**Chapter 1**

# Deformation, stress and strain

## Terms and concepts

### *Deformation*

In structural geology, deformation refers to a change in the shape of a rock body, without implying deterioration. Studying deformation involves understanding how rocks behave under varying temperature and pressure in response to applied forces. This requires knowledge of the relationships between force, stress, and strain, which are strictly defined in geology and engineering, differing from their general everyday meanings.

### *Stress and strain*

In structural geology, deformation results from forces acting on a rock body, creating a stress field that leads to a change in shape, known as strain. Strain is a geometrical concept that must be measured accordingly, often using complex mathematical models, though qualitative approaches can also be useful. However, it is impossible to reverse-engineer the initial unstrained state or the exact causal stress from the final strain, as there are countless possible paths leading to the observed deformation.

### *Force, stress and pressure*

The term '**force**', defined strictly, means '*that which causes acceleration in a body*'; it is the product of the mass of the body and its acceleration. However, this concept is not much use to us in understanding geological structures; with the exception of earthquakes and volcanic explosions, acceleration is irrelevant to understanding natural deformation in rock.

For most practical purposes, the geologist will wish to convert a force, or system of forces, into a stress or stress field. A **stress** is a pair of equal and opposite forces acting on a unit area of a surface (Figure 4.1A) and a **stress field** is the system of stresses acting in three dimensions on a body.

The effect of a force of given magnitude or strength depends on the size of the area on which it acts. This can be easily demonstrated by placing a weight on a large piece of floating wood, say, which stays afloat, and comparing it with the effect of the same weight on a smaller piece of wood, which sinks under the weight (Figure 4.1B). Thus the same size of force will produce different sizes of stress depending on the surface area of the body on which the force acts, since *stress equals force divided by area*.

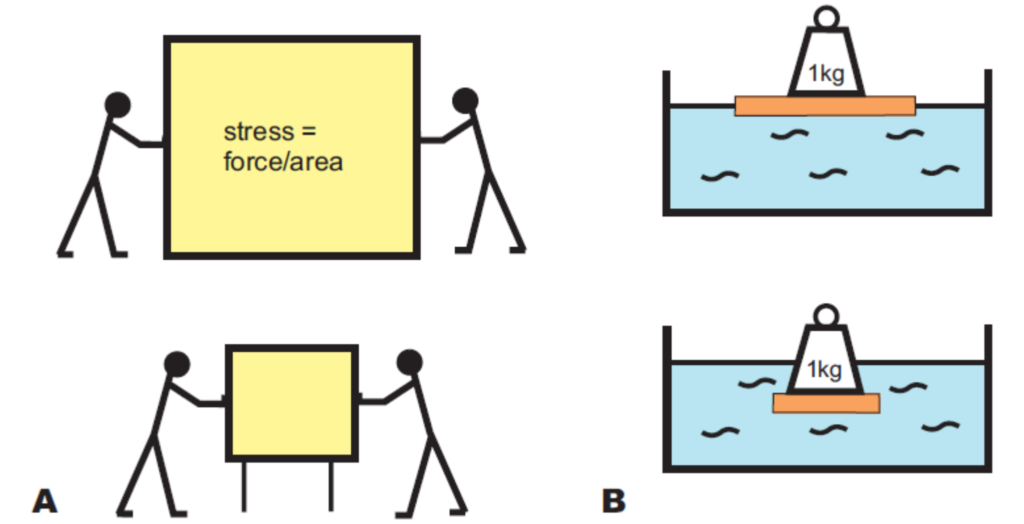


Figure 4.1 Force and stress. A. The stress acting on the yellow box is produced by two equal and opposite forces, divided by the area on which the force acts (i.e. the side of the box); a smaller box, subjected to the same forces, will be subject to a larger stress, since the force is divided by a smaller area. B. A 1 kg weight resting on a flat slab floats, whereas the same weight resting on a smaller slab sinks, because the force (the 1 kg weight) is spread over a smaller area.

At depth, rocks experience **lithostatic pressure**, which results from the weight of overlying rocks and is **equal in all directions**, similar to water pressure in deep seas. In addition to this gravitational stress, **tectonic forces** introduce **directional stresses**, which can be either **compressional** (positive) or **extensional** (negative).

The total stress on a rock at depth consists of two components: **lithostatic pressure (mean stress)** and **directional stress** (the difference between maximum and minimum stress), with the latter being responsible for potential **deformation**.

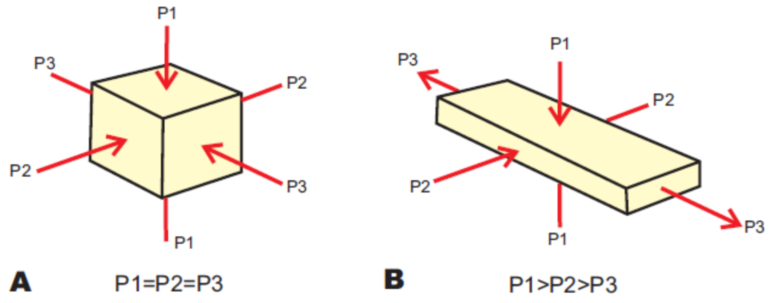


Figure 4.2 Pressure. A. in lithostatic pressure, the stresses exerted on a body in every direction are equal; this can be represented by three mutually perpendicular stress axes P1, P2 and P3, where P1=P2=P3. B. A variable stress field can be represented by three mutually perpendicular stress axes such that P1 is the maximum stress, P3 the minimum and P2 the intermediate, i.e. P1>P2>P3. The shape of the box illustrates diagrammatically the strain of an initial cube subjected to a stress field of this type. Note that P3, the minimum stress here, is negative, i.e. an extension.

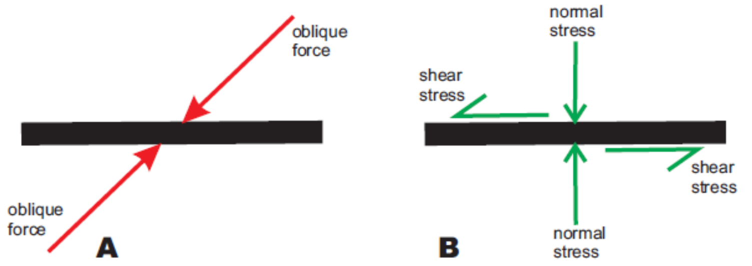


**Figure 4.3** Normal stress and shear stress. An oblique force opposed by an equal and opposite force acting on a thin slab (**A**) can be represented by a normal stress acting at right angles to the slab and a shear stress acting parallel to the slab (**B**) and tending to move the top of the slab to the left and the bottom to the right; this is sometimes referred to as a rotational stress; in this case the material inside the slab will tend to rotate in an anticlockwise sense, i.e. the strain will be rotational (see Figures 4.4, 4.5).

## *Normal stress and shear stress*

Forces and stresses often act **obliquely** to a surface rather than just at right angles. In such cases, the resulting **stress** can be divided into two components:

- **Normal stress**: Acts **perpendicular** to the surface, either inhibiting movement (if compressive) or facilitating it (if extensional).
- **Shear stress**: Acts **parallel** to the surface, trying to move the two sides in opposite directions.

**Shear stress** plays a key role in **faults and shear zones**, influencing how easily rock blocks move relative to each other. The balance between **shear and normal stress** determines the likelihood of movement along a surface.

## *Strain*

**Strain** refers to a change in a rock's **shape and/or volume**. A change in **volume alone** is called **dilation**, which can be an **increase or decrease**. Measuring dilation in rocks is challenging, but **volume reduction** is commonly linked to **high lithostatic pressure**, leading to **pore space reduction, fluid expulsion, and mineral transformations** into higher-density forms.

Changes in **shape** due to deformation create **characteristic rock fabrics** that are often **visible and measurable**. Shape changes can involve **distortion, rotation, or both**. **Shear strain (rotational strain)** occurs when a rock undergoes **distortional strain** in response to **shear stress**, causing elements of the original rock body to **rotate**.

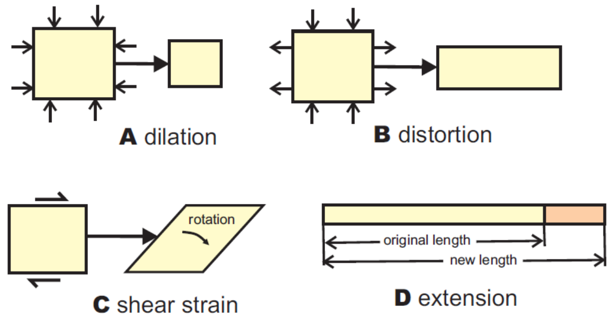


**Figure 4.4** Types of strain. **A.** Dilation = volume change; **B.** Distortion = shape change; **C.** Shear strain: the distortional strain produced by a shear couple; **D.** Strain in one dimension: the extension = the change in length (pink) divided by the original length (yellow).

## *The description of strain*

**Strain** in one dimension is measured as **extension**, which can be **positive or negative** and represents the proportional change in length (e.g., 10% shortening). Since strain is a **ratio**, it has **no units**.

In **three dimensions**, strain can be determined by measuring how the lengths of specific lines in a body have changed. While the original shape may vary, it is usually described as if it were **equidimensional**. Strain is then analyzed using **three perpendicular axes**, corresponding to the **maximum, intermediate, and minimum extensions**.

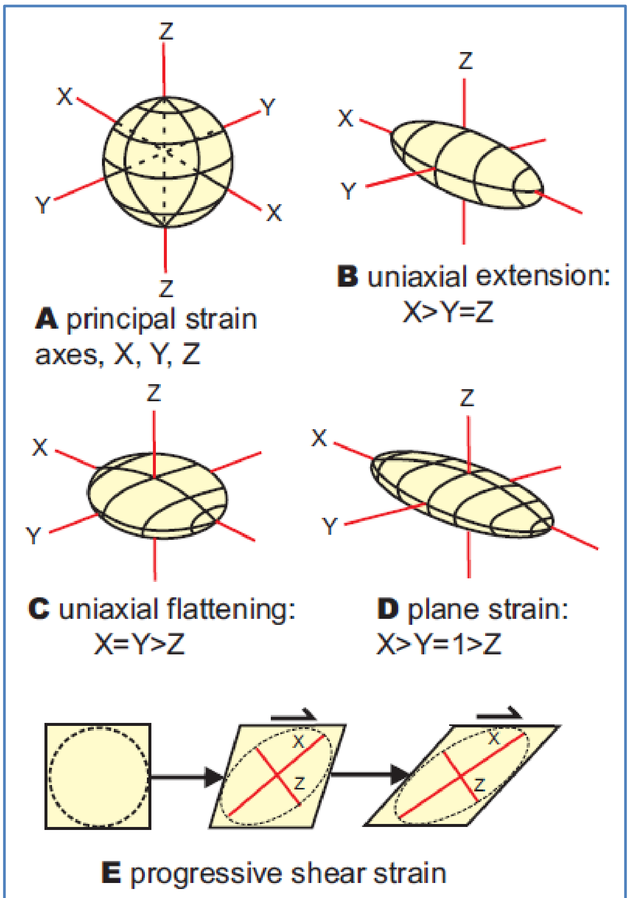


**Figure 4.5 A–D.** Co-axial strain in three dimensions affecting an initial sphere: **A.** The three principal strain axes X, Y, and Z. **B.** Uniaxial extension: the maximum strain axis (X) is greater than the intermediate (Y) and minimum (Z) axes, which are equal; **C.** Uniaxial flattening: the minimum strain axis (Z) is smaller than the intermediate (Y) and maximum (X) axes, which are equal; **D.** Plane strain: the intermediate strain axis (Y) is unchanged, and all the strain is in the plane of X and Z. **E.** Progressive shear strain: in the XZ plane, the strain axes X and Z progressively rotate with increasing strain.

The **principal strain axes** define the shape of a **strain ellipsoid**, which represents deformation as if the original body were a **sphere**. The ellipsoid's shape depends on the ratio of its axes:

- **Prolate ellipsoid (rugby ball-shaped) → Extensional deformation** (maximum strain axis significantly larger than the other two, which are nearly equal).
- **Oblate ellipsoid (pancake-shaped) → Flattening deformation** (minimum strain axis significantly smaller than the other two, which are nearly equal).
- **Plane strain** → The **intermediate strain axis remains unchanged**, with all three axes being different.

The strain ellipsoid can also take **intermediate shapes** between these endmember types.

***Co-axial and rotational strain***

**Distortional strain** can be classified into two types:

- **Co-axial strain**: The **strain axes remain fixed** in orientation during deformation (Figures 4.5B–D). This is sometimes called **pure shear**, but the term can be misleading.
- **Rotational strain (shear strain)**: The **strain axes rotate progressively** during deformation (Figure 4.5E). Also called **non-co-axial strain** or **simple shear**, but "rotational strain" is the preferred term.

Since only the **final strain geometry (finite strain)** is observable, it may not be immediately clear whether deformation was co-axial or rotational. However, if the **original rock had planar features with different orientations**, the **finite strain pattern** can help determine the deformation type.

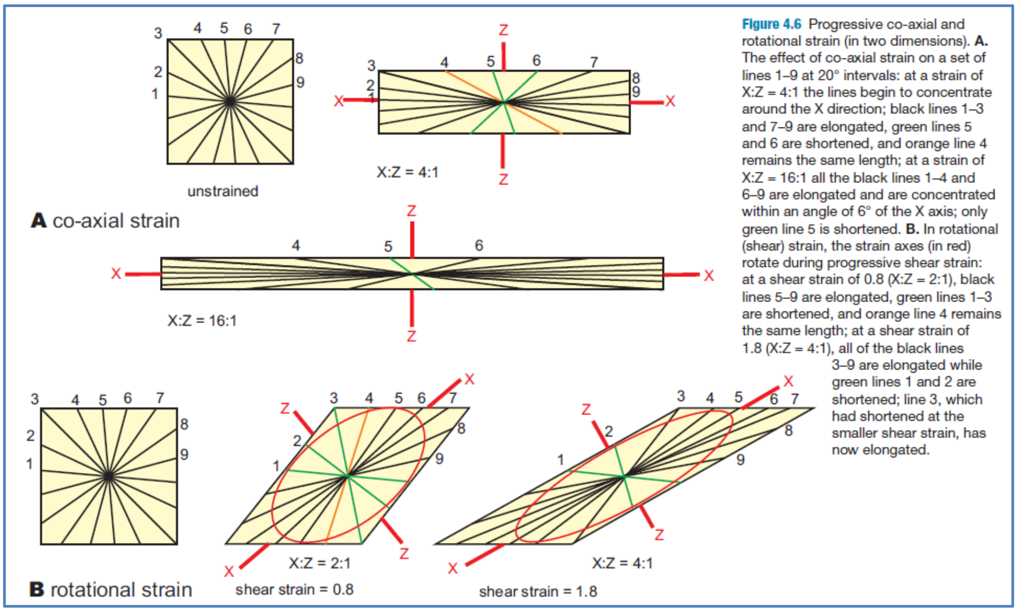


**Figure 4.6** Progressive co-axial and rotational strain (in two dimensions). **A.** The effect of co-axial strain on a set of lines 1–9 at 20° intervals: at a strain of X:Z = 4:1 the lines begin to concentrate around the X direction; black lines 1–3 and 7–9 are elongated, green lines 5 and 6 are shortened, and orange line 4 remains the same length; at a strain of X:Z = 16:1 all the black lines 1–4 and 6–9 are elongated and are concentrated within an angle of 6° of the X axis; only green line 5 is shortened. **B.** In rotational (shear) strain, the strain axes (in red) rotate during progressive shear strain: at a shear strain of 0.8 (X:Z = 2:1), black lines 5–9 are elongated, green lines 1–3 are shortened, and orange line 4 remains the same length; at a shear strain of 1.8 (X:Z = 4:1), all of the black lines 3–9 are elongated while green lines 1 and 2 are shortened; line 3, which had shortened at the smaller shear strain, has now elongated.

## *Geometrical features of progressive strain*

In **strained rocks**, pre-existing **planes or lines** undergo **progressive geometrical changes** as deformation increases. Depending on their initial orientation, some features **extend**, while others **contract**.

- **Co-axial strain (Figure 4.6A)**: Fields of **extension and contraction** remain symmetrical.
- **Rotational strain (Figure 4.6B)**: Extension and contraction fields shift, allowing determination of the **sense of rotation** (e.g., **dextral** (right-lateral) vs. **sinistral** (left-lateral) shear).

In cases of **large rotational strain**, some planes may **first contract and later extend**, which can be observed in deformed layers that were initially **folded and later stretched apart**.

## *The measurement of strain*

Due to the **variable physical properties** of rocks, strain measurements in one part of a rock body may not represent the **entire structure**. Instead of focusing on **precise measurements in a single location**, a **statistical approach** collecting multiple approximate strain measurements over an area is more effective.

However, an important limitation is that **only certain rocks contain identifiable features of known initial shape** that can be used to measure strain. These features are called **strain markers**.

The **best strain markers** are those that were originally **spherical** before deformation. These include:

- **Spherulites** in lavas
- **Ooliths** in limestones
- **Reduction spots** in slates

By analyzing how these markers have been **deformed into ellipsoidal shapes**, geologists can estimate the **strain** in the rock (Figure 4.7A).

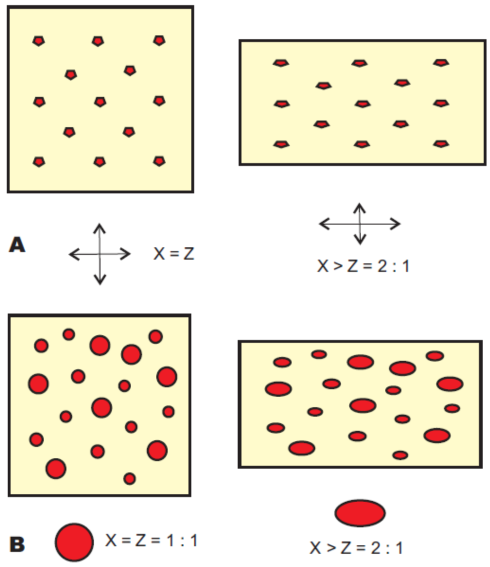


**Figure 4.7** Two easy ways to measure strain. **A.** In a rock containing evenly spaced objects (e.g. pebbles or phenocrysts), measured in the XZ plane, the mean separation between the objects in the X direction compared to the Z direction gives the X:Z strain ratio; a similar measurement in either the XY or YZ planes gives X:Y:Z. **B.** In a rock containing objects that can be assumed to have been initially spherical, or whose initial shapes had no preferred orientation, the mean X:Z ratio of the strained objects can be assumed to represent the X:Z ratio of the rock as a whole, provided that the strain in the matrix is equivalent to the strain in the objects.

The strain in the rock matrix is assumed to match that in the measured objects, which is valid only if both materials have similar strength. Pebbles in conglomerates are commonly used as strain markers, but their initial shape variability must be considered.

If the initial shape variation is random, meaning there is no preferred orientation, the method can provide a reasonable approximation of the overall strain. Grain aggregates and deformed phenocrysts in igneous rocks are also used as strain markers, as they tend to become ellipsoidal under moderate to large strains. Certain fossils can serve as strain markers as well, though the geometric calculations required can be complex and time-consuming.

An alternative approach that avoids measuring individual objects is to analyze the spacing between them. If their original distribution was random or uniform within the rock body, their spacing can provide an estimate of the strain (Figure 4.7B).

To measure strain, one method involves selecting an XZ plane and counting the number of intersections along a traverse in the X direction, then comparing it to the number in the Z direction. A similar approach in the XY or YZ planes allows for determining the strain ratio X:Y:Z.

A third technique applies when the original unstrained body contains linear or planar elements (e.g., elongate crystals in igneous rocks) with no initial preferred orientation. This situation, illustrated in Figure 4.6, shows that at a strain ratio of X:Z = 4:1, most lines are concentrated within 35° of the X axis. At X:Z = 16:1, nearly all lines fall within 22°. Thus, at large strains, the spread of orientations around the X axis can provide a rough strain approximation.

### *Structures produced by compression and extension*

When a rock body undergoes **compression** or **extension**, its deformation depends on the uniformity of its material properties.

- If the rock is **homogeneous**, deformation occurs **uniformly** (Figure 4.7).
- If the rock contains **stronger layers** within a **weaker matrix**, these layers respond differently to stress.

**Compression vs. Extension**

- **Under compression**, stronger layers tend to **contract** and form **folds** (Figure 4.8A).
- **Under extensional strain**, stronger layers may **break into segments**, separated by narrow necks, or become completely **isolated** within the weaker material.

This process is called **boudinage**, and the resulting segments, which may be **sausage-shaped** or **brick-like**, are known as **boudins** (from the French word for a type of sausage) (Figures 4.8B and 4.8C).

### *Transpression and transtension*

A block or layer of material may be subjected to compressional stress and shear stress simultaneously, in which case the stress state is termed **transpression**. Shear stress added to extension, is termed **transtension**.

***Strain and fabrics*** The effect of large strains, mainly on metamorphic rocks but also on some unmetamorphosed rocks, is to produce a set of new microstructural elements that are collectively known as the **fabric**.

Fabrics, consisting of foliations and lineations, provide insight into strain axes and, in some cases, the magnitude of strain. A foliation representing a flattening plane contains the principal strain axes X and Y, with Z perpendicular to it. Similarly, a lineation within the XY plane corresponds to the principal strain axis X, indicating the direction of elongation. By measuring the deformation of objects within these planes, the strain ratios X:Y:Z can be estimated.

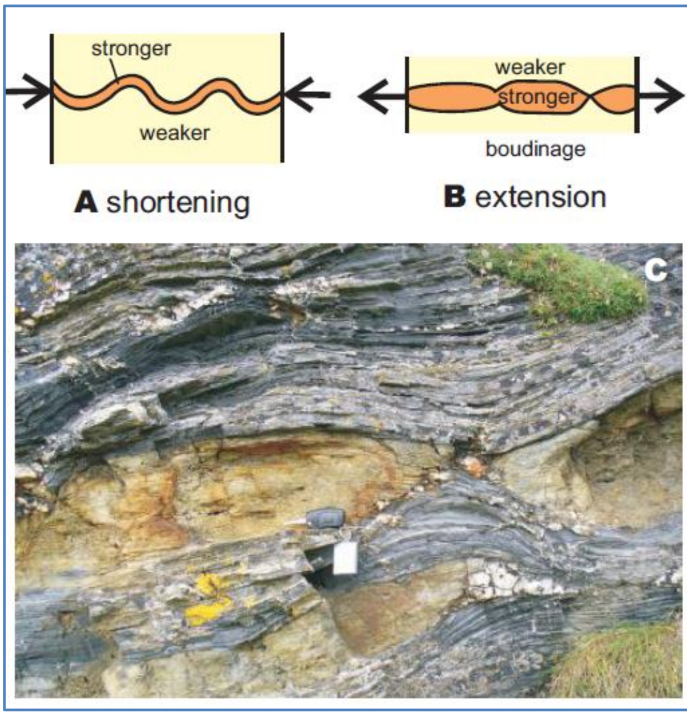


**Figure 4.8** Contraction and extension in layered rocks. Contraction of a strong layer within weaker material will produce folds (**A**), whereas extension may produce boudinage (**B**). **C.** Extension has resulted in the stronger, sandstone layer being pulled apart forming boudins separated by narrow 'necks'; the weaker shale layers have flowed into the areas between the boudins; see car key for scale.

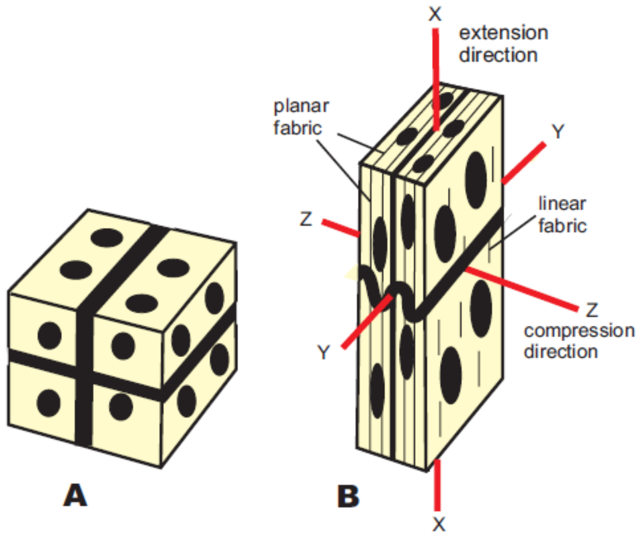


**Figure 4.9** Strain and fabric. Undeformed block (**A**) containing spherical objects is subjected to co-axial strain (**B**): planes parallel to the YZ strain axes are shortened (folded) while planes parallel to the XY strain axis are lengthened and flattened (to give a planar fabric, or foliation). The spheres are extended in the X direction and flattened in the XY plane. A linear fabric, in the form of an elongation lineation is parallel to X.

## Behaviour of materials under stress

The deformation of rock material depends on the interplay between its lithology (physical and chemical composition) and the environmental conditions (temperature and pressure) under which deformation occurs. Although it is difficult to imagine solid rock changing shape, laboratory experiments under high temperatures and pressures help simulate these conditions.

Additionally, conducting experiments over long durations allows for a more accurate approximation of geological timescales, as natural rock deformation often takes thousands to millions of years.

### *Elastic, plastic and viscous behaviour*

The best way to understand rock deformation is by first examining how familiar materials behave under stress. For instance, a stretched rubber band or spring demonstrates **elastic strain**, where the material returns instantly to its original shape when the stress is removed. In an **ideal elastic strain**, the extension of a spring is proportional to the applied force (Figure 4.10A).

Rock material can exhibit some degree of elastic strain when stress is initially applied. However, if the stress continues or increases, the strain behavior changes. Most rocks fracture after only a small amount of elastic strain. When stress reaches a certain threshold, materials—including rocks—undergo **plastic strain**, meaning the deformation becomes permanent and does not reverse when the stress is removed.

Plastic strain can be simulated by bending plasticine, putty, or a thin metal sheet. Initially, some elastic strain occurs, but with increasing stress or prolonged application, plastic deformation takes over. An **ideal plastic strain** can be illustrated by dragging a block of wood across a rough surface, where resistance must be overcome before movement occurs (Figure 4.10B).

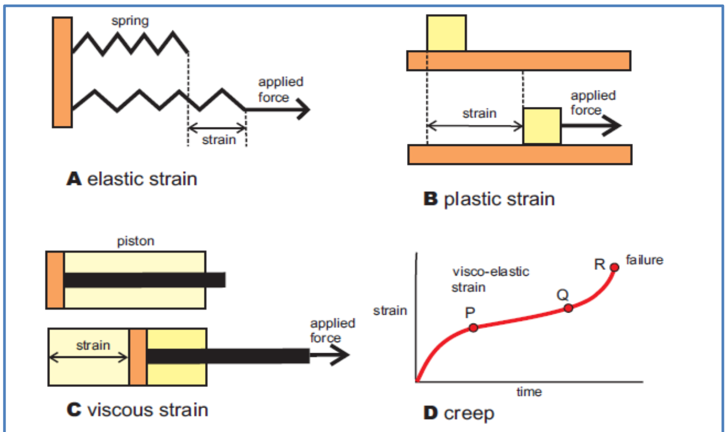


**Figure 4.10** **A.** Ideal elastic strain is represented by an applied force extending a spring; the amount of strain equals the amount by which the spring has been extended; there is a steady (linear) increase in strain with increase in stress and a steady increase in strain with time at constant stress. **B.** Ideal plastic strain is represented by a block being pulled by an applied force along a rough surface, opposed by sliding friction; the strain is represented by the distance travelled by the block; the frictional resistance represents the yield stress; the strain increases without any increase in stress. **C.** Ideal viscous strain is represented by a piston, drawn through a cylinder by an applied force, the strain being represented by the distance travelled by the piston; the force is opposed by the resistance caused by the viscosity of the fluid in the cylinder; there is a steady (linear) increase of strain-rate with stress and strain with time, at constant stress. **D.** Visco-elastic behaviour and creep: strain-time graph showing a more realistic curve of variation of strain with time, known as a creep curve; initial visco-elastic strain is followed at P by a period of steady-state strain increase (viscous) followed in turn at Q by a period of accelerating strain ending in failure (R).

Steady **plastic strain** occurs once an initial force overcomes friction, allowing the block to move at a constant speed. The **strain amount** corresponds to the distance traveled. The **initial force** required to move the block represents the **yield strength** of a material—the stress level at which **permanent strain** begins. In **ideal plastic strain**, this force remains constant; applying a larger force would accelerate deformation, eventually leading to **failure**.

The concept of **viscosity** is usually applied to the behaviour of liquids although, by analogy, it can be extended to rock materials undergoing solid-state flow. The term **viscosity** is defined as the *rate of flow* of material subjected to a stress; in the case of liquids, it is measured by the rate of flow through a narrow tube, subjected, for example, to gravitational force. Ideal viscous behaviour can be represented using the familiar analogy of the piston, as in the shock absorber of an automobile (Figure 4.10C). Here, the rate of flow (i.e. the *strain* *rate*) is proportional to the magnitude of the stress; in viscous strain, therefore, it is the *rate of increase* of the strain that is proportional to the size of the stress.

Since viscosity is measured as rate of flow (or strain rate) divided by stress, the larger the value of the viscosity, the more slowly the material deforms. Thus a material with a high viscosity, like rock, deforms more slowly than one of low viscosity, such as oil.

The terms plastic and viscous are commonly used interchangeably although, strictly speaking, plastic strain in a given material should exhibit a single strain rate, whereas with viscous strain, a range of strain rates is possible, depending on the size of the stress.

Real rock material deforms usually in a more complex way than these idealised elastic, plastic and viscous models. A closer approximation to the behaviour of real materials is the **visco-elastic** model, which combines an initial period of elastic strain with a period of viscous strain (Figure 4.10D).

To simulate the geological conditions under which permanent strain occurs in rocks, such as in the formation of folds, a smaller stress has to be applied for a very long period of time. The resulting variable visco-elastic behaviour is usually known as **creep**.

Figure 4.10D shows a typical creep curve of the kind that represents the deformation of real rock materials in laboratory experiments. In this type of behaviour, an initial short period of visco-elastic strain is followed by a much longer period of steady plastic or viscous strain, which may end, after a period of accelerating viscous deformation, in failure.

Under geological conditions, the behaviour of rocks will generally fall into two categories: high values of stress lead to an accelerating strain rate and failure after relatively short periods of time, whereas low values of stress lead to long-term, steady, viscous creep at low strain rates.

It is possible to demonstrate creep behaviour of rock in a relatively short time span by suspending a thin slab of rock, such as sandstone, between two supports at each end of the slab.

After a period of perhaps months, or even years, depending on the strengthof the slab, it will bend downwards and eventually fracture under constant gravitational force. If the amount of downwards displacement (i.e. the strain) is measured over time, it should correspond to a typical creep curve.

***Brittle and ductile behaviour***

Materials that fail (fracture) after there has been no, or very little, plastic or viscous deformation when a stress is applied are said to be **brittle**, whereas those that experience considerable plastic or viscous flow are said to be **ductile.** These terms are relative and somewhat subjective, and materials that are brittle at low temperatures become ductile at higher temperatures. It follows that, in general terms, brittle behaviour characterises faulting and ductile behaviour, folding.

***Effects of temperature, lithostatic pressure and pore-fluid pressure***

An increase in temperature has a marked effect on the way a rock deforms, since it decreases the yield strength of the material, so that permanent deformation will commence at a lower stress and the strain rate will increase for a given stress level. Thus an increase in temperature makes a rock more ductile.

Increasing the **lithostatic pressure**, however, has the opposite effect, by raising the yield strength of the material, and thus requiring a larger directed stress to achieve permanent deformation.

The effect of raising the lithostatic pressure can be counteracted by the effect of pore fluids. In a rock with a high proportion of pore fluid, the pressure of this fluid (the **hydrostatic pressure**) will approach that of the lithostatic pressur since the fluid is subject to the same gravitational pressure as the surrounding rock).

Thus in rocks at a considerable depth in the crust, the **effective pressure** consists of the lithostatic pressure minus the **pore fluid (hydrostatic) pressure**. This means that, at a given temperature, the ductility of a rock will depend critically on the pore-fluid pressure.

The yield strength of rocks is thus dependent both on temperature and effective pressure, but the effect of temperature becomes more important at greater depth. Consequently, rock in general will become stronger with depth in the crust down to a critical level at which the strength reaches a maximum value, below which it decreases as the temperature increases. This level willbe reached at different places with different rocks, but in general will lie at mid-crustal levels.

Here there is a transition between broadly brittle behaviour and broadly ductile behaviour in a zone known as the **brittle–ductile transition**, usually regarded as between 15 km and 20 km depth in normal continental crust, below which earthquakes are not usually generated.

An alternative way of looking at this is to compare the differential stress required to maintain a geologically realistic strain rate with increasing depth as shown schematically in Figure 4.11. Studies of real rocks in laboratory experiments indicate that the variation in the yield strength is significantly affected by important changes in the mechanism of deformation, which we shall now discuss.

## How rocks deform

The various models of deformation discussed above treat rocks as uniform materials in analysing their response to stress. However, to understand their behaviour in practice, it is necessary to examine rocks in detail, on a small scale, especially under the microscope.

Since rocks are aggregates of crystals or grains, in most cases involving several different mineral types, their deformation will depend on the various responses of the different minerals to stress.

Another important factor is the influence of the temperature pressure environment during the deformation, since this affects the way in which individual minerals deform. The behaviour of the weakest minerals in a deforming rock is critical, since this controls the strain of the whole rock; in the case of many rocks, this mineral is quartz; in others it is feldspar, or calcite; and in the case of mantle materials, it is olivine; and much experimental work has been done on these minerals.

Unfortunately, the short time periods available for experimental deformation means that the strain rates achieved during this deformation are much too high (around 10-7 per second, equivalent to 300% per year), whereas we know from plate tectonic velocities that strain rates for lithospheric deformation should be in the range 10-12 – 10-15 per second, or 100,000 to 10 million times slower! Nevertheless, the correspondence between the types of crystal behaviour and rock fabrics seen in these experimentally deformed materials and those characterizing natural rock deformation suggests that they do provide a useful guide.

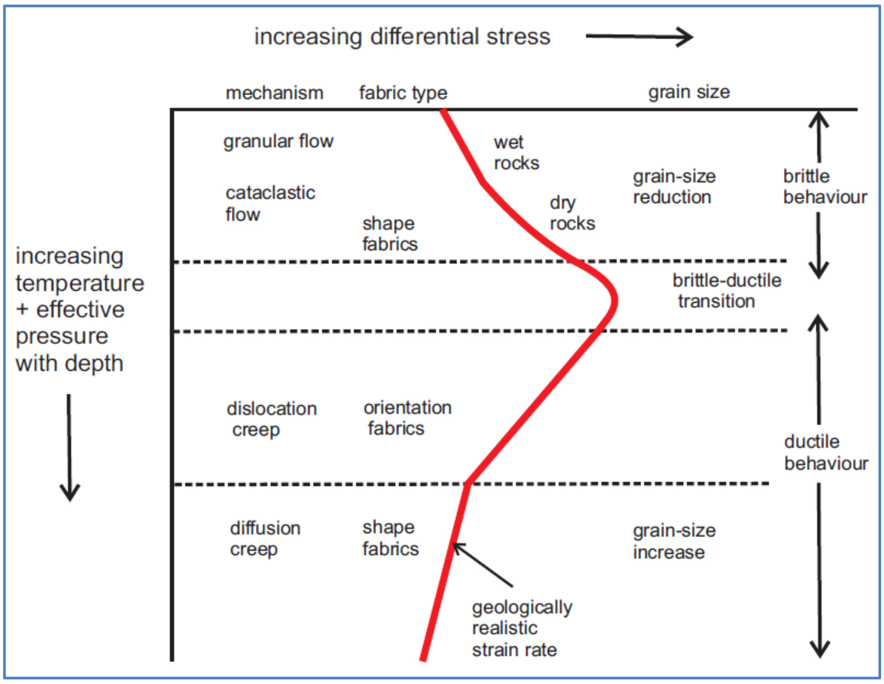


**Figure 4.11** Schematic deformation diagram showing the fields occupied by the various types of viscous permanent strain associated with increasing depth (i.e. increasing temperature and effective pressure) and increasing levels of differential stress. The red line represents the behaviour of an imaginary rock material deforming at a geologically realistic strain rate (usually assumed to be around $10^{-14}$ per sec). At low temperature and high differential stress, viscous deformation takes place by granular flow or, with increasing lithostatic pressure, by cataclastic flow. The strength increases with depth to a maximum at the brittle-ductile transition. Above this level is brittle behaviour, dominated by fracture processes and leading to grain-size reduction. At higher temperatures, lower stresses are required to produce the equivalent strain rate. The mechanisms here are dislocation creep and, at higher temperatures again, diffusion creep. This is ductile behaviour, leading to grain-size increase.

## *Elastic behaviour*

Temporary or elastic strain in a crystalline solid is achieved by distortion of the **crystal lattice**, which is the molecular framework of the crystal (Figure 4.12A). When the stress is removed, the crystal returns to its original shape, but if the stress is maintained for too long, or is increased, the distortion may become permanent and the material begins to behave in a ductile manner.

## *Permanent strain*

Ductile deformation, or **creep**, in rocks is achieved by means of several different mechanisms that produce changes in shape (strain) within the rock that are not recoverable. There are a number of factors that determine which mechanism is chosen in a particular case;

these include the nature of the rock (lithology, grain size etc.), the temperature and lithostatic pressure, the pore fluid pressure, and the differential stress.

Sedimentary rocks near the surface will usually deform by means of **grain boundary** sliding, since the grain boundaries are the weakest part of the rock. In unconsolidated material such as beach sand, for example, this behaviour is obvious, as the grains will merely roll past each other under stress (**granular flow**).

In sedimentary material where the grains are cemented together, and in crystalline rocks, small fractures along the grain boundaries will enable the grains to slide past each other. In this process, the grains will be internally undeformed. Under higher stresses, the grains themselves may be fractured.

Creep deformation that is dominated by fracturing is termed **cataclastic flow**. This process produces fabrics characterised by angular-shaped grains and a progressive decrease in grain size. Alignment of elongated grains or grain aggregates may give rise to a **shape fabric**. Under higher pressure and temperature, fracturing is replaced by a process termed **dislocation creep**, whereby the crystals themselves are internally deformed (e.g. by **dislocation gliding** or **twin gliding**) so as to achieve a different shape compatible with the overall strain (Figure 4.12B, C).

Various features such as **deformation twins** and **kink bands** characterise crystals deformed in this way. Dislocation or twin gliding accompanied by rotation can orient a set of crystals in such a way as to produce the maximum strain effect, and leads to a crystal orientation fabric (Figure 4.12D). These changes in crystals may be accompanied by recrystallisation as the strain progresses, and will be aided by higher temperatures.

Crystalline metamorphic rocks deformed in this way will be characterised by a **preferred orientation** of the crystallographic axes, since the newly shaped crystals will tend towards an orientation such that their crystallographic slip planes are aligned favourably with the stress axes. This is termed an **orientation fabric**. Original grains may survive, surrounded by small re-grown grains, producing a characteristic texture.

With further increase in temperature and pressure, another mechanism becomes dominant. Here, material is transferred by diffusion through the crystal lattice or along grain boundaries from areas of high compressive stress to areas of low stress either in the solid state (**diffusion creep**) or by solution and redeposition from a fluid (**pressure solution** or **solution creep**) (Figure 4.12E, F).

Ultimately, complete recrystallisation of the rock may take place, leading generally to an increase in grain size. This process will tend to form a shape fabric dominated by elongate crystal shapes that serve to define the new strain geometry; it may or may not produce an orientation of the crystal axes. Solution creep is the typical mechanism in low-grade metamorphic rocks where fluid is available, whereas diffusion creep dominates in higher-grade metamorphic rocks.

Figure 4.11 shows schematically how these various mechanisms relate to variations in differential stress and depth for geologically realistic strain rates.

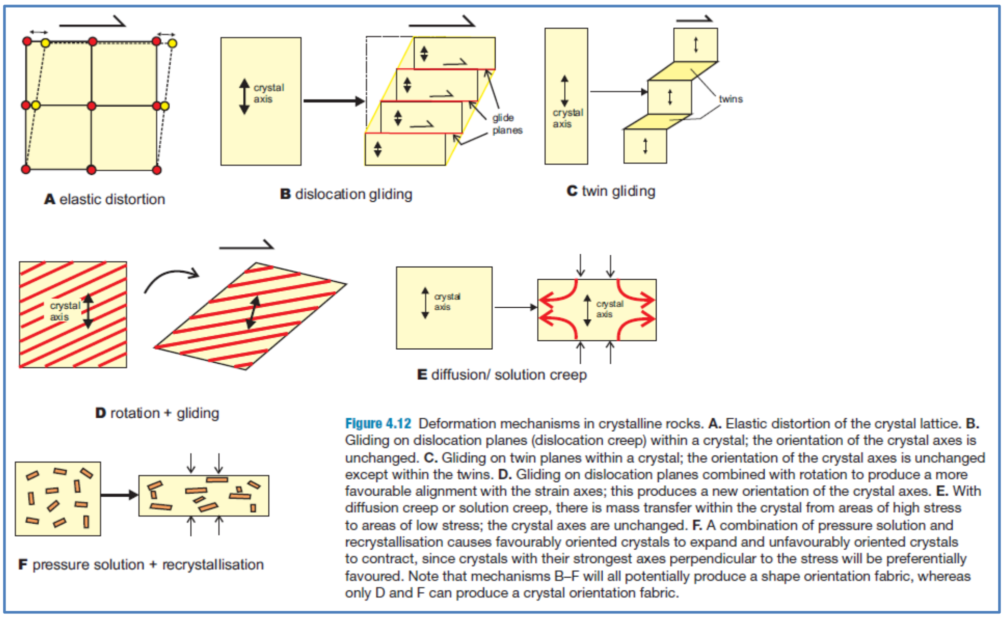


**Figure 4.12** Deformation mechanisms in crystalline rocks. **A.** Elastic distortion of the crystal lattice. **B.** Gliding on dislocation planes (dislocation creep) within a crystal; the orientation of the crystal axes is unchanged. **C.** Gliding on twin planes within a crystal; the orientation of the crystal axes is unchanged except within the twins. **D.** Gliding on dislocation planes combined with rotation to produce a more favourable alignment with the strain axes; this produces a new orientation of the crystal axes. **E.** With diffusion creep or solution creep, there is mass transfer within the crystal from areas of high stress to areas of low stress; the crystal axes are unchanged. **F.** A combination of pressure solution and recrystallisation causes favourably oriented crystals to expand and unfavourably oriented crystals to contract, since crystals with their strongest axes perpendicular to the stress will be preferentially favoured. Note that mechanisms B–F will all potentially produce a shape orientation fabric, whereas only D and F can produce a crystal orientation fabric.