Note to readers:

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The Physical Environment
A New Zealand Perspective

Edited by
Andrew Sturman and
Rachel Spronken-Smith

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Study of the physical environment preferentially focuses on the processes and phenomena occurring at the Earth's surface, where most human activity takes place. It is, however, important at the outset to describe the less visible yet fundamental internal Earth processes and phenomena, as well as the Earth materials that form the lithosphere, which is the foundation to our physical environment.

The Earth is a complex, dynamic planet that has evolved and changed throughout its 4.6-billion-year history. The internal and external Earth processes are to a large extent interdependent and interactive. For example, heat-driven internal 'tectonic' processes have created and moved continents about on the Earth's surface, and so opened and closed ocean basins throughout geologic time. These in turn have influenced ocean and atmospheric circulation patterns. Another example, more regional in nature, occurs where continents collide and also along continental margins where tectonic processes form high mountains, which in turn influence air circulation patterns and regional climates. Fundamentally the Earth's rock materials and internal tectonic processes provide the platform upon which all physical environmental processes exist and function.

Internally driven tectonic geological processes occur slowly and over long time periods. For example, it takes many tens to more than 100 million years for large ocean basins such as the Pacific Ocean to open and form, or conversely, to be destroyed. Similarly, large mountain chains form over millions of years. The three major groups of rock (igneous, sedimentary, and metamorphic) and their geological structures record the history as well as the rates of internal geological processes taking place. Igneous rocks record magma formation in the crust and upper mantle and associated volcanism, metamorphic rocks record episodes of large-scale tectonic movements and associated mountain building, and sedimentary rocks record the timing of uplift, erosion, basin formation, and deposition.

Within the geomorphic environment, landforms are the result of the interaction between the Earth's internal and external processes, with Earth materials exposed at the surface. Landforms are first and foremost related to tectonics, the geological structure and history of a region, as well as the erosion and sedimentation taking place under the influence of climatic and biological processes. The purpose of this chapter is to introduce the reader to the internal structure of the Earth, and the processes that
are fundamental to forming and controlling the distribution and location of the major surface features of the Earth's oceans and continents.

**The interior of the earth**

Internally, the Earth is comprised of three concentric layers. In the centre is the core, surrounded in turn by the mantle, which in turn is surrounded by the crust (Figure 2.1). This division is a function of matter differentiation, which is thought to have commenced at the time the Earth formed, along with the rest of the Solar System, from a solar nebulae made up of dust and gases. This cloud of matter is inferred to have been largely composed of compounds of silicon, iron, magnesium, oxygen, aluminium, and small quantities of other chemical elements. This compositionally uniform cloud of matter began to heat internally, and gravitationally began to differentiate heavy matter to the centre, while lighter minerals were sorted out to the outer layers, so forming the three-layered internal structure. Gases are also thought to have been progressively released from the interior of the Earth, ultimately leading to the formation of oceans and development of the atmosphere.

![Figure 2.1](image-url)  
The physical and compositional layering of the Earth. The chemical and mineral compositional layering is depicted on the right side, with variations in physical properties shown on the left. (Adapted from Abbott 1996)

The Earth's internal layers of differing density reflect variations in composition, as well as temperature and pressure, which both increase with depth. Density also increases from the surface to the centre (Figure 2.1). The core makes up approximately 16% by volume of the Earth. Analysis of earthquake waves passing through the interior of the Earth indicate that the core is divided into a solid inner core and a surrounding liquid outer core, both inferred to consist of iron and nickel with a calculated density of 10–13 gm cm⁻³ (Figure 2.1). Surrounding the core, the mantle is estimated to comprise approximately 83% of the Earth's volume. Its density is calculated to range
from 3.3 to 5.7 gm cm\(^{-3}\), and compositionally inferred to be dominated by minerals of iron and magnesium.

The mantle is further subdivided on the basis of physical contrasts in temperature and pressure conditions. The lower mantle, or mesosphere, is solid and makes up the bulk of the volume of the mantle. Surrounding this is the asthenosphere (weak sphere). Here, rocks compositionally similar to those in the lower mantle are weak because of the temperature-pressure conditions. Rocks in the asthenosphere behave in a plastic manner, slowly and constantly yielding, effectively flowing, but in a solid state. Partial melting occurs locally within the asthenosphere, generating magma (molten material), which may rise to the Earth’s surface and be erupted through volcanic activity.

Above the asthenosphere, the solid upper mantle and the compositionally distinct outermost layer of the Earth—the crust—form the lithosphere (or strong sphere) (Figure 2.2). For this outer lithospheric zone, pressure and temperature conditions are such that the rocks remain in a solid and rigid state. The lithosphere is in turn divided into plates that are moving over the weaker asthenosphere and are driven by slow mantle convection cells. These convection cells are maintained by heat exchange from the interior of the Earth and associated cooling. Thus mantle convection induces and maintains a complex array of rigid lithospheric plate motions, and these are

![Figure 2.2](image)
reflected at the Earth's surface by the formation of mountains, chains of volcanoes, ocean basins, deep submarine trenches, and belts of earthquake activity.

The Earth's crust is divided into two distinct types, with continental crust ranging in thickness from 25 to 60 km, with an average density of about 2.7 gm cm\(^{-3}\), and compositionally dominated by silicon and oxygen. Oceanic crust is much thinner, typically 5-10 km thick, much denser (\(-3.0\) gm cm\(^{-3}\)) than continental crust, and composed mainly of basaltic rocks.

**The principle of isostasy**

Isostasy is the mechanism by which regions of low-density crustal rock rise or subside until the mass of their topography (effectively that which lies above or below sea level) is buoyantly supported as it 'floats' on denser mantle rock below. Gravity surveys show that mountains are supported by low-density crustal 'roots', which project down into the mantle (Figure 2.3). The analogy is frequently drawn between this and icebergs floating in the ocean, with much of their mass below sea level. With continents and mountains, their elevation above the Earth's equilibrium surface (which is closely approximated by sea level) is compensated for by the presence of a thicker crust projecting into the mantle. The weight of crustal rocks is balanced by the weight of displaced mantle at depth. If a mountainous region is in isostatic equilibrium, then the higher the mountains, the deeper the crustal rocks will project into the mantle. Similarly, oceanic basins are underlain by thin denser oceanic crust, with the mantle rocks elevated with respect to sea level, compensating for the mass deficiency caused by the thin crust and ocean water. The lithospheric rocks generally achieve isostatic balance or equilibrium by crustal flexuring, involving regional crustal uplift and/or downwarping (Figure 2.3). For example, rapid erosion and sediment removal from elevated mountainous regions will reduce load, and isostatic adjustment will occur, causing the region to rise. Equally an adjacent low-lying region of thinner crust will receive the eroded sediment, increasing the load, thickening the crust, and inducing isostatic adjustment expressed as downwarping or subsidence. These crustal isostatic adjustments are thought to be accommodated by slow yield or plastic flow of the asthenosphere rocks in the upper mantle.

Perhaps the best example of the process and rates of isostatic adjustment is provided by the large regions of uplift since the close of the last glaciation in North America and Scandinavia. The loading by the continental ice sheet caused the continental crust to isostatically adjust by subsiding. Rapid melting and retreat of the ice sheet, so unloading the crust, caused the crust to isostatically rebound. In Canada this rebound continues, and in places amounts to more than 250 m of uplift in the last 10,000 years. Similarly, in Scandinavia isostatic rebound since ice retreat continues at a rate of about 1 cm per year.

Thus the tectonic cycle, to a large degree, underpins all other Earth system cycles. The geomorphic cycle in particular is dependent on the formation of hills and mountains, which in turn are eroded by wind, water, and ice. Ocean basins control global circulation patterns, which in turn are a factor in global climatic changes. It is at the surface that the complexities of the Earth's internal and external processes are most evident.
Continental drift, seafloor spreading, and the theory of plate tectonics

In 1915 German meteorologist Alfred Wegener first proposed the hypothesis of continental drift, basing this on the fit of the different widely separated continents, as well as the distribution of similar fossils and rocks of the same age. Wegener argued that all continents were once joined in a supercontinent he named Pangaea. Wegener's continental drift hypothesis did not receive general acceptance because it provided no explanation as to how the continents moved over the oceanic crust and what the mechanism of continental drift entailed.

In the 1950s and 1960s much new data, such as seafloor topography and magnetic geophysical data, became available from the deep ocean basins. This led to the next major advance in understanding—the concept of seafloor spreading. Two scientists,
Harry Hess and Robert Dietz, independently proposed that new ocean crust is formed along mid-oceanic spreading centres, coinciding with mid-oceanic ridges found in each of the major ocean basins. They proposed that continents and segments of oceanic crust move together, and suggested that the driving mechanism is some form of thermal convection within the Earth. Soon after, in the late 1960s, the theory of Plate Tectonics became established as the unifying hypothesis, accounting for seafloor spreading and continental drift as part of rigid plate movements. These rigid plates are comprised of oceanic and continental crust, as well as the upper mantle, together forming the lithosphere (Figure 2.4).

Seven large and five or more smaller lithospheric plates form the Earth's surface (Figure 2.5). These plates are in constant relative motion, driven by the Earth's internal heat transfer mechanism within the convecting asthenosphere. The relative rates of movement of one plate with respect to another along a plate boundary is typically of the order of centimetres per year. While the plates are internally strong and rigid, and are effectively 'rafted' on the yielding asthenosphere, they are also characterised by complex deformation and plate interactions along their margins. Here plates are created, deformed, and destroyed, giving rise to concentrated narrow belts of volcanic and earthquake activity. The plate boundaries are equally well defined morphologically by the huge submarine mountain chains forming the mid-oceanic ridges, deep narrow ocean trenches, and large mountain belts, and also by major faults (Figure 2.4).

There are three types of plate boundary recognised, and each is accompanied by distinctive large-scale morphologic expression at the Earth's surface (Figure 2.5): 

- **divergent boundaries** form where plates separate and move apart, allowing new lithosphere to form from the upwelling mantle-derived magma
- **convergent boundaries** form where plates collide and one subducts beneath the other, ultimately recycling lithosphere back into the mantle
- **transform fault boundaries** occur where plates slide past each other and lithosphere is neither created nor destroyed.

Divergent plate boundary margins form mid-ocean ridges (Figure 2.4). Here, the newly formed oceanic crust is less dense and buoyant, and remains elevated, so forming the many thousands of kilometres of submarine volcanic mountain chains found in each of the world's major ocean basins. The crests of these mid-ocean ridges are characterised by central rift valleys, and it is within these that the rising magma is emplaced, cools, and solidifies, and is accreted onto the separating oceanic lithospheric plate. The Mid-Atlantic Ridge represents perhaps one of the most impressive examples of a divergent plate boundary, and reflects the separation of North and South America from Europe and Africa, and the opening up of the Atlantic Ocean over the last 200 million years (Figure 2.5). Iceland straddles the Mid-Atlantic Ridge, having formed there because of the exceptionally large volume of magma erupted, and so offers the rare opportunity to study the rifting process and associated volcanic activity along the divergent plate margin.

New divergent plate boundaries may also be initiated within the continental lithosphere, and in such a setting linear rift valleys form. They signal the early stages in continent break-up, and ultimately the formation of a new ocean basin between two separated parts of the continent. The East African Rift is interpreted to reflect the
Convergent plate boundaries occur where plates collide, and generally, but not always, lithosphere is recycled by subduction to the mantle. The collision and subduction of one plate beneath the other is accompanied by the formation of a narrow, deep submarine trench, and belts of volcanism and earthquakes. Convergent
Margins can be further subdivided into several distinctive types, depending on the type of crust of each plate involved in the convergent margin setting. Andean-type margins form where the denser, thinner oceanic crust is subducted beneath the lighter and thicker continental crust. The margin of the overriding continental crust is deformed and rocks are complexly faulted and folded, forming mountain belts. The subducted lithospheric plate descends and becomes heated, and partial melting occurs in the wedge of overriding lithosphere generating magma. This magma in turn rises and may erupt onto the Earth's surface, forming a chain of volcanoes. The convergent South American–Nazca Plate boundary on the Pacific Ocean rim is a classic example of such a setting. In contrast, where oceanic crust is subducted beneath oceanic crust, such as is the case in the New Hebrides and near Japan, a trench and parallel chain of volcanoes forming an island arc mark such convergent margins.

The collision of two plates along a convergent margin is also responsible for shallow (<70 km), intermediate (70–300 km), and deep (300–700 km) focus earthquake activity (Figures 2.4 and 2.5). The shallow earthquake activity is confined mainly to the overriding crustal rocks near the plate margin, while the intermediate and deeper earthquakes are associated with the subducting plate.
When two continents collide, subduction is unlikely to occur, as both are comprised of lighter, more buoyant rock. In this situation one plate will override the other and an exceptionally thick continental crust results. This is best exemplified by the collision of India and Asia, with the Eurasian plate overriding the Indian plate, and resultant formation of the Himalaya Mountains.

Transform plate boundaries form where plates slide horizontally past each other, and lithosphere is neither created or destroyed. There are two locations where transform boundaries form. The first is where the divergent plate boundaries along mid-oceanic ridges are offset. The second location occurs in a continental setting, and the best known and most studied example is the San Andreas Fault in California, where the Pacific Plate is sliding northward past the North American Plate at a relative annual rate of about 55 mm. Earthquake activity along transform faults is confined to shallow crustal depths.

**Earth materials**

The materials that make up the Earth’s crust are very diverse, reflecting their varied origins. Since the formation of the Earth about 4.6 billion years ago, crustal and mantle materials have continuously been cycled as the result of the combined effects of internal and external processes. For example, molten rock from the lower crust and mantle is erupted on to the Earth’s surface. It is then exposed to external processes that break it down, erode, and transport it, ultimately depositing the particles on the ocean floor, where it is buried, consolidated, and cemented to become rock again.

This section outlines the mineral ‘building blocks’ of the many rock types known, and also introduces the three main groupings of rock based on their mode of origin. Simple classification schemes for each of the three rock groups are also presented, and define the most common rock types.

**Minerals and rocks**

The Earth’s surface physical environment is fundamentally dependent on the geological materials comprising the crust, and the internal and external processes continually modifying it. The Earth’s crust is composed of rocks, and rocks in turn are formed by aggregations of one or more minerals. We define minerals as naturally occurring inorganic crystalline solids, with definite chemical compositions and physical properties. Minerals have a characteristic crystal shape, indicative of a stable atomic structure. Well over 3000 naturally occurring minerals have been recognised and described, but the eight most common rock-forming minerals make up more than 90% of all crustal rocks (Table 2.1).

The relatively small number of common rock forming minerals is a direct reflection of the few elements found in major abundance in the Earth’s crust. Almost 99% of the crust is made up of just eight elements. In relative abundance by weight, these are oxygen (46%), silicon (28%), aluminium (8%), iron (6%), magnesium (4%), calcium (2.4%), potassium (2.3%), and sodium (2.1%). Mineral classification is based on
Table 2.1 Common rock-forming minerals.

<table>
<thead>
<tr>
<th>Mineral</th>
<th>Abundance in crust (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Plagioclase*</td>
<td>39</td>
</tr>
<tr>
<td>Quartz</td>
<td>12</td>
</tr>
<tr>
<td>Orthoclase*</td>
<td>12</td>
</tr>
<tr>
<td>Pyroxenes</td>
<td>11</td>
</tr>
<tr>
<td>Micas</td>
<td>5</td>
</tr>
<tr>
<td>Amphiboles</td>
<td>5</td>
</tr>
<tr>
<td>Clay minerals</td>
<td>5</td>
</tr>
<tr>
<td>Olivine</td>
<td>3</td>
</tr>
</tbody>
</table>

* Feldspar group of minerals

their chemistry. In turn, the most important group of minerals is the silicates, formed by oxygen and silicon. Within this group, quartz, feldspar, and the ferromagnesium minerals are the most important and common. Other important mineral groups include oxides, hydroxides, sulphides, and carbonates.

Geologists have long recognised three major groupings of rock: igneous, sedimentary, and metamorphic. The basis of these groupings is their mode of origin. The appearance of rock is determined by its mineralogy, or the relative proportions of its constituent minerals, and by its texture—that is, the size and the shape of the mineral grains or crystals and the way these are associated within the rock.

**Igneous rocks**

Igneous (from Latin ignis—fire) rocks form by crystallisation from a magma, a molten rock originating deep beneath the Earth's surface, within the lower crust or mantle, where temperatures are in excess of 700°C. As magma cools, different minerals will crystallise out of the melt at different temperatures, and the resulting rock will be made up of an assemblage of different interlocking mineral crystals. Thus the type of igneous rock formed during cooling will depend on the original chemistry of the magma. The darker-coloured rocks contain less silica (SiO₂) and are referred to as basic igneous rocks, while the lighter-coloured rocks are more silica-rich and are referred to as acidic. Two major groups of igneous rocks—intrusive and extrusive—can be distinguished by crystal size.

Intrusive igneous rocks form where magma is intruded into rock deep beneath the Earth's surface (Figure 2.6), and crystallisation occurs slowly, favouring the growth of large interlocking mineral crystals visible to the naked eye. Rock bodies formed by igneous intrusion are referred to as plutonic rocks. Granite is a common example, while other common igneous rock types are listed in Table 2.2.

Extrusive igneous rocks form from volcanism, where magma is extruded onto the Earth's surface and cools rapidly to form volcanic rocks (Figure 2.6). The rocks formed by volcanic activity range from lava flows to more explosive volcanism. In the former, the erupted magma has flowed for some distance on the surface before solidifying, and is characterised by a fine crystalline or glassy texture. In the latter, fragmented magma is erupted into the atmosphere and almost instantaneously crystallised forming ash particles.
General Concepts and the Origins of New Zealand

Figure 2.6 Schematic diagram illustrating the origin of intrusive and extrusive igneous rocks. (Adapted from Skinner & Porter 1989)

Table 2.2 Classification scheme for common igneous rocks.

<table>
<thead>
<tr>
<th>Magma type</th>
<th>Typical minerals</th>
<th>INTRUSIVE Coarse-grained</th>
<th>EXTRUSIVE Fine-grained</th>
</tr>
</thead>
<tbody>
<tr>
<td>Basic</td>
<td>Dark-coloured: pyroxene, amphibole; calcium plagioclase feldspar</td>
<td>Gabbro</td>
<td>Basalt</td>
</tr>
<tr>
<td>Intermediate</td>
<td>Both light- and dark-coloured minerals</td>
<td>Diorite</td>
<td>Andesite</td>
</tr>
<tr>
<td>Acid</td>
<td>Light-coloured: quartz, orthoclase feldspar; sodium plagioclase feldspar; mica</td>
<td>Granite</td>
<td>Rhyolite</td>
</tr>
</tbody>
</table>

Sediments and sedimentary rocks

Rocks that are exposed at the Earth’s surface are either broken into fragments, forming sedimentary (clastic) particles or grains, or are slowly dissolved by physical, chemical, or biological weathering processes. These particles and dissolved materials may in turn be eroded and transported by ice, water, or wind, and become deposited as sediment. Clastic sediments include gravel, sand, silt, and clay, and chemical and biochemical sediments formed by precipitation from solutions such as halite (NaCl) and calcite (CaCO₃, typically forming shells). The diversity of environments in which sediments may accumulate to form sedimentary rocks are shown in Figure 2.7.

Once accumulated, sediments are lithified into hard rock in two ways: by compaction, where the individual grains are progressively more tightly packed after burial by the increasing weight of overlying sediments; and by precipitation of minerals into the pore space between particles, so cementing them together and ‘hardening’ the sediment to form a sedimentary rock. Common examples of sedimentary rocks include conglomerate from gravel, sandstone from sand, siltstone from silt, mudstone from silt and clay, and limestone from shells and shell fragments.
Sedimentary rocks are often characterised by sedimentary structures indicative of the environment of deposition. Sedimentary bedding is formed when distinctive layers reflect variations in mineralogy and particle or grain size. Within individual beds more detailed sedimentary structures such as laminations, ripple marks, and cross-bedding are often indicative of the depositional environment where the sediments accumulated. Together with any fossils contained in the sediment, these structures are useful in interpreting the geological history of a region.

Sedimentary rocks cover much of the Earth's surface but form only a relatively thin layer over the igneous and metamorphic rocks, which constitute much of the crust as a whole. The most common sedimentary rock types are listed in Table 2.3.

**Table 2.3** Classification scheme for common sedimentary rocks.

<table>
<thead>
<tr>
<th>Particle size (mm)</th>
<th>Clastic sedimentary rocks</th>
<th>Chemical sedimentary rocks</th>
<th>Biochemical sedimentary rocks</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Sediment</td>
<td>Rock</td>
<td>Limestone (CaCO₃)</td>
</tr>
<tr>
<td>Boulders 256-</td>
<td>gravel</td>
<td>Conglomerate</td>
<td>Limestone (CaCO₃)</td>
</tr>
<tr>
<td>Cobbles 64-</td>
<td>Breccia</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Pebbles 4-</td>
<td>Sand</td>
<td>Sandstone</td>
<td></td>
</tr>
<tr>
<td>Granules 2-</td>
<td></td>
<td>Greywacke</td>
<td></td>
</tr>
<tr>
<td>Silt 0.062-</td>
<td>silt</td>
<td>Siltstone</td>
<td></td>
</tr>
<tr>
<td>Clay 0.004-</td>
<td>clay</td>
<td>Mudstone/Claystone</td>
<td>Halite (NaCl)</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Gypsum (CaSO₄·2H₂O)</td>
</tr>
</tbody>
</table>
**Metamorphic rocks**

Metamorphic (from Latin meta—change, and morphe—form) rocks form when rock is subjected to high temperatures and pressures and fluid rock interaction deep beneath the Earth's surface, changing the mineralogy, texture, and/or chemical composition without melting. All types of pre-existing rocks (igneous, sedimentary, and metamorphic) may be altered by metamorphism.

Regional metamorphism occurs over geographically extensive areas along convergent plate boundaries, and such metamorphosed rocks will be exposed at the Earth's surface by the associated episode of mountain building (Figure 2.8a). Contact metamorphism occurs in more restricted areas, associated with a major igneous intru-

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**Figure 2.8** The formation of metamorphic rocks by (a) regional metamorphism in response to burial and tectonic deformation associated with the development of mountain systems or though large igneous intrusions such as granite batholiths; (b) contact metamorphism surrounding an igneous intrusion. (Adapted from Fitcher & Farmer 1977)
The type of metamorphic rock produced depends on the distance from the intrusion and its temperature, the inner metasomatic aureole forming in response to both high temperatures and fluid flux, while the outer thermal aureole is solely temperature-dependent. Rocks altered by regional metamorphism typically form a fabric known as foliation. These are closely spaced planes formed during structural deformation and progressive recrystallisation. Examples of common foliated metamorphic rock types, ranging from low grade to higher grade, include slate, phyllite, schist, and gneiss. Typical contact metamorphic rocks include quartzite and marble, which are characteristically made up of equi-dimensional-shaped mineral grains.

The most common metamorphic rocks are listed in Table 2.4.

Table 2.4  Classification scheme for common metamorphic rocks.

<table>
<thead>
<tr>
<th>Mineral grain appearance</th>
<th>Metamorphic rock name</th>
<th>Pre-metamorphic parent rock type</th>
</tr>
</thead>
<tbody>
<tr>
<td>Finely foliated fabric</td>
<td>Slate</td>
<td>Mudstone, shale or greywacke</td>
</tr>
<tr>
<td></td>
<td>Phyllite</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Schist</td>
<td></td>
</tr>
<tr>
<td>Coarsely foliated fabric</td>
<td>Gneiss</td>
<td>Sandstone, conglomerate, granite</td>
</tr>
<tr>
<td>Equi-dimensional grains</td>
<td>Hornfels</td>
<td>Mudstone, shale, limestone sandstone</td>
</tr>
<tr>
<td></td>
<td>Marble</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Quartzite</td>
<td></td>
</tr>
</tbody>
</table>

Figure 2.9  The conceptual rock cycle, illustrating the complex inter-relationships that exist between the Earth's internal and external processes with respect to each of the three major rock groups. (Redrawn from Monroe & Wicander 1994)
**The rock cycle and plate tectonics**

The rock cycle is a conceptual approach to understanding the way that Earth materials are continually being recycled between the igneous, sedimentary, and metamorphic environments. The cycle is illustrated in Figure 2.9 and shows how the Earth's internal processes (including magma formation and metamorphism) and the external processes (including weathering, erosion, sediment transportation, deposition, and lithification) and the three rock groupings are fundamentally interrelated. Effectively, the Earth's surface represents the interface between these processes. However, it must be remembered that the rock cycle is fundamentally driven by the Earth's internal tectonic processes, and that this is expressed at the surface by lithospheric plate movements.

**Rock deformation and geological structure**

The effects of tectonic forces are concentrated and especially evident in relatively narrow zones along the world's plate margins. Here, rock deformation is fundamentally related to the relative motion between adjacent plates, and rock formations are complexly deformed and displaced by faulting and folding in response to the plate motions. That this process is active and ongoing along plate margins is demonstrated by the associated regions of earthquake activity.

There is much to be learnt from studying the deformation of rocks along active plate boundaries: it not only reveals the geological history of the margin, but also provides insights into how our landscapes are fundamentally controlled by the tectonic forces operating. Analysis of the landforms further reveals much about the geometry and rates of earth deformation processes. Before we can turn our attention to the mountain building along the Australia–Pacific Plate boundary in New Zealand, we need to outline the basic terminology and processes of rock deformation.

**Rock deformation**

**Stress** (directional force) is important in changing the volume and shape of rock masses. Rock may be affected by elastic, brittle, or ductile strain, where strain is the deformational response of the rock mass to the imposed stresses. The three types of stress are compression, tension, and shear (Figure 2.10). Compression occurs when a volume of rock is squeezed by external forces and the resulting strain involves a shortening or contraction. Tension occurs when the external forces stretch a volume of rock and the resulting strain involves a lengthening or extension. Shearing occurs when the external forces, acting on opposing sides of a volume of rock, are pushing in opposite directions and the resulting strain is an angular distortion or change in shape.

Elastic strain of rock is that component of the deformation that is recoverable when the stresses are reduced or released. The maximum elastic strain that may be accommodated by rock in practice is very small, and once forces deform rock beyond the elastic limit, a permanent deformation will occur. This represents plastic strain and may be recognised by folding and/or faulting of the rock mass (Figure 2.10).
Brittle rock deformation occurs in the upper crust and at the Earth’s surface. Here, rocks accommodate the accumulating tectonic forces or stresses by movement on fractures called faults. This brittle strain is frequently associated with earthquake activity, when movement and deformation is accommodated suddenly and episodically once the frictional resistance to slip on a fracture in the rock mass is overcome. Larger volumes of rock may also respond to compressional forces by folding at or near the Earth’s surface.

Ductile rock deformation generally occurs at depths of 10 or more kilometres beneath the surface, where temperatures and pressures are such that the rock will continuously and slowly respond by yielding, and slow recrystallisation of minerals into more stable forms also occurs. This results in the formation of metamorphic rocks.

The tectonic forces dominating the style of rock deformation along plate boundary zones are: compressive forces along convergent margins, tensional forces along divergent margins, and shearing forces along transform margins.

**Describing deformed rocks**

Field geologists record and describe deformed rocks by recording information about their geometry. Data are collected from rock outcrops, where the underlying bedrock is exposed at the surface and not obscured by soil or vegetation cover. For example, most sedimentary rocks are deposited in near-horizontal layers separated by bedding planes. Thus, if such stratified sequences are subsequently tilted, their orientation relative to the horizontal needs to be recorded. This is achieved by measuring the strike
and **dip** (Figure 2.11). The measurements are made with a compass, and the strike is recorded as a bearing relative to north; the dip angle is recorded with an inclinometer, which is usually included in a geological compass. The strike is the direction of the line of intersection between the rock layers represented by bedding planes and the horizontal plane. The dip angle represents the maximum angle of tilt relative to the horizontal of the bedding planes. The direction of dip is at right angles to the strike.

![Figure 2.11](image)

**Figure 2.11** Definition of strike and dip as applied to the orientation of rock strata.

Where strata have been more complexly deformed by faulting and folding, a number of measurements are required to correctly record and describe the deformation of the rock sequence. Faults are planar structures, too, and may be described in a similar way.

**Folding** of rock is a permanent bending deformation resulting from compressional forces, and while many folds have formed beneath the ground surface, in tectonically active regions such as convergent margins it is now well documented that large-scale folding at the Earth's surface is common and is an important mechanism in mountain-building processes.

The two basic types of fold are an **anticline**—where the strata are up-arched or bent upward into a ridge—and a **syncline**—where strata are down-arched or bent down to form a trough (Figure 2.12).

![Figure 2.12](image)

**Figure 2.12** Basic fold terminology with regard to (a) horizontal anticlinal and synclinal axes, and (b) a plunging anticlinal axis. (Adapted from Skinner & Porter 1989)
The sides of a fold are referred to as the limbs and are separated by the axial plane, which divides a fold into two parts. The intersection of the axial plane with the crest of an anticline and the trough of a syncline is referred to as the fold axis. Where a fold has an inclined axis, it is said to be a plunging fold, and the angle between a fold axis and its horizontal projection is referred to as the fold plunge (Figure 2.12). Anticlines and synclines typically form together and share a common limb.

Folds may also be described as symmetrical and asymmetrical (Figure 2.13). A symmetrical fold is one where the axial plane is vertical, and the limbs dip at approximately the same angle. If the axial plane is inclined and limb dips are different, the fold is described as asymmetric. Overturned folds occur where both limbs dip in the same direction, and the strata of one limb are actually overturned from their original orientation. In the case of a recumbent fold, the axial plane is near horizontal and the fold is then referred to as a recumbent structure. In geologically more complexly deformed mountainous regions, such as at convergent plate margins, the folds developed are typically more complex and there may be several phases of folding superimposed.

![Types of folds: (a) symmetric, (b) asymmetric, (c) overturned, and (d) plunging. (Adapted from Press & Siever 1998; Monroe & Wicander 1994)](image)

Rock formations close to the Earth's surface are brittle (inelastic and rigid) and therefore readily fracture. Rock fractures across which there has been no movement of the block on one side relative to the other are referred to as joints, while those fractures across which there has been relative movement parallel to the fracture plane of the block on one side relative to the other are referred to as faults.

Joints vary widely in scale, from microscopic fractures of individual mineral crystals, to rock fractures extending over hundreds of metres to kilometres. Most commonly joints form in response to rock deformation. Typically, such joints occur as near-parallel and relatively evenly spaced sets, although multiple sets of joints with
different orientations are commonly observed in the field. Joints also form during the cooling of igneous rocks emplaced as lava flows at the surface, or as magma intrusions at depth beneath the surface. Tectonic uplift, erosion, and associated unloading of rock masses are also responsible for the development of sheet and/or exfoliation jointing. Such jointing may be less systematic and form near parallel to the topographic surface. Joints are an important factor in facilitating surface weathering and erosion processes. Fractures and joints are also important in facilitating groundwater movement within rock formations. For example, in regions where limestone outcrops at the surface, solution weathering may lead to the development of a karst landscape. Typically, sinkholes will form, aligned along joints within the limestone where solution weathering is preferentially concentrated, resulting in cave formation (see Chapter 17).

Faults represent the brittle deformation response of the rock mass to compressive, tensional, or shearing forces. Faults are classified according to the relative movement or slip across the fault plane. The orientation of a fault plane is defined by the strike and dip, as described earlier. Accordingly, we can define different types of faults with respect to the fault plane orientation (Figure 2.14).

Dip-slip faults are those where the relative movement of one side to the other is parallel to the direction of dip (Figures 2.14a–c). If the hanging wall block has moved down the fault plane, the fault is called a normal fault, and this occurs under tensile forces. If the hanging wall block has moved up the fault plane, it is called a reverse fault. When a fault plane has a low angle of dip, less than $45^\circ$, and the hanging wall block has moved up the fault plane, it is called a thrust fault. Reverse and thrust fault movement occurs under compressional forces.

Where two or more normal faults are present, block faulting may lead to the development of a downthrown block called a graben, or alternatively, an elevated or uplifted block referred to as a horst (Figures 2.14d–e). Horst ridges and graben depressions are bounded on two sides by near-parallel normal faults and may form through a wide range of scales. In regions affected by horizontal extension, very large-scale block faulting may form a horst, which is characterised by elevated topography and dissection, leading to the formation of mountains and ranges bounded along their margins by normal faults. Graben valleys, also bounded by normal faults, will receive sediment eroded from the surrounding elevated topography.

Strike-slip or lateral faults are defined by horizontal movement parallel to the fault plane strike orientation (Figures 2.14f–g). Such faults occur because of shearing forces acting across a region. Strike-slip faults are described as right-lateral or left-lateral, based on the relative direction of movement of the ground on one side of the fault to the other.

Oblique-slip faults occur where relative movement across the fault includes both components of horizontal and vertical slip (Figure 2.14h).

**Landscape development**

The preceding sections have outlined the main processes responsible for the large-scale geomorphic structure of the Earth's surface. The following sections examine,
Mountains and mountain building

A mountain can be defined in a general sense as any large mass of rock rising above the common level of the surrounding land. While some mountains occur as isolated, single peaks, more frequently mountains occur as a series of associated peaks forming a mountain range, and typically these will be of similar age and origin. At an even larger scale the Rocky Mountains in North America, the Andean Mountains in South America, the Himalayas, and the Southern Alps in New Zealand are examples of mountain systems, characterised by a tremendously complex geological history resulting from structural deformation accompanied by metamorphism and plutonic and
volcanic activity. Such mountain systems are also typically associated with significant crustal thickening along active convergent plate margins.

The classification of mountains is closely linked to their origin. Four basic types of mountain landforms are commonly recognised (Figure 2.15): volcanic mountains, block-faulted mountains, folded mountains, and thrust mountains. This subdivision closely reflects the role that igneous and tectonic processes play in mountain formation.

The term mountain chain is used to link several mountain systems, with no implicit relationship in terms of origin or age (Figure 2.16). An example of such a chain is the American Cordillera, stretching from Alaska to the southern tip of South America.

Mountainous landforms may form away from plate boundaries and involve little tectonic deformation. For example, the formation of intra-plate isolated volcanic mountains such as the Hawaiian Islands is related to the presence of a magma hot spot deep beneath the surface in the mantle. However, the majority of mountain systems, and the most spectacular, have formed along plate margins, such as along the margins of the Pacific Ocean, and also the Alpine–Himalayan mountain belt (Figure 2.16). In the deep ocean basins the globally linked mid-oceanic ridges represent huge mountain chains associated with magma intrusion and volcanism along the divergent plate boundaries. Elsewhere, along convergent plate margins, mountain systems are forming

Figure 2.15 The classification of mountains based on mode of origin and geological structure include:
(a) volcanic mountains and volcanic arcs, (b) block-faulted mountains, (c) fold mountains, (d) thrust mountains.
(Adapted from Press & Siever 1998)

Figure 2.16 (facing page) (a) Global mountain chains associated with plate margin settings (depicted in blue), illustrating: (b) subduction of the oceanic Nazca Plate beneath the continental South American Plate to form the Andes Mountains; (c) subduction of the oceanic Pacific Plate beneath the oceanic Eurasian plate, forming the volcanic island arc of Japan; (d) continent–continent plate collision leading to the formation of the Himalayan Mountains. (After Monroe & Wicander 1994; Press & Siever 1998)
in response to the compressional forces associated with the collision of tectonic plates. Mountains can also form along transform plate margins in continental crust. The mountain-building process, or orogenesis, is discussed below with respect to the Australia–Pacific Plate boundary zone in the New Zealand region, and contrasting tectonic elements of the plate margin are then used to illustrate the tectonically driven orogenic processes involved.

**Present-day plate boundary tectonics of New Zealand: from oblique subduction to continental collision**

More than 600 million years of geological history is recorded by the diversity of rock groups exposed in the New Zealand landmass, which are representative of the larger, submerged micro-continental block (Figure 2.17). Over much of this time the micro-
continental block was located adjacent to the Australia–Antarctic sector of Gondwana, prior to its break-up about 90 million years ago by rifting and seafloor spreading.

Today, the largely submerged New Zealand micro-continent straddles the boundary between the Australian and Pacific Plates, and the relative motion of these plates over the last 20 million years has controlled the evolution and present shape of the emergent New Zealand landmass (Figures 2.17 and 2.18a, b). The present-day obliquely convergent relative motion between these plates ranges from about 50 mm per year near East Cape in the North Island, to approximately 30 mm per year near Fiordland in the South Island (Figure 2.18b). The magnitude of the components of motion perpendicular and parallel to the plate boundary varies systematically from

![Figure 2.18](image)

**Figure 2.18** The main structural features of the New Zealand micro-continent straddling the Australia–Pacific Plate boundary zone, with regional setting shown in the inset. Numbered arrows show rates of relative convergence in mm per year. Inverted V notation represents active volcanic arcs. Shading represents anomalously thick oceanic crust associated with higher sections of the seafloor.
north to south, as a function of the movement of the tectonic plates on the globe. Thus the rate of plate convergence is highest in the east coast North Island region, while the component of plate boundary parallel (right-lateral) shear is greatest in the South Island.

Although the annual rate of relative motion between the plates is small, the cumulative effect over geologically long periods of time is spectacular. This is shown by the separation of once-continuous belts of rock by over 460 km along the Alpine Fault in the South Island, the formation of the Southern Alps, and volcanic activity in the North Island. The ongoing deformation along the plate boundary zone is also reflected by widespread earthquake activity (Figure 2.19a). Shallow earthquakes occur in a broad zone parallel to the plate boundary through New Zealand (Figure 2.19b), and reflect the extent of the active earth deformation occurring in response to the oblique plate collision. The present-day zone of earth deformation associated with the Australia–Pacific Plate boundary in New Zealand gives insights into the evolution of a tectonically formed mountainous landscape (as shown in Figures 2.19c and d). The Southern Alps represent the major component of this landscape, formed largely in response to uplift on the Alpine Fault, which is the largest active fault in New Zealand.
Zealand. It extends for over 800 km from the Puysegur Trench and offshore Fiordland at the southwest tip of the South Island, to the northeastern coast at Cook Strait (Figure 2.18). The dominant component of movement on the fault is horizontal, with rock belts east of the fault in southern South Island separated by movement on the fault from matching strata more than 460 km to the north and west of the fault. Generally, the fault is an east-dipping oblique right-lateral strike-slip fault.
Present-day slip rates on the Alpine Fault are best considered in terms of the horizontal and vertical components. The horizontal slip rate varies along the length of the fault, but for the section of the fault adjacent to the central Southern Alps, estimates based on geological and geodetic surveying field work indicate annual rates ranging from 22 to 30 mm horizontal, and 6 to 8 mm vertical.

Along the west margin of the Southern Alps a narrow band of high-grade metamorphic rock, formed at mid to lower crustal levels, lies parallel and adjacent to the central section of the fault (refer Figure 2.17), reflecting the substantial long-term uplift of the eastern or Pacific Plate side. Given the present-day rate of uplift, these rocks may have been elevated to the surface from a depth of 20 km within the last three million years. Equally, the horizontal separation of matching rocks across the Alpine Fault can be achieved in less than 20 million years if present-day slip rates are applied throughout this period of time.

Summary

This chapter has reviewed the internal composition and layering of the Earth, the materials that make up the Earth's crust, and the tectonic processes that are fundamentally responsible for shaping the Earth's surface. The theory of Plate Tectonics provides the underpinning, unifying framework for an understanding and interpretation of the Earth's composition and the internal processes, the crustal rock cycle, and the formation and significance of rock structures. In terms of Earth systems, it provides a rational understanding for the formation and destruction of continents and ocean basins, and how the Earth's internal processes may significantly influence the lithosphere–atmosphere–hydrosphere interactive system. The emphasis has focused on the large-scale dynamic processes responsible for the major features of New Zealand's landscape, while volcanic landform development is described in more detail in Chapter 3.

Further reading