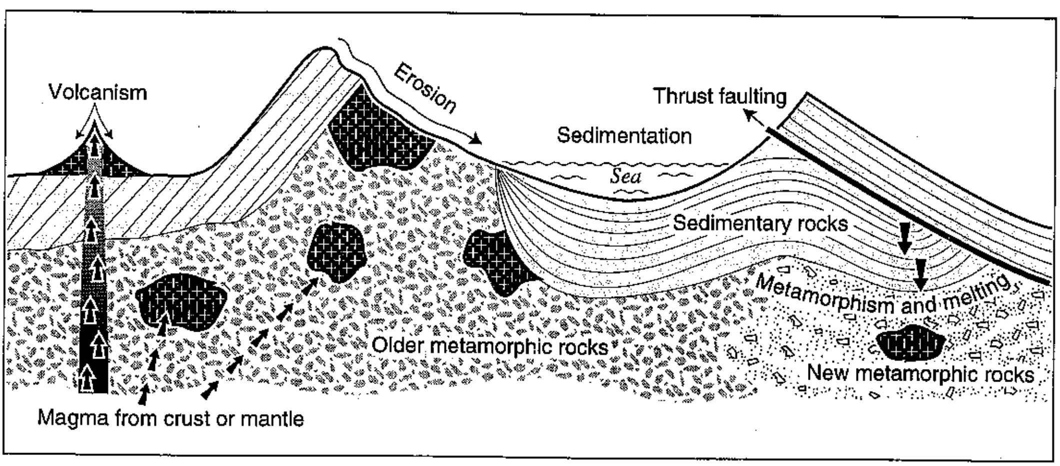


Fig. 1.6. Simplified rock cycle. New igneous rocks crystallise from magma produced by melting of the crust or mantle. They are eroded, along with existing rocks, and the sediments are commonly deposited in the sea or lakes. Thrust faulting thickens the crust, forming mountains and forcing the rocks downwards into regions of higher temperature and pressure, where they are transformed into metamorphic rocks.

ture. Rocks can be divided up into those that were at one time molten, termed igneous, those that formed on the surface of the earth, termed sedimentary, and existing rocks that have been transformed by heat and pressure, termed metamorphic. The interchange between these three groups of rocks is illustrated in Fig. 1.6.

Igneous rocks

All igneous rocks have crystallised from molten rock, termed magma, formed in the interior of the earth. Magmas poor in silicon and rich in iron and magnesium, such as basalt, form at depths of 100-200 km by the melting of rocks in the mantle. Magmas rich in silicon, such as granite, commonly form at much shallower levels in the crust by the melting of sedimentary, metamorphic or existing igneous rocks. In each case only part of the original rock melts, so that the magma has a different composition from its source.

The names of different igneous rocks reflect both their composition and the level at which they crystallised: slow cooling of magma below the surface of the earth produces rocks with mineral crystals that are readily visible to the unaided eye. Such rocks are termed plutonic (after Pluto, the Greek god of the underworld) and an example is granite. Faster cooling on the surface of the earth produces volcanic rocks (after Vulcan, the Roman smith god), which are generally fine-grained. These rocks may contain large crystals, called phenocrysts, but these formed at depth before the magma was erupted.

Volcanic rocks are further subdivided into lavas and pyroclastic rocks. Lavas were still liquid when they were extruded, generally quiescently, onto the surface. Pyroclastic rocks are produced during explosive eruptions. Liquid magma is blasted into the air where it solidifies before it hits the ground. The finer-grained material is called volcanic ash or tuff (not to be confused with tufa, a soft limestone). Volcanic breccias include blocks of pre-existing rocks and pumice, a solidified volcanic foam.

Sedimentary rocks

Sedimentary rocks are produced on the surface of the earth by accumulation of rock fragments and biological materials or by crystallisation of minerals from water.

Many sedimentary rocks, such as clay, shale (the term schist is used in some older texts), siltstone, sandstone and conglomerate, are accumulations of rock fragments, and are termed clastic rocks. Weathering breaks down existing rocks into smaller pieces (sediments), which are transported by wind or more commonly water. These sediments accumulate in dunes, river valleys, lakes or most commonly in the sea. Crystallisation of calcite, and other minerals, from fluids percolating through the loose material cements the grains together. Harder rocks are produced by more cementation. This process can continue even after the rock is quarried: as it dries out further crystallisation of the interstitial liquid will strengthen the cement.

Two series of clastic rocks have been given special names that reflect the environment in which they form. Flysch is a series of rocks deposited in the deep sea adjacent to a rapidly rising mountain chain. It is dominated by sandstone, with finer siltstone and clay. Molasse is another type of clastic rock associated with mountain chains. However, it is produced after the mountains have formed and it is deposited in shallow lakes or river beds. It is dominated by conglomerates, with sandstone and siltstone.

The term clay applies to both a group of platy minerals, and a rock dominated by these minerals.[179, 116] As mentioned above, clay can form by weathering, but it can also be produced by the alteration of rocks in contact with hot water (200-400°C) associated with volcanism or the emplacement of intrusions. The amount of clay minerals is increased when the rock is eroded, transported by water and redeposited.

Most clays are mineralogically unsuitable for making pottery as they are too stiff or they do not fire well. However, suitable clays are commonly associated with layers of lignite, a low-grade coal.[283] The original environment was a shallow lake in which fine clay accumulated. Eventually the water was sufficiently shallow for plants to grow abundantly. The plants died but did not decay in the stagnant water and their remains were transformed into lignite. Red pottery clays were also obtained from 'terra rossa' soils (see below) by washing and settling of the liquid.

Limestone is a sedimentary rock formed almost entirely of the mineral calcite and can form in a number of ways: many limestones are accumulations of marine shells, some microscopic in size. These rocks are generally formed in shallow seas, but can also form on the ocean floor, or even accumulate in sand dunes. Under certain conditions calcite (or the similar mineral aragonite) can crystallise directly from the sea to produce accumulations of calcite-mud, which is then cemented into solid rock. Finally limestone can be precipitated from freshwater in caves and around springs to form stalagmites, stalactites and travertine (tufa). Part of the calcium in calcite may be replaced by magnesium to form dolomite, a rock very similar in the field to limestone, but slightly more resistant to erosion. Marls are soft, impure limestones with large amounts of clay and sand.

Some organisms, such as sponges, have a skeleton made of silica. These skeletons can recrystallise to form the hard rock chert (flint), which commonly occurs as nodules or layers in limestone.

Prolonged evaporation of seawater in enclosed basins or irregularly flooded coastal flats can produce sedimentary rocks called evaporites. The first mineral to crystallise from the brine is gypsum, followed by common salt (sodium chloride) and then potassium and magnesium salts. Evaporites are too soluble to be generally observed on the surface, except for gypsum rock (alabaster), and even this is weathered very rapidly.

Fossils are an important part of many sedimentary rocks. Fossils are the remains of animals and plants or of their activities, such as worm tubes, tracks or faeces. Some fossils, such as shells, are essentially unchanged since formation. Others, however, have been chemically or physically changed during fossilisation: carbon, silica or calcite may replace organic matter or the organism may rot away and the hole become filled with fine-grained sediments.

Metamorphism and metamorphic rocks

When rocks formed under a certain pressure and temperature are moved to a different part of the earth with a different pressure and temperature, either higher or lower, then some of the minerals in the original rock may become unstable and new minerals will be formed. In some cases the only change is the recrystallisation of minerals to form larger grains. This is the process of metamorphism, and it can act on sedimentary, igneous and already metamorphosed rocks. Many metamorphic rocks have a distinctive layered structure, produced by deformation of the rock when it is weakened by the metamorphism.

Metamorphism of clay-bearing sedimentary rocks initially produces a platy mineral, mica, which gives a sheen to the rock, termed phyllite. With increasing pressure and temperature the size of the mica crystals increases and their good cleavage gives a fissility, or ability to be split into sheets, to the rock, which is now called schist, from the Greek *schizein*, to split. Mica is not stable at higher temperatures and pressures and is replaced by other minerals, such as amphibole. The resulting rock, which does not split very well, is termed gneiss (pronounced 'nice').

Rocks rich in iron and magnesium, such as basalt and gabbro, are metamorphosed into dark-green to black amphibolites, i.e. rocks made up of amphiboles. Metamorphism of peridotite at low temperatures yields a rock rich in serpentine minerals called serpentinite.

Very high temperature metamorphism can produce melting. If the magma is retained in the rock then a streaky rock called migmatite is produced. At higher degrees of melting so much magma is produced that it can not be retained in the rock and will start to separate. When it crystallises it will form an igneous rock, commonly a granite.

A 'typical' major volcanic eruption

Although every volcanic eruption is different in detail, many of the large explosive eruptions that have occurred in our region have similar features. The eruption begins with a plinian phase, named from the eruption of Vesuvius in AD 79, as observed by the Roman writer Pliny. An initial explosion, or possibly a landslide, opens up a pathway for the magma from a temporary storage chamber (magma chamber), at a depth of several kilometres, to the surface. The first magma to use this pathway is from the top of the magma chamber, where it contains much dissolved water. The water boils as the pressure is reduced and produces a light froth which accelerates upwards, fragments and leaves the volcano as a high-speed jet of ash. This rises into the atmosphere, sometimes to a height of 30 km, and takes the form of an umbrella-shaped cloud. Deposits from this stage of the eruption are called ash-fall tuffs. Their thickness will depend only on the distance from the vent and the prevailing wind direction and will not be affected by local topography.

As the eruption continues, water in the magma chamber is reduced and there is less force in the jet. Eventually the jet is unable to support the weight of the ash column and collapses. When the collapsing column hits the top of the volcano the ash has nowhere to go but sideways. The result is a strong lateral blast of ash and gases in all directions, termed a base-surge, and it can occur several times during an eruption. A base-surge is so fast that it is not deflected by topography, such as valleys, and will deposit equal amounts of material on both the valley floors and the ridges.

During the final stages of the eruption the magma does not normally have sufficient power to overflow the rim of the crater. Periodically, the activity increases and a hot mixture of fine ash and gas may roll down the side of the volcano. This mixture behaves like a liquid and will generally flow down valleys like a snow avalanche, leaving deposits that are thicker on the valley floors than on the sides. If the flow is very hot then it may glow, and the particles may weld themselves together to give a hard, dense rock called ignimbrite. This type of flow is termed an ash flow or nuée ardente (glowing cloud).

Lava flows are generally produced by magmas with smaller amounts of water. This may arise either because they have lost it in earlier eruptions or because they are naturally poor in water, as is basalt, for example.

Weathering and soils

When rocks are exposed to rain, frost and sun many of their minerals are no longer stable, and new minerals form, a process called weathering. Part of this process is physical, as cracks develop due to thermal expansion and contraction; part is chemical, reactions between the minerals and the air and water; and part is biological, caused by the growth of roots, bacteria, fungi and lichens. Generally minerals formed at high temperatures, such as pyroxenes, olivine and feldspars, are replaced by minerals stable on the surface, such as quartz, iron oxides and clay.

Soils are one of the materials produced by weathering. The nature of the soil and its thickness depend on the nature of the original rock, the climate and the time available before the soil is removed by erosion.[47, 30, 84] Stony or thin soils are produced where there is rapid erosion of weathering products, such as on steep hillsides where there is little vegetation. Such erosion may have occurred in the distant past, or it may be related to recent climate change or human activities. Red and brown Mediterranean soils cover large areas of Greece and western Turkey. These fertile soils were probably produced during the last glacial period, when the climate was wetter and cooler. They can be produced from a variety of different rocks. Soils in recent volcanic areas are usually very thin as there is not sufficient time for the breakdown of the rocks. However, where the streams flow into closed basins that have little or no access to the sea, the sediment may stay there long enough for chemical weathering to produce soil. Old volcanic areas may have deep fertile soils. Weathering of pure limestone or marble produces a thin red soil called terra rossa. This soil is produced by weathering of the impurities in the original rock and is rich in iron oxides (hence the colour) and clay. Although fertile, it is usually too thin on the hillsides for successful agriculture. One of the best soils in the Aegean region, known as rendzina, is produced on marl and flysch. However, the most productive soils are alluvial soils, formed in river valleys and former lakes. The amount of these soils available for agriculture has been greatly increased by drainage programmes, both recent and ancient, but they are also lost to urban development.

Soils in dry regions commonly develop a hard crust rich in calcite just below the surface. This crust is a type of soft limestone and is commonly called caliche or calcrete (from its common resemblance to concrete). Caliche develops where evaporation exceeds rainfall. Under these circumstances groundwater from deeper levels percolates upwards and evaporates below the surface. Small amounts of calcite in solution in the water are deposited at this level and form a harder layer. Subsequent erosion by rain or wind may remove the overlying softer soils and leave a pavement of 'fossil' caliche. Some of the rock locally termed 'poros' is caliche, but this term also applies to rocks with very different origins.

Extreme weathering of limestones under tropical conditions (in the Tertiary period in the Aegean) gives bauxite, a loose, red rock that is the ore of aluminium. In a similar way weathering of peridotite or gabbro can produce a loose, brown rock called laterite, used in antiquity as an ore of iron, and now also as an ore of nickel.

Groundwater

Rainwater or melting snow either flows along the surface into rivers and lakes, or is absorbed into the underlying rocks, where it is called groundwater (Fig. 1.7). Water can move through permeable rocks such as sandstone, or along cracks in less permeable rocks such as granite or limestone. Clay is impermeable as the material is weak and will reseal any cracks which could transport water.

The water descends until it reaches the water-table, which is the upper surface of the zone of rock completely saturated in water. This is the level of water in wells which are not being pumped. The water-table may be quite close to the surface, especially in valleys, where

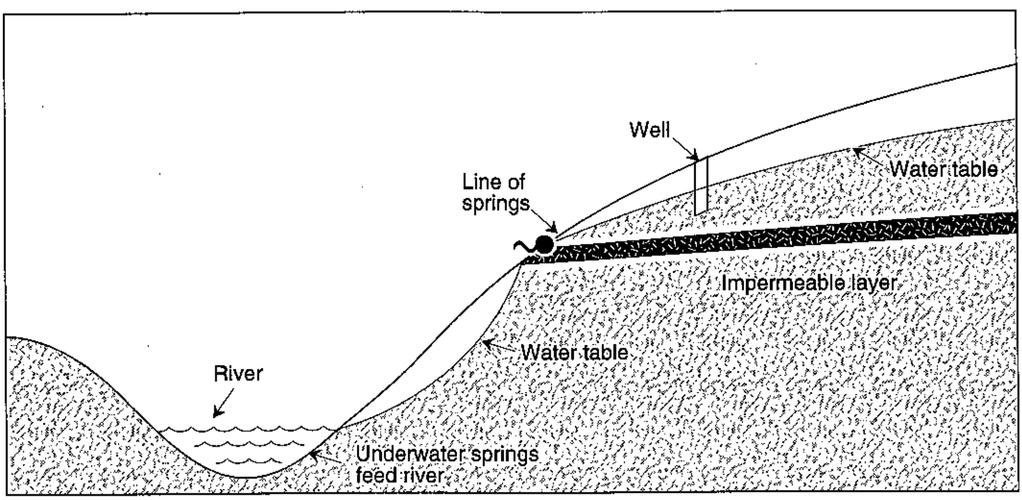


Fig. 1.7. Groundwater cycle. Some rainwater falling on land runs off along the surface and some percolates into the rock, where it descends to the water-table, the zone of permanently saturated rock. Loss of vegetation can increase the amount of run-off, and hence decrease the amount of water absorbed and available for springs. Springs form where the water-table intersects the surface, such as above an impermeable layer. Most springs debouche underwater and directly feed rivers.

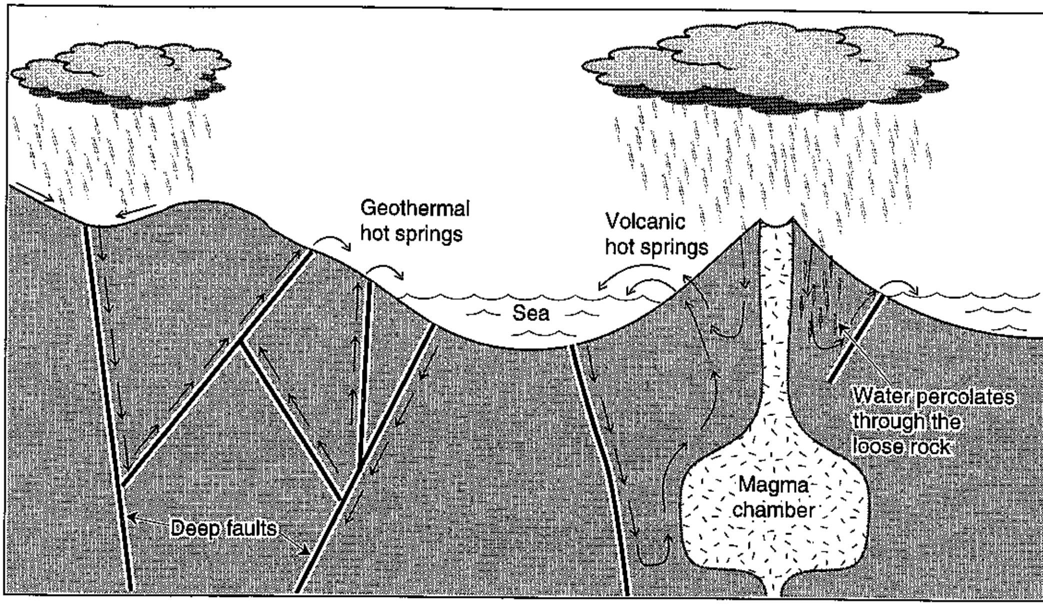


Fig. 1.8. Geothermal springs. The source of heat for most hot springs is the natural increase in temperature with depth of about 3°C per 100 m. For some springs in volcanic areas the heat source is cooling magma. Seawater or freshwater descends one fault or percolates through loose rocks, is heated and returns to the surface as hot springs.

it can be tapped with wells. However, in limestone or marble the water-table may be very deep (see below). Water can sometimes be perched on top of layers of less permeable rock, such as clay.

Where the water-table intersects the surface groundwater will rise as springs (but for karst springs see below). Although we usually think of springs discharging on land, most debouche directly into river beds. The passage of water through rock can be interrupted by an impermeable or less permeable layer of rock. This can lead to a line of springs at the boundary between the two rocks.

Hot springs

Most hot springs are related to deep faults, and not to volcanism, as is commonly supposed (Fig. 1.8).[1, 180] As mentioned above, temperature increases with depth everywhere in the earth. Two intersecting faults are necessary to tap this source of heat. Cool surface water (seawater, rivers or lakes) descends one fault to a depth of several kilometres. There the water is heated by the surrounding rocks, expands and starts to rise up the other fault, creating a convection system.

These hot waters can dissolve parts of the surrounding rocks, especially limestone and marble. As the liquids rise to the surface their temperature falls and minerals, commonly calcite, crystallise in the upper parts of the faults and around springs to form travertine (but note that travertine can also form around cold springs). These deposits tend to clog up the faults, and consequently in many areas periodic movements of the faults are necessary to maintain the flow of hot water.

The glacial period and recent climatic change

The climate of the last 1.6 million years, the Pleistocene and Holocene periods, has been marked by dramatic fluctuations. Within this period, loosely called the Ice Age, the climate has been generally cool, with warmer conditions typically lasting for about 10,000 years every 100,000 years. During the cooler periods temperatures were about 5-8°C lower, the glaciers advanced and covered much of northern Europe, and the climate of the Mediterranean region was generally wetter.

We are now in one of the warmer intervals, called the Holocene or Recent period, which started about 10,000 years ago.[235] It began with rapid warming, but about 8,000 years ago the climate became more stable. Since then climatic variation has been less extreme, especially during the last 4,000 years.[235, 237]

Sea-level

Although the sea is often viewed as unchanging, its level with respect to the land was often very different in the past (Fig. 1.9).[261] These differences reflect both global changes in the level of the sea and local changes in the height of the land, and the history of sea-level changes is therefore different at each location.

During the cooler periods of the last few million years much water was stored on the continents as ice-sheets, lowering sea-level by as much as 120 m (Fig. 1.9). The last glacial maximum was about 20,000 years ago. This water was released as the climate warmed up, returning the sea-level to its earlier condition. Since then this mechanism has had little effect on sea-level.

Local changes in the height of the land are commonly related to large-scale tectonic movements, commonly along faults. The Aegean is the most active region of the Mediterranean, and hence tectonic control on ancient sea-level is the most important process.[85]

Evidence of ancient sea-level stands (periods of static sea-level with respect to the land) can be found in coastal landforms, in the occurrence of recent marine fossils on land and in archaeological remains. Erosion by the sea is strongest where the waves break on the shore, hence prolonged action of the waves produces a notch in the sea-cliffs and a surf bench (see Plate 1A). The absence of significant tides in most of the Mediterranean means that these effects are concentrated in a narrow zone, and lowering of sea-level will preserve these features from further erosion. The occurrence of marine fossils on ancient platforms serves to distinguish them from terraces produced by faulting, and can be used to determine the age of the sea-level stand. Archaeological remains that were once at sea-level can also be used to determine the height and age of ancient sea-levels.[256]

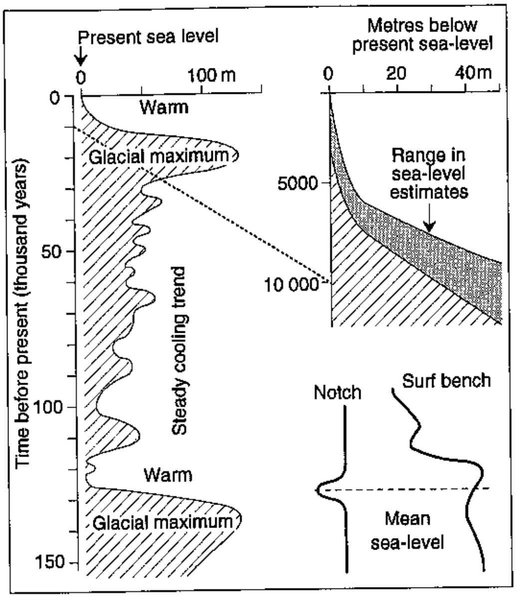


Fig. 1.9. Global sea-level variations during the last 150,000 years have been controlled by global climate: during cold spells water was tied up in glaciers and sea-level dropped. Estimates of recent global sea-level changes are complicated by local tectonic effects, so only a range in sea-level estimates is given. However, most estimates agree that there has been little global sea-level change during the last 5,000 years. When sea-level is maintained at one level for a period of time then a notch or surf-bench develops. In regions with little tide the roof of the notch is about 30-40 cm about sea-level.

Rivers and alluvium

Rivers are a major force in shaping the land. It is easy to underestimate their effect during the dry Mediterranean summer, but winter floods can rapidly reshape valleys. It is not only water that moves in a river, but also sediment, from clay particles to boulders. When deposited on land this material is called alluvium.

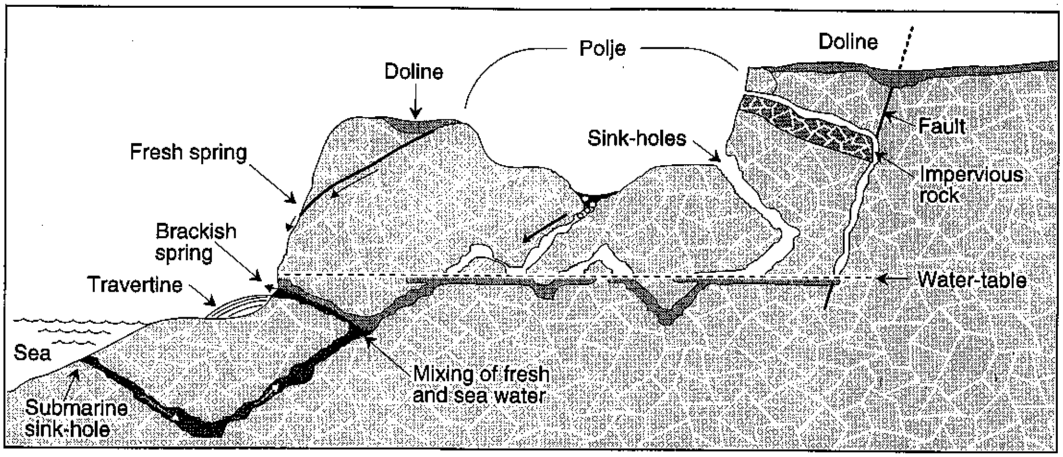


Fig. 1.10. Karst features. Rainwater or snow-melt is drained into dolines or sink-holes whence it flows underground, through fissures and caves, to reappear at karst springs.

The upper, steeper and straighter sections of rivers tend to erode their beds, and the eroded material is transported to the lower parts of the river where it is deposited to form alluvial plains. Rivers flowing across alluvial plains tend to follow a sinuous pattern, termed meandering after the river Meander (now the Büyük Menderes). Alas, this river now runs in a straight canal.

Rivers can cut down into old alluvial plains if the bed of the river is steepened by tectonic movements, a drop in sea-level or other causes. The old alluvial plain will be preserved as terraces on the sides of the valleys. Conversely, an increase in the amount of erosion in the upper parts of a river can rapidly bury old alluvial plains under new alluvium.

Limestone and marble (karst) landscapes

The ability of limestone and marble to dissolve in and crystallise from water has led to the development of a distinctive landscape called karst (Fig. 1.10), named after a region of Croatia where it was first described.[86] The gas carbon dioxide dissolves in water to produce a weak acid that can dissolve the calcite (and dolomite) in limestones and marbles. Calcite crystallises from these waters when the concentration of carbon dioxide in the water drops. In karst regions there are few streams or rivers and most rain or snow-melt sinks directly down into the rocks, to appear as springs, commonly far away.

We start with a small-scale surface feature of limestones and marbles: the ridges and grooves produced by weathering. A great variety of different types have been described, but the most distinctive are razor-sharp ridges produced by solution of the limestone in rainwater. The effect occurs immediately after the raindrop hits the rock. Smoother grooves are formed underneath soils and sediments.

On a larger scale closed, cone-shaped basins 2-100 m deep and 10-1,000 m in diameter, commonly floored by sediments and soils, make up an important part of the surface of most karst regions, and give it a pitted relief. These basins are called dolines and may be widely separated or occur as swarms. Rain and snow-melt drain into dolines and then into the underlying rocks, hence they take the place of valleys in non-karst areas. Dolines form by solution of the underlying limestone along cracks, and the settling down of the smaller blocks. Solution may be enhanced by higher levels of carbon dioxide in the soils, produced by plant decay.

Dolines absorb rainwater, but rivers and streams disappear into sink-holes (swallow-holes). These may just be areas of loose sediments where the water sinks away, rather like dolines, or they may be caves in a cliff or holes in the ground. Most sink-holes develop from dolines. Sink-holes generally lead into cave systems, but these may not always be accessible.

Valleys are not completely absent from karst areas. Major rivers that originate outside the region can cross in a valley. Blind valleys end when the river drops into a sink-hole. Some valleys are normally dry, but may contain water when flow exceeds the capacity of the sink-holes.

Flowing water may dissolve the walls of cracks and faults to produce caves. Once established, caves can also be enlarged by the abrasive action of sediments in the water. In the upper parts of caves water commonly flows only in the lower part of the passage, and processes similar to those seen in rivers on the surface, such as meandering, beaches etc. may occur there. In the deeper parts water completely fills the caves, like water in a pipe. Here the cave tends to have a different shape: it is commonly rounded in section and much straighter.

Caves are guided in direction by existing structures in the rocks, such as joints and faults. However, most caves have one or more prominent levels, corresponding to ancient water levels in the cave system that were maintained for a long period. There are many reasons why solution should be most active near the water level, but the dominant effects are flow speed and the time available for reactions: below the water surface flow is slower, so that dissolved material is not removed so fast. Above the water level most of the rocks are not continually immersed in water, hence there is less opportunity for solution.

Although karst caves are formed by solution it is the opposite process, crystallisation, that has decorated them, and is the main reason for their interest. In most caves calcite crystallises in response to loss of carbon dioxide rather than evaporation of the water. A slow drip rate promotes formation of stalactites from the roof, and a faster rate favours production of stalagmites on the floor. Continuous flow produces sheets of calcite on the walls, called flowstone, and curtains of rock from the roof. Seepage of water too slow to form drops, can produce weird shapes like springs, called helicites, and irregular branching forms 2-5 cm long, called cave coral. Water flowing along the floor also deposits calcite in areas where more carbon dioxide is lost. This occurs where water overflows a barrier, such as the edge of a pool. Deposition of calcite builds a small dam, and the pool deepens with time. Such processes can also occur on the surface, producing a series of pools and deposits of calcite called travertine. Most of the decorations in Greek caves formed when the region had a wetter climate, probably during the last glacial period.

Another distinctive feature of karst landscapes is the formation of large basins, known as poljes, surrounded by hills that slope quite steeply to a flat floor. No rivers or streams leave these basins, instead they are drained by sink-holes in the floor, or at the edge of the plain. Drainage is commonly inadequate so that many poljes are flooded in winter but dry in summer. Poljes commonly develop by differential erosion along contacts with less readily weathered rocks, or by tectonic movements along faults. Once they have started to form, the ponded water will dissolve away the limestone or marble of the floor, and the solution will be drained away into the sink-holes. This is why the floors are commonly so flat.

Water enters the rocks of a karst region at many points, but tends to leave at only a few major springs.[146] Some springs rise underneath alluvium, but others flow from an open cave or rise up a steep conduit to the surface. Loss of carbon dioxide from the waters as they leave the ground can lead to crystallisation of calcite and the production of travertine terraces. Some sink-holes can become spring outlets during heavy rains, especially those around poljes. Some springs debouche offshore in the sea.

Beach-rock

Many beaches in the Aegean have a gently shelving platform of well-cemented beach sand and gravel that resembles concrete (see Plate 1B). This is called beach-rock and it can form extremely rapidly, in 25 years or less. The sediments are commonly cemented by calcite, but the origin of the cement is not always clear.[115] Some may originate through evaporation of fresh groundwater in a process similar to that which produces caliche. It has also been suggested that the cement forms from evaporation of seawater, aided by small variations in the level of the sea or by mixing of sea and fresh water. In any case cementation takes place below the surface and the beach-rock is exposed by subsequent erosion.

2

Geological History of the Mediterranean

The Mediterranean has a very complex geological history, mostly spanning the last 200 million years. Interactions between the European/Asian and African continents have created and destroyed several seas, of which the Mediterranean is but the most recent. The geological complexity, the landscape and the climate of this region have long attracted geologists. The best compilation of their work, at least for the European side of the Mediterranean, is *The Geology of Europe.*[2] The geology of Greece and Turkey has also been treated in detail in two books,[127, 34] as have the mineral deposits.[172,24] Here we will give an outline of the geology of the Mediterranean, but will deal in more detail with the Aegean and adjacent areas. More details of the geology are given in the regional chapters 3-16.

The geological framework

The Mediterranean region has a complex internal geological history, but the external events that controlled the development of this region are simpler to explain, and we will start there. Our history begins 190 million years ago, in the Mid-Jurassic period. Although there are many rocks older than this in the region, we are uncertain of the position of the continents in earlier times, and hence cannot reconstruct the earlier history of plate tectonics.

The North Atlantic Ocean did not exist in the Early Jurassic period and both Africa (more correctly Gondwanaland: Africa with other continents) and Eurasia (Europe and Asia) were united with the North American continent to form a super-continent named Pangea (Fig. 2.1). Between the future continents of Africa and Eurasia there was a wedge-shaped ocean, named Tethys (after the daughter of the earth-goddess Gaia), which opened out to the east where it joined the other oceans. This state of affairs changed when the North Atlantic Ocean began to form about 190 million years ago.

A new plate margin started to form with the rifting of the continent along huge normal faults, following closely the position of an older ocean that had closed up to form the Appalachian, Caledonian and Scandinavian Mountains. New crust was created in the rift, forming a new ocean, the North Atlantic. Initially this rifting was concentrated in the south, separating Africa from North America, and moving it to the east. The position of continental rifting, and new ocean formation, moved slowly to the north, like a tear, and by about 110 million years ago Eurasia had started to separate from North America.

During the next period, the southern parts of the Atlantic Ocean continued to form more rapidly than those in the north, forcing the African plate to rotate in a clockwise direction about an axis in the Atlantic off Gibraltar, closing up the Tethys Ocean. This rotation slowed down considerably when the Arabian part of the African continent hit Eurasia, producing the mountains of Turkey, and has now almost ceased. Of course, within this framework there is much detail. For example, continental fragments originally attached to Africa, such as much of Italy and parts of Greece, have broken off, crossed the closing ocean and hit Eurasia. It is this part of the story that will be expanded below. But first we must deal with some of the special terms commonly used in Alpine and Aegean geology.

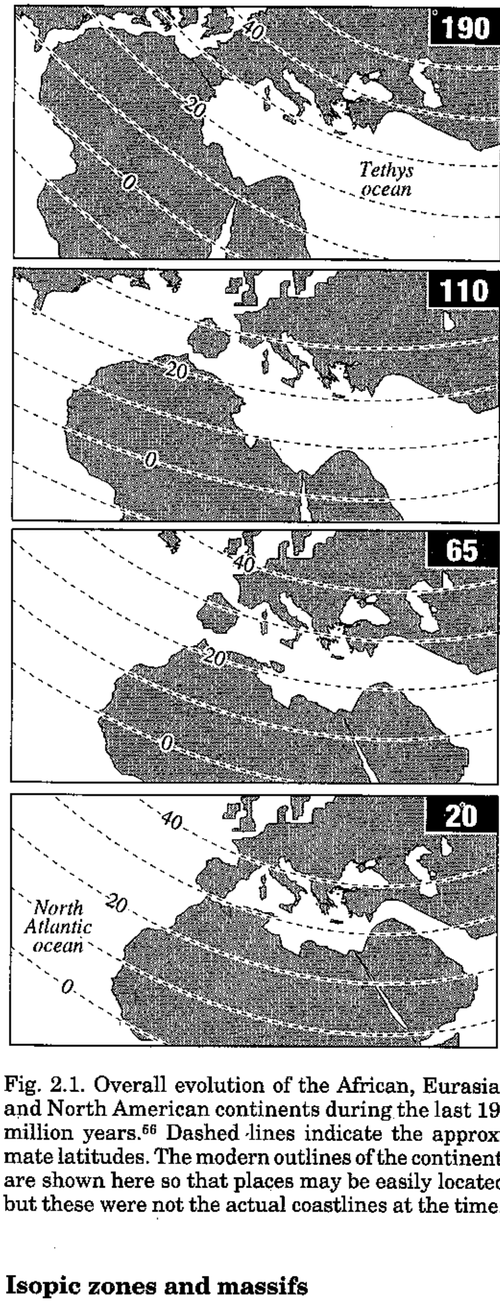


Fig. 2.1. Overall evolution of the African, Eurasian and North American continents during the last 190 million years.[66] Dashed lines indicate the approximate latitudes. The modern outlines of the continents are shown here so that places may be easily located, but these were not the actual coastlines at the time.

Isopic zones and massifs

The overall geology of the Alpine region has traditionally been described in terms of isopic zones and massifs. Isopic zones are groups of widespread rocks that share a common history, both in the ancient environments of deposition of sediments (deep ocean, shallow sea, continent, etc.) and their faulting and folding. They were originally continental fragments, islands, oceanic ridges or parts of the ocean floor. Isopic zones may be hundreds of kilometres long, and up to several kilometres thick. They are bounded by faults, commonly shallow-dipping thrust faults, formed during regional compression, and are hence nappes or groups of nappes. These compressions have also stacked the isopic zones up onto each other and against the massifs.

Massifs are blocks of metamorphic and plutonic rocks, formerly assumed to be much older and more resistant to folding and faulting than adjacent sediments. Nowadays the distinction between massifs and rocks of the other isopic zones is not so clear: some of the metamorphism is quite recent, and parts of the massifs may not be much older than adjacent sedimentary rocks. Massifs are better considered as slightly lower levels of the continental crust exposed by faulting or erosion.

The isopic zones and massifs of the Aegean were once a series of continents, continental margins, deep troughs and ocean basins (Figs. 2.2, 2.3, 2.4). The present geographic distribution of these zones is now very different as they were piled up on top of each other during the Alpine crustal compression (see later). The various groups of rocks will be described in order from the 'internal' zones of the north-east to the 'external' zones of the south-west.

The most extensive area of metamorphic and plutonic rocks in the Aegean region is the Rhodope massif in Greek Thrace and Bulgaria, and the adjacent Serbo-Macedonian massif to the west (Fig. 2.2). These massifs may continue under the Thrace basin into north-eastern Turkey, as the Sakarya zone (or Western Pontides). All these massifs have long and complex histories, which may stretch back into the Precambrian epoch over 600 million years ago. They were partly metamorphosed and faulted during the Alpine compressions. More recently granites have been emplaced and volcanic rocks erupted.

The Vardar (Axios) isopic zone lies to the west of the massifs, and continues northwards into the former Yugoslavia (Fig. 2.2). To the

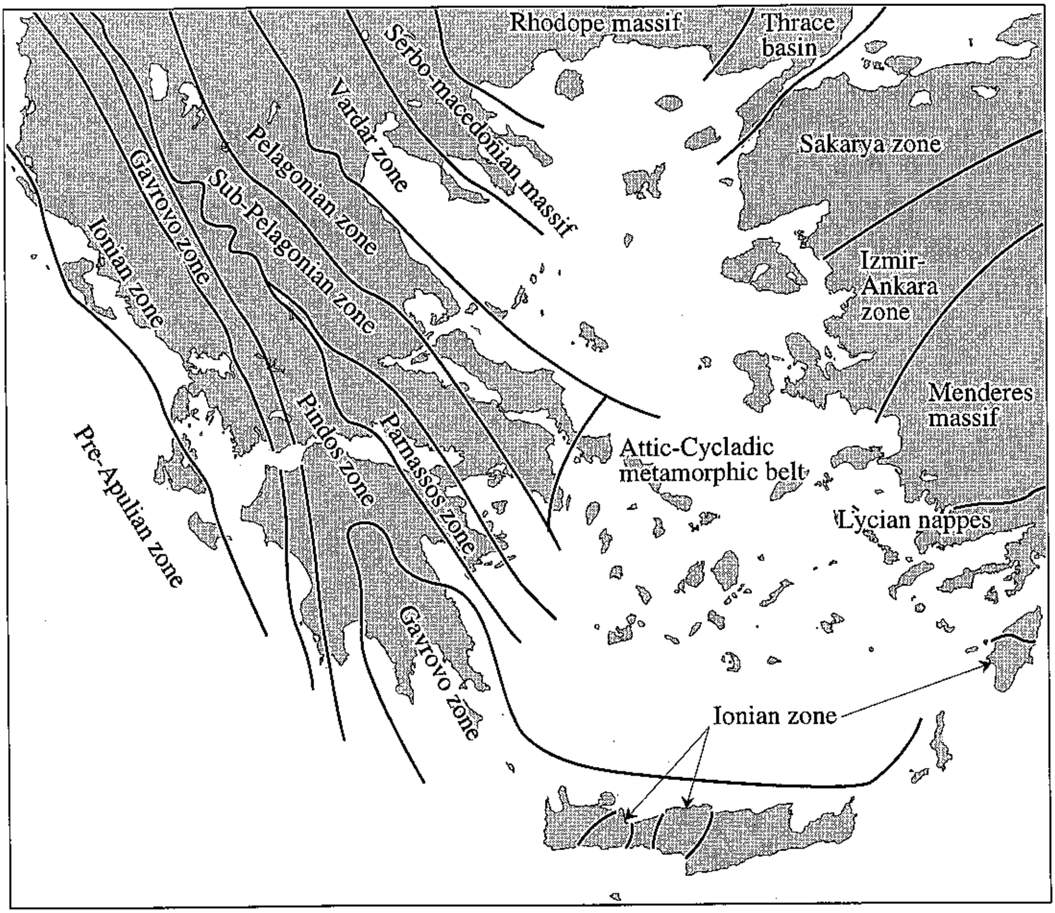


Fig. 2.2. Isopic zones and massifs of the Aegean region.

east it sweeps under the Aegean Sea, possibly to reappear in Chios. It may continue on the Turkish mainland as the Izmir-Ankara zone. It is a complex zone that has been sub-divided by some geologists into separate zones. However, it is dominated by Mesozoic deep-water sediments and ophiolites, and is hence an old ocean basin, part of Tethys. Within this basin were a number of older continental fragments, expressed as ridges.

The next isopic zone is the Pelagonian, which is exposed in much of eastern peninsular Greece. During the Triassic and Jurassic periods shallow-water limestones were deposited on this continental fragment, in an environment similar to that of the Bahamas today.

Some geologists consider that the Pelagonian zone continues to the south as the Attic-Cycladic metamorphic belt (or massif). The oldest rocks are schists, gneisses and granites some of which may be Palaeozoic. However, many of the rocks are similar in age to those of the adjacent zones, and some metamorphism was the result of the Alpine compressions.

The Menderes massif starts near the Aegean coast and extends far into central Turkey (Fig. 2.2). It is geologically similar to the Attic-Cycladic metamorphic belt but also shows some affinities with metamorphic rocks in Africa. It is not clear if it was separate from the Attic-Cycladic metamorphic belt, or merely a lateral extension.

The Sub-Pelagonian zone is the great belt of ophiolites and associated rocks (limestones, cherts) of the Aegean, and continues to the north into Albania (Fig. 2.2). It may continue to the south-east within the Lycian nappes of Turkey. These rocks were originally part of a continental margin, between the Pelagonian continent and the Pindos Ocean. The region was lifted up during the Late Jurassic and Early Cretaceous. After a period of emergence, reef limestones were deposited during the Late Cretaceous. During the Oligocene and Miocene a deep continental trough developed, which filled up with up to 5 km of continental and shallow-water sediments, called molasse, and deep water flysch sediments.

The Parnassos zone is only exposed in south-central Greece. For most of the Triassic to the Palaeocene this small continental fragment was under shallow water, and limestones were formed. When the region was dry land tropical weathering produced bauxite deposits. These rocks were followed by flysch, which terminated with conglomerates in the late Eocene, indicating nearby mountain-building.

The Pindos zone crops out in central Greece and the Peloponnese, but is much more extensive as it underlies or rode over much of the other zones. Its sediments were mostly deposited in a deep oceanic trough, but there are no ophiolites in this zone. During the early part of its history it was a true ocean basin.

The Gavrovo zone (in places called the Tripolitza zone) is narrow along the coast of Albania and in central Greece, but widens out in the Peloponnese. Like the Parnassos and Pelagonian zones, this region was a continental fragment for the early part of its history. Initially, during the Mesozoic, thick shallow-water limestones were deposited. However, in late Eocene times mountains had been formed and they shed flysch sediments onto the limestones, almost completely covering them up.

The Ionian zone comprises much of Epirus, and parts of the Ionian islands and Peloponnese. Like the Pindos zone, in early times this zone was an area of deep water, but probably floored by thin continental crust. Again it contains Mesozoic limestones deposited in a trough. As in the Gavrovo zone, and for the same reasons, flysch makes its appearance in the Eocene.

The most external (westerly) zone is the Pre-Apulian, and it barely touches our region. It comprises little deformed sediments of both deep and shallow water origin, and is really part of the Apulian platform of Italy. It is a block of continental crust.

To these zones and massifs we must add the Thrace basin of north-western Turkey. It is not a true isopic zone as it was formed after most of the Alpine compressions, in response to movements of the North Anatolian fault zone, which divides the Eurasian plate from the Turkish plate. A large thickness of sediment was deposited here from the Early Tertiary onwards, but recently sedimentation has ceased and depression of the crust has shifted southwards to the Sea of Marmara.

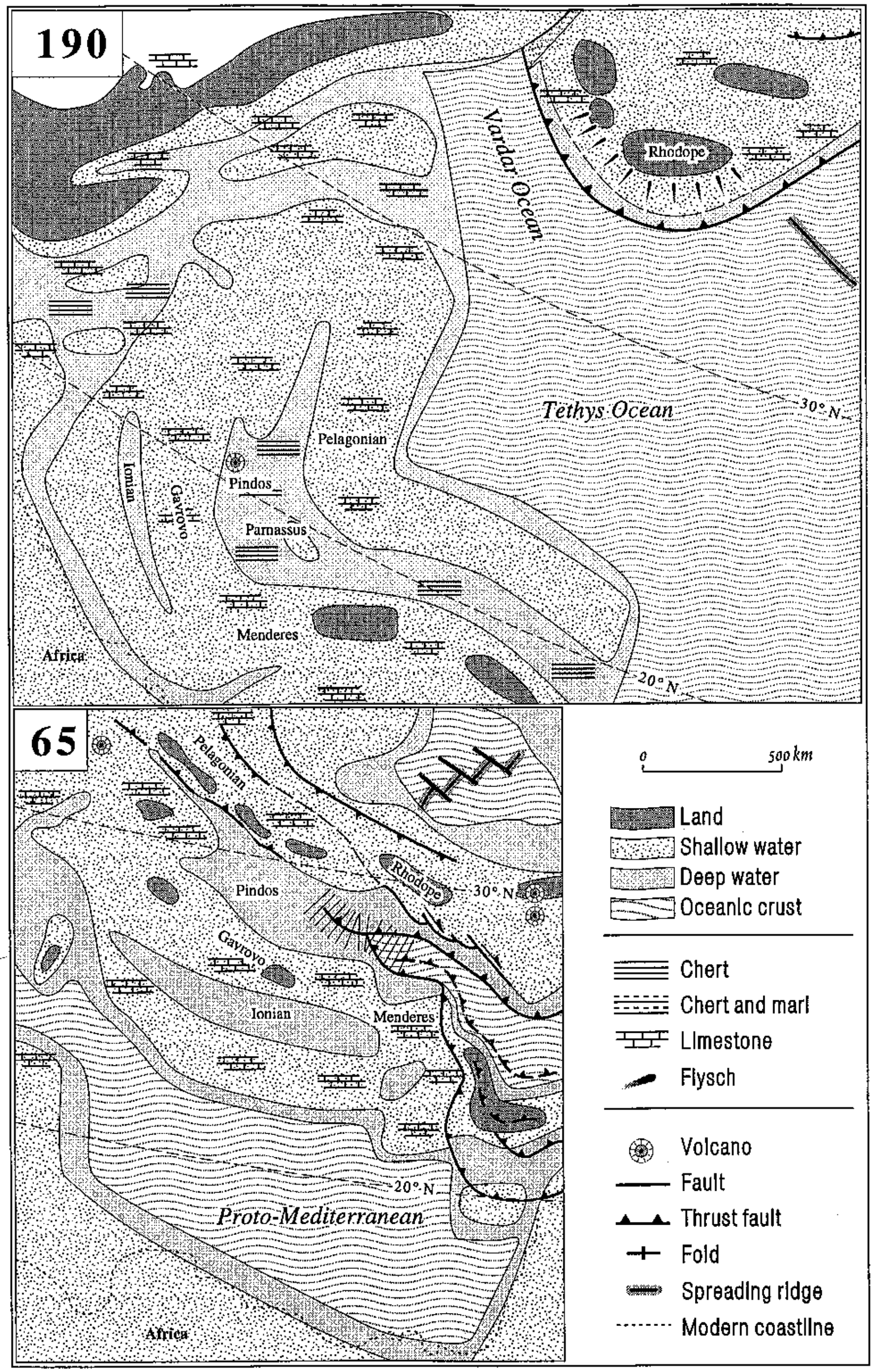
Thus overall the various zones and massifs represent an alternation of micro-continents, deep troughs on continental crust and true ocean basins (Fig. 2.4). Next we will trace the interactions between these units and their assembly into the present Aegean region.

The geological development of the Aegean

There have been many different reconstructions of the history of the Mediterranean region, but here we will follow that of Dercourt et al. [56] which is the most comprehensive. The overall history is that of the closure of an ocean basin filled with islands and underwater ridges and the thrusting of each component onto the adjacent parts generally to the north and east (Figs. 2.3, 2.4).

In the Early Jurassic, 190 million years ago, the Rhodope and Serbo-Macedonian massifs and the Sakarya zone lay along the north shore of the Tethys Ocean. These rocks are considered to be the southern margin of the Eurasian continent. To the south a subduction zone plunged northwards under the continent. All the other zones lay further south, adjacent to the African continent. Between was a wedge of true ocean, part of the Tethys Ocean.

The Pelagonian, Parnassos and Gavrovo zones were continental platforms, covered by


190
Vardar Ocean
Rhodope
Tethys Ocean
30° N
Pelagonian
Ionian
Gavrovo
Pindos
Parnassus
Menderes
Africa
20° N
65
Pelagonian
Rhodope
30° N
Pindos
Gavrovo
Ionian
Menderes
20° N
Proto-Mediterranean
Africa
0
500 km
Land
Shallow water
Deep water
Oceanic crust
Chert
Chert and marl
Limestone
Flysch
Volcano
Fault
Thrust fault
Fold
Spreading ridge
Modern coastline

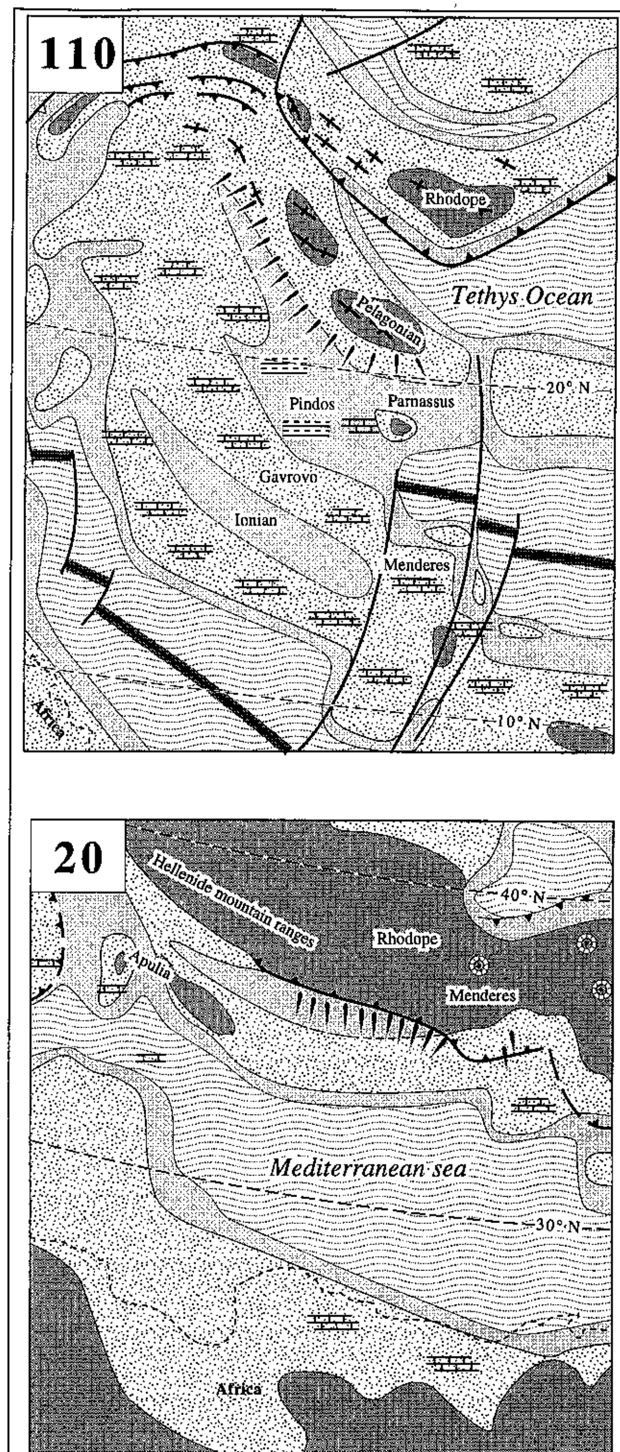


Fig. 2.3. Ancient positions of the isopic zones of the Aegean region for 190, 110, 65 and 20 million years ago.[56] Most of these zones, except the Rhodope massif, originally lay adjacent to the African continent. The abundant limestones of the Aegean region formed in these warm, mostly shallow, seas. Convergence of Africa and Eurasia forced the Tethys sea to close up, but at the same time a new ocean basin, the early Mediterranean, opened up to the south. Expansion of the basin separated the Aegean isopic zones from Africa and helped to force them against Eurasia.

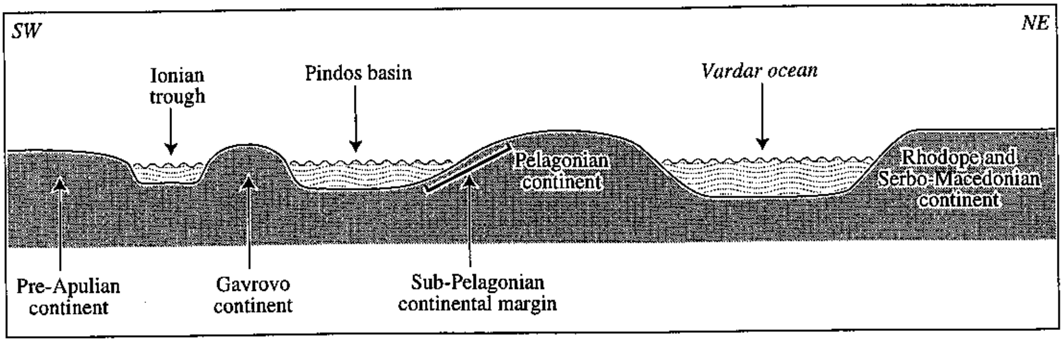


Fig. 2.4. Schematic section across the Aegean region before compression. The Pre-Apulian, Gavrovo and Pelagonian continental zones were covered with shallow seas for much of the time.

shallow, warm tropical seas, where limestones were deposited in an environment similar to that of the Bahamas today (Fig. 2.4). The Ionian zone was a deep trough on the African continental shelf and the Pindos basin was floored by true oceanic crust. The Sub-Pelagonian zone was the continental margin adjacent to this basin. This arrangement was relatively stable until the end of the Jurassic (130 million years ago) when parts of the Sub-Pelagonian continental margin and Pindos basin were thrust eastwards onto the Pelagonian continental platform.

A major change had occurred by the Mid-Cretaceous, about 110 million years ago; a new plate boundary had formed south of the Pre-Apulian zone and an ocean basin, the Proto-Mediterranean, was growing. This new ocean detached from Africa the zones that would later form the Aegean region and Apulia (southern Italy). As there was relatively little change in the positions of Eurasia and Africa, the Tethys Ocean basin to the north had to close up. It was partly subducted beneath Eurasia and partly piled up against the massifs as in the Vardar zone. As the Pelagonian zone neared Eurasia the regional compression was converted into uplift by movements along thrust faults. Parts of the Pelagonian zone above water shed flysch sediments into the Pindos Ocean basin to the south-west. Similarly the Parnassos continental fragment was raised above sea-level and intense tropical weathering produced bauxite and laterite deposits.

During the next period the Aegean isopic zones were forced to the north by expansion of the ocean to the south. Ocean basins were subducted and/or thrust westwards over the adjacent zones. Continental fragments were similarly stacked, one on top of the other. By 35 million years ago all that remained of the vast platform once attached to the African mainland were the Ionian and Pre-Apulian zones. Even the Menderes massif had been welded onto Eurasia. By 20 million years ago the Ionian zone had almost been completely consumed, and continental crust, albeit thinner than normal, linked Apulia and the Hellenide mountain ranges.

The overall process of regional compression has continued in the Ionian islands to this day, but elsewhere in the Aegean major regional expansion became important. This expansion has shaped much of the landscape of the Aegean and continues today.

Subduction of oceanic crust northwards beneath the Aegean started during the Miocene, or possibly a little earlier, initially along an east/west line (Fig. 2.5). The crust above the subduction zone arched upwards to form the non-volcanic Hellenic arc, now represented by the islands of Crete, Karpathos and Rhodes, the western edge of the Peloponnese, and south-eastern Turkey.[13] Further north melting of the subducted slab, and the overlying mantle, produced the volcanoes of the South Aegean volcanic arc and the more diffuse, but important, Early Miocene volcanic rocks in the western Aegean and Turkey. Volcanism con-

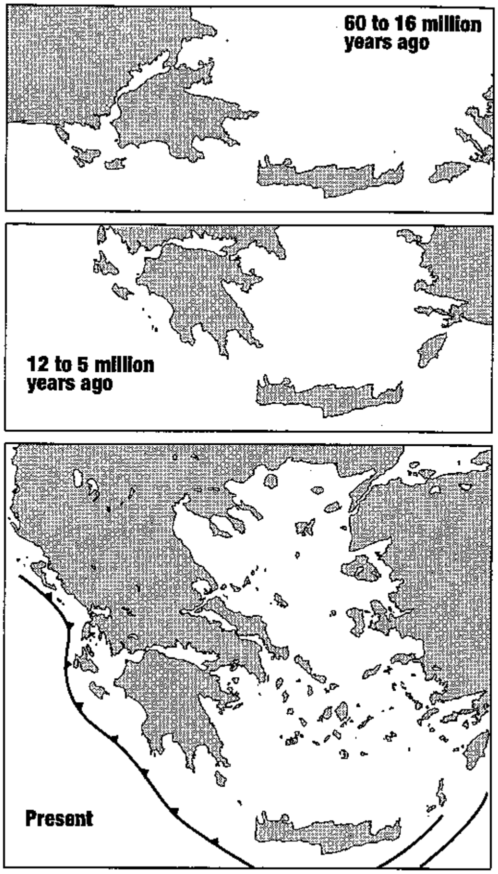


Fig. 2.5. Evolution of the Hellenic arc over the last 60 million years.[137] Initially the arc ran almost straight, east/west. The curvature developed by expansion of the crust north of the Aegean arc, forming the present Aegean Sea.

tinues now on the islands of Thera (Santorini) and Nisyros.

Crustal extension associated with the subduction was concentrated in the region north of Crete, above the subduction zone, so that the initially straight subduction zone was inflated into the curve we see today (Fig. 2.5).[137] This extension was coupled to overall plate motions, so that a large part of the region has experienced Neogene extension. At the surface the crust has responded to this extension by developing grabens and horsts. The orientation of these structures is rather variable, but many are approximately north-west/south-east, although some of the youngest faults in the Peloponnese and Crete are north/south. It is these grabens and horsts that have produced many of the mountains and valleys, as well as the islands.

Mention should be made of an unusual event that occurred about six million years ago: The Mediterranean sea almost completely dried up.[117] This happened when tectonic forces closed up the Straits of Gibraltar. Rivers flowing into the basin, principally from the Black Sea, were insufficient to maintain sea-level and the sea dried out, except for a few saline lakes like the modern Dead Sea. There may even have been a salty waterfall at Gibraltar, also feeding the lakes. Evaporation of water from the lakes produced deposits of gypsum and other minerals.

The present plate tectonic configuration of the Aegean region is rather complex and the details are much disputed (Fig. 2.6, Plate 2).[260] The African plate is moving north-east with respect to Europe and descending along a subduction zone underneath the Aegean plate. To the east the plate motions are almost parallel to the plate boundary, hence the relative motion is taken up by strike-slip faulting. Still further east, towards Cyprus, the subduction zone resumes. In the north the Anatolian plate that carries much of Turkey is moving westwards with respect to Europe, and most of this motion is taken up by a strike-slip fault, the North Anatolian Fault zone. This fault runs across the northern Aegean Sea, as the North Aegean Trough, and fades out somewhere near Volos. The motion is also taken up by a number of parallel faults to the south in western Turkey and the Aegean. There are no simple plate boundaries within peninsular Greece to link up the major structures; motions are shared between a number of smaller faults of all types.

Geologically recent events

During the last cool interval of the glacial period, about 130,000 to 20,000 years ago, the climate of the Aegean was moist and cool. Glaciers were restricted to the highest mountains on the Greek and Turkish mainlands, and the island of Crete. Sea-level was considerably

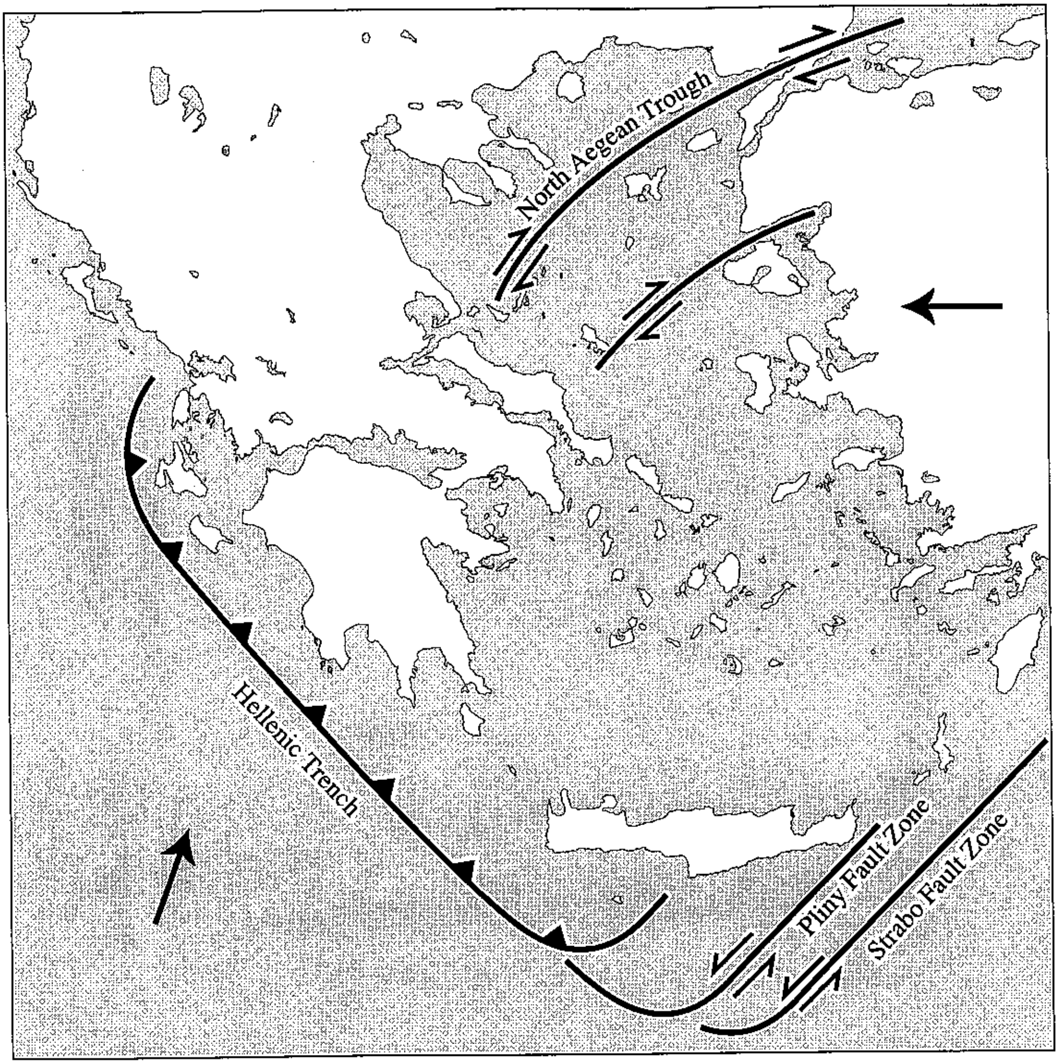


Fig. 2.6. The present-day plate tectonics of the region (after 260).

lower that it is today, and hence the river-beds were steeper and there was rapid erosion. Mountains were deeply incised and broad, flat valleys were developed. It is this landscape that we see in Greece today, modified by post-glacial effects. Many of the red and brown soils were produced during this period.

At the end of the last glacial period the warming was accompanied by a rapid increase in sea-level. The sea invaded the valleys, forming deep inlets, which were rapidly filled in by sediments transported by the rivers. The alluvial sediments that fill most Mediterranean valleys have been divided into the 'older' and 'younger' fills, originally thought to have been two synchronous events.[277] However, it is clear that in many places there are several different younger fills, which are not synchronous. The older fill was produced by erosion associated with the post-glacial warming, but the cause of

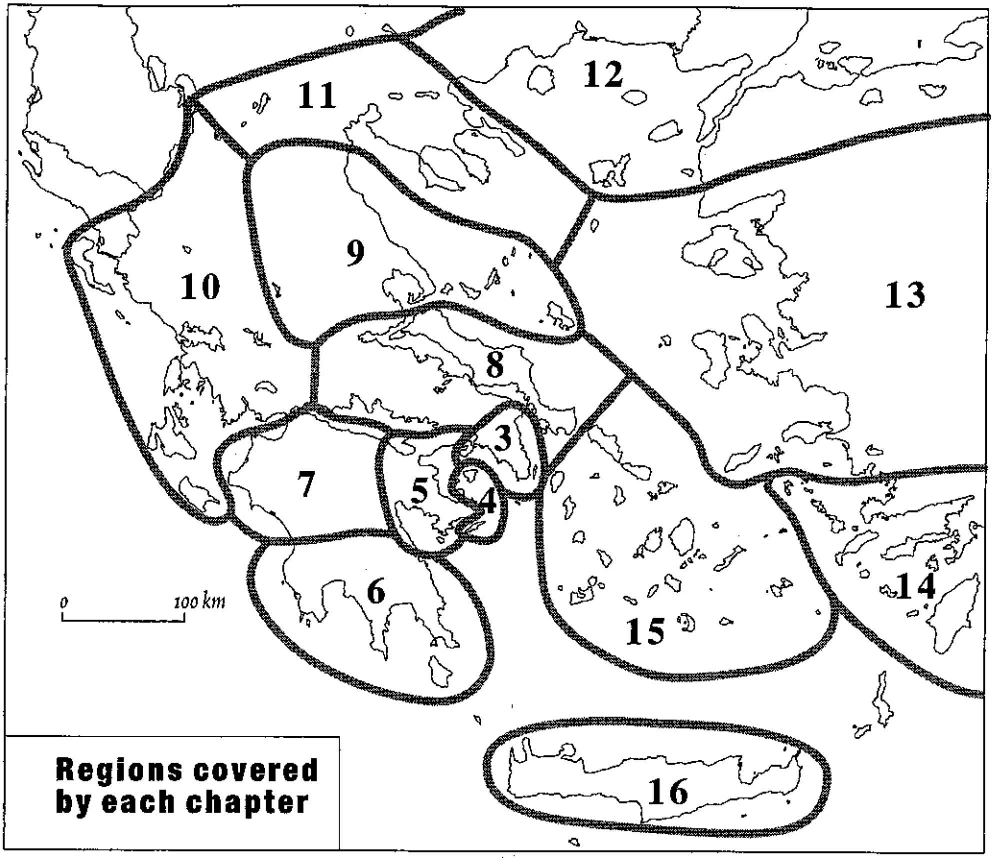


Fig. 2.7. The regions covered by Chapters 3 to 16.

the younger fill(s) is more problematical. It was originally suggested that it was due to the climate becoming drier, but there is little evidence for significant changes over the whole region.[191,235,237] A more likely cause is deforestation produced by people and maintained by grazing animals.[270,271]