3 Stress

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Abstract

A layman may laugh at you when you say that a rock feels the pinch of a force imposed on it! Yes, it does feel irrespective of the amount of force. In fact, the rock gets disturbed when a force is imposed upon it. The disturbance thus developed in a rock is called *stress* which is expressed as the force acting per unit area of a rock. If the applied stress persists uniformly till its amount exceeds the strength of the rock, the latter undergoes deformation, thus developing *strain*. Depending upon the specific conditions in the earth's crust, stress can be of various types such as hydrostatic stress, differential stress, deviatoric stress and lithostatic stress. The stress that is locked in the rocks when they were formed in the geological past is called *palaeostress* that has implications for the amount of deformation (strain) in rocks as well as for the directions in which the stresses had acted upon the rocks. Precise knowledge of the *present-day state of stress* inside the earth's surface is important in geology and in engineering geology, especially for various engineering and mining projects as well as for our preparedness for earthquakes. This chapter highlights some aspects of stress as relevant to structural geology.

Keywords

Stress · Principal stress axes · Uniaxial stress · Biaxial stress · Stress ellipse · Stress ellipsoid · Mohr stress diagram · Palaeostress · Stress tensor · Stress inside the earth

3.1 Introduction

Rocks in the earth's crust are under the burden of the overlying rocks as well as of their own weight. A rock mass is thus always being affected by externally applied forces. These forces develop *stress* within the rock. If the amount of stress developed in a rock exceeds its strength, the rock then yields and undergoes deformation; that is, it changes its shape and/or size to accommodate the induced stress. Although the rock deforms due to stress, the primary cause of stress is the forces that have acted upon the body. So, any study on stress should also take into account the nature of forces that act upon a body. Estimation of stress in rocks is a complicated subject. Although estimation of stress in rocks located below the ground surface up to depths of a few hundreds of metres is possible, it would give the present-day or contemporary stress only and not the actual stress that was operative when the rock was formed or deformed. Study of stress is important especially for getting an idea of direction of application of forces during deformation as well as in the estimation of strain in rocks. In this chapter, we shall discuss some basic concepts of stress in rock that are necessary for structural studies.

3.2 Force

A force is that which changes or tends to change the state of rest or of uniform motion of a body. The force that acts upon a body is applied from some external agency that either makes or tends to make the body move from its original position or stops or tends to stop a moving body. In the first case, the body is said to have been affected by an acceleration, which is measured as the force acting per unit time and tends to increase the motion of the body. A force is best expressed by *Newton's second law of motion*, i.e.

$$F = ma \tag{3.1}$$

where F is a force, m is the mass of a body and a is the acceleration produced in the body due to the force F. In the second case, the body is said to have been affected by a retardation, which is the force acting per unit time and tends to decrease, and finally stops, the motion of the body. A force is a *vector quantity*, while the mass is a *scalar quantity*. A vector quantity has a magnitude and a direction such as weight, velocity and gravity, while a scalar quantity has a magnitude only such as mass, speed and heat.

Considering a rock mass as a deformable solid or fluid, the forces acting upon it can be of two types, *body forces* and *surface forces*.

1. **Body forces** act upon a body and are therefore given by the volume or size of the body. Such forces act on every point of the body and are therefore measured in three dimensions. Thus, the bigger a body is, the larger its body force is such as the planetary bodies. Common examples of body forces include gravity and magnetic force. Effect of gravity produces weight to a body.
2. **Surface forces** act upon the boundaries of a body, i.e. along the area where the forces are in contact with the body. If we push a small wooden block by hand, the block moves for a distance. The distance moved is dependent upon the forces that the block has received along the contact area only. Further, instead of being displaced, a body, say a rock mass, may get deformed. In this case, the external forces have developed stress in the rock mass, and in order to accommodate the internally developed stress, the rock mass gets deformed. The deformation is thus an expression of the stress that the rock mass has developed. Surface forces are therefore relevant to structural geology

because rocks deform by the application of some external forces along a contact plane.

3.3 Types of Force

In structural geology, contact forces are more common, while the direct role of force at a distance has not yet been significantly highlighted and can thus be considered negligible. Balanced forces cause stress, while unbalanced forces produce stress plus displacement. The common types of forces (Fig. 3.1) are briefly described below.

Confining pressure (Fig. 3.1a) at a point is one that is exerted equally from all sides. In the earth's crust, the confining pressure is the lithostatic pressure developed due to the load of the overlying rocks. Since this pressure is equal on all sides of a body, it can be compared with the hydrostatic pressure produced by fluids. In the context of the rocks in the earth's crust, use of confining pressure is more common. *Shear* (Fig. 3.1b) can be compared to the effect produced when the cards in a pack are displaced past each other. Development of shear in a rock takes place by the formation of a couple, also called shear couple, caused by relative movement of rock material past one another. Shear can be produced by a shear stress. *Bending* (Fig. 3.1c) involves deformation of a beam when a load is applied across it. The beam deforms plastically. Bending occurs when the bending stress exceeds the yield strength of the material but below the ultimate strength. Because of the bending load, both compressive and tensile stresses develop in the beam. These stresses are of opposing nature (depending on, say, whether a fold is an anticline or a syncline), and as such there is a neutral surface in the beam where there is no bending stress. Depending upon the direction and location of the transverse stress, the beam assumes a variety of shapes after deformation. In structural geology, bending mainly has implications for fold mechanics. Some of the folds in crustal rocks have been interpreted to have formed by bending (see Chap. 8). *Torsion* (Fig. 3.1d) involves deformation of a body when it is twisted by applying a stress in one direction at one end while the other end remains either motionless or twisted in opposite direction. In other words, torsion involves application of couples that act in parallel planes but in opposite directions and about the same longitudinal axis of rotation. In crustal rocks, situations favouring the development of torsion practically do not exist. As such, torsion as a stress in rocks is an unlikely process. A torque (Fig. 3.1e) is a force that tends to rotate a body. A *torque* may increase or decrease the speed of rotation of the body. Since rotational forces seldom develop in rock masses of the earth, application of torque in structural geology is therefore negligible to absent.

3.4 Stress

Force acting upon a body or a rock develops stress within it. The body is then said to be under stress due to the applied force. If F is the amount of force acting upon a unit area A of a body, then the stress (σ) developed in the body due to this force is given by

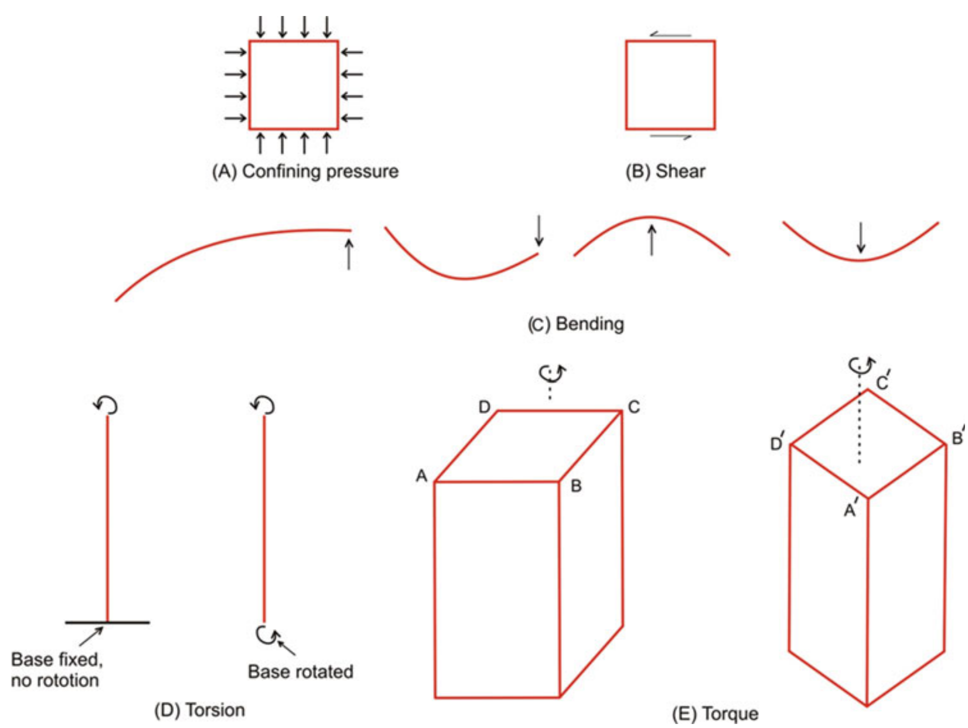

Fig. 3.1 Common types of forces. (**a**) Confining pressure. (**b**) Shear. (**c**) Bending. (**d**) Torsion. (**e**) Torque. (See text for details)

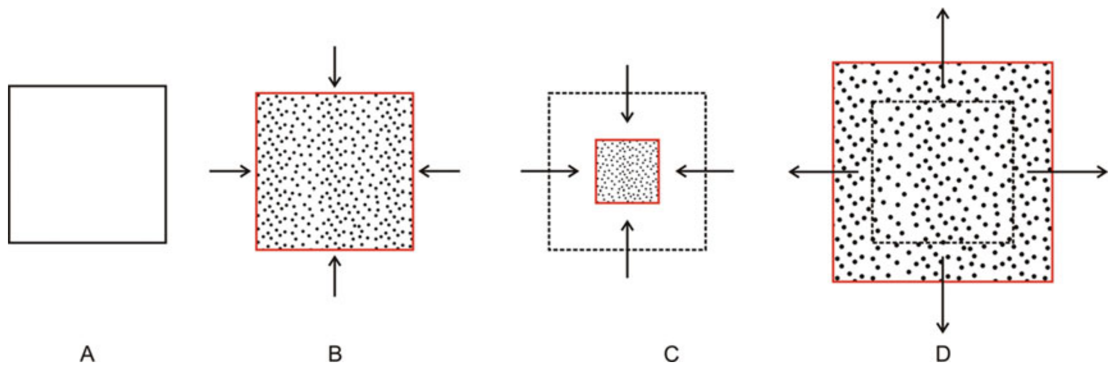

Fig. 3.2 Stress acting upon a body as explained diagrammatically. (**a**) A body without any externally applied stress continues to remain in its original shape. (**b**) The body under the influence of equal amount of stress acting from all sides develops stress within it. (**c**) If the applied stress persists uniformly and its amount exceeds the strength of the body, the latter undergoes dilatation. Since the stress is compressive, the body shows reduction of volume, which in two dimensions causes reduction in area as shown in **c**. (**d**) If the applied stress is tensile, the body would show increase of volume, which in two dimensions causes increase in area as shown in **d**

$$\sigma = F/A \quad (3.2)$$

Unlike force, a stress has no directional significance but has a point of application. The stress developed in a rock interacts with the constituents of rock and tends to develop physical change or deformation (Fig. 3.2) to the rock, which we call *strain*. Thus, stress causes strain in a rock. If stress is the cause, then strain is the effect. Obviously, if a rock is not subjected to any stress, it continues to remain in its original shape and volume (Fig. 3.2a). If the body is under the influence of equal amount of stress acting from all sides, it is said to be under stress (Fig. 3.2b). If on the other hand the stress overcomes the strength of the rock, the latter succumbs to the stress and shows changes in its shape or volume, and we then say that the rock is strained or deformed. A stress will cause a change in shape as shown in Fig. 3.2c (compressive stress) and in Fig. 3.2d (tensile stress).

In the above definition of stress, we have assumed that the distribution of stress is uniform within the body. In practical sense, it is generally not so, and the stress developed in a body due to a force F may vary from cross section to cross section. In such case, the stress at a point can be given as

$$\sigma = \lim_{\Delta A \to 0} \Delta F/\Delta A \quad (3.3)$$

It may be noted here that although the above two equations, i.e. Eqs. (3.2) and (3.3), provide values of stress, there are subtle differences. The former equation provides the value of stress developed in a body due to a force F assuming that the distribution of stress is *uniform* within the body. The latter, on the other hand, gives value of stress at a point assuming that the distribution of stress, in practical sense, is *not uniform*.

The stress that causes deformation to a rock can be estimated for many minerals via experimental mechanics. The study and analysis of stress help in throwing light, among others, on the directions of the applied forces and explaining the geometry, stages and pattern of deformation of a rock mass. *Stress analysis* helps in the designing of structures such as dams and tunnels. It is also used in several geological phenomena such as plate tectonics, earthquakes, volcanism and landslides.

3.5 Units of Stress

The unit of stress can be derived from the unit of force. In the International System of Units (SI), the unit of force is newton (N). One newton is defined as the force required to produce an acceleration of 1 m/s to a mass of 1 kg. In the CGS system, the unit of force is dyne; 1 dyne = 10^{-5} N. The most commonly used unit of pressure is *bar* and that of stress is *megapascal* (MPa). Atmospheric pressure is expressed in *pascal*. On the surface of the earth, 1 atmospheric pressure is equal to 100,000 Pa, which for the sake of simplicity is taken to be equal to 1 bar. 1 MPa is equal to 10 bars, and so 100 MPa is equal to 1000 kilobars or kba. Thus, any further conversions can be easily made.

3.6 Tensile Stress and Compressive Stress

In stress analysis, it is important to know the orientation of stress with respect to the body. We can imagine a plane upon which a stress is being applied at a point. This enables us a two- or three-dimensional study of stress, which can be of three important types (Fig. 3.3): (a) *Tensile stress* (Fig. 3.3a) that develops when the stress acts along a plane in opposite directions when the body is in equilibrium: as a result, the

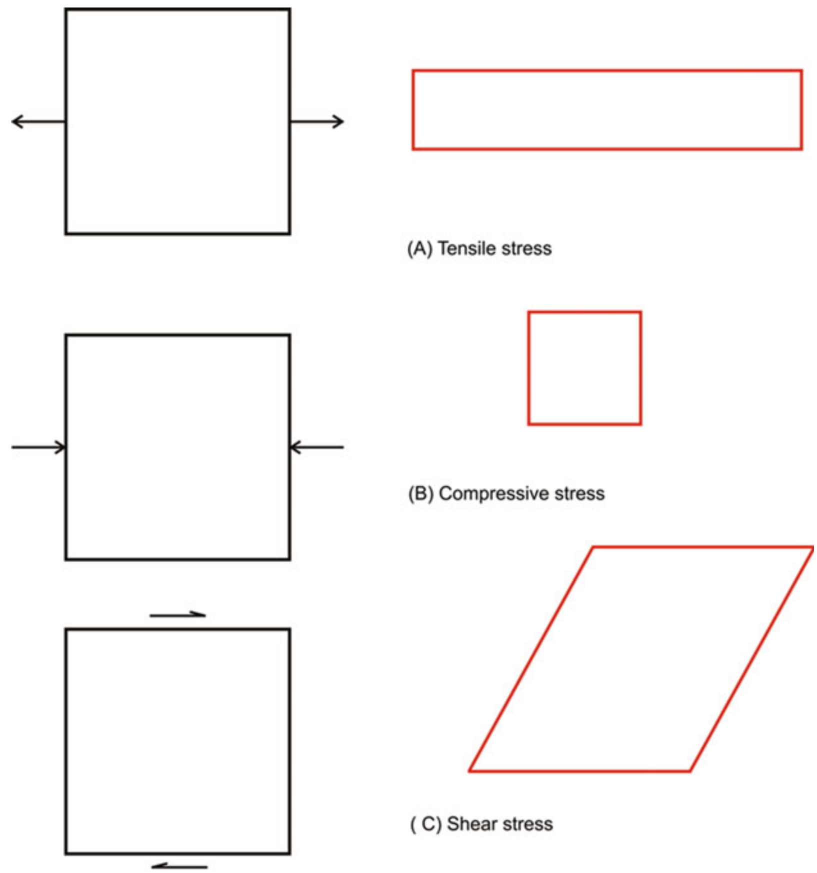

Fig. 3.3 Diagrammatic sketches to show the three common types of stresses. (**a**) Tensile stress acting in opposite direction; the body gets stretched in the direction of stress. (**b**) Compressive stress acting towards each other; the body gets shortened. (**c**) Shear stress. A square changes to a rhomb

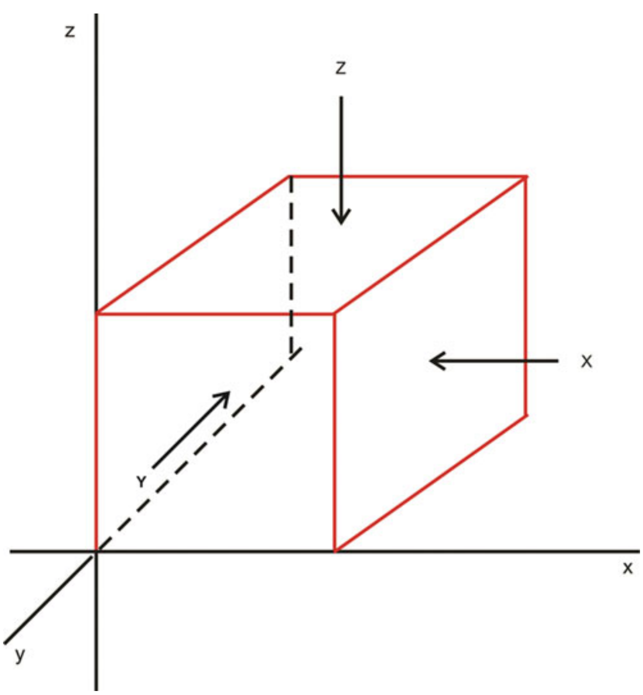

Fig. 3.4 Principal stress axes. Assuming that the cube is under stress from all directions, we can imagine that all the stresses are represented by only three stresses acting parallel to each of the coordinate axes x, y and z. Each of these stresses is called a principal stress. Thus, the stress acting parallel to x-axis is designated as X, while the others are Y and Z

body gets stretched. Tensile stress is assigned a negative (−) sign in structural geology. (b) *Compressive stress* (Fig. 3.3b) that develops when the stress acts perpendicular to the plane when the body is in equilibrium: it means that an equal and opposite stress has been active. As a result, the body gets shortened. Compressive stress is assigned a positive (+) sign (because in geology tensile stress is least likely to exist in nature). Since each of the above two types of stresses, i.e. compressive and tensile, acts perpendicular to the plane of the body, each is called direct stress or normal stress. (c) Shear stress (Fig. 3.3c) that develops when the stress acts in opposite directions but not along the same plane or line: as a result, the body shows angular changes of its original planes and thus changes its shape. A square for example is changed to a rhomb, while a rectangle is changed to a parallelogram.

3.7 Principal Stress Axes

Analysis of stress becomes easy if we imagine that all the stresses acting upon a body are represented by only three stress components acting parallel to each of the coordinate axes x, y and z (Fig. 3.4). The normal stress acting parallel to each axis is called a principal stress. The principal stress is normal stress acting on a plane that has zero shear stress. Of these three stresses, one is considered to take a maximum value, while others take a minimum and an intermediate value. The coordinate axes along which the stresses act are called *principal axes of stress*. The stress acting parallel to *x*-axis is designated as X, while the other two stresses are designated as Y and Z. This convention enables to know the orientation of any principal stress in the coordinate system. It is also assumed that shear stresses acting along each of these principal axes of stress are zero. As such, each principal stress represents the normal stress.

3.8 Uniaxial Stress

If a stress acts along a linear body (Fig. 3.5a), say a rod or a bar, we say that the body is subjected to forces that act along one direction only, i.e. along its length. In this case, two of the three principal stresses are not acting and can therefore be considered to be of zero value. This type of stress is called a *uniaxial stress*, and the bar is said to be under uniaxial stress. If σ_1 is the principal stress acting along the length of the body, then $\sigma_1 \neq 0$, $\sigma_2 = 0$ and $\sigma_3 = 0$. If the amount of force is *F* and the cross-sectional area of the rod is *A*, the uniaxial stress thus developed in the rod is given by $\sigma = F/A$. If the forces act in opposite directions along the length of the rod, the latter is said to be under uniaxial tensile stress (Fig. 3.5b) that causes increase of length of the rod. With forces acting towards each other, the rod is said to be under uniaxial compressive stress (Fig. 3.5c) that causes shortening of length of the rod. In structural geology, according to the *convention of sign* as mentioned above, the tensile stress is assigned a negative (−) sign while the compressive stress a positive (+) sign.

3.9 Stress Ellipse

It is possible to know the two-dimensional stress at a point if the normal stress (σ_n) and shear stress (τ) are known. In this case, these two stress components act on a plane of any orientation passing through the point. An infinite number of planes are thus possible. We can however consider a simpler case in which all the normal stresses acting on the planes are represented by either a compressive stress or a tensile stress. In such case, if we represent all the stresses acting on the body by lines, the longest and shortest lines trace an ellipse (Fig. 3.6) called the *stress ellipse*. This ellipse can be traced if two of the stresses acting perpendicular to each other are

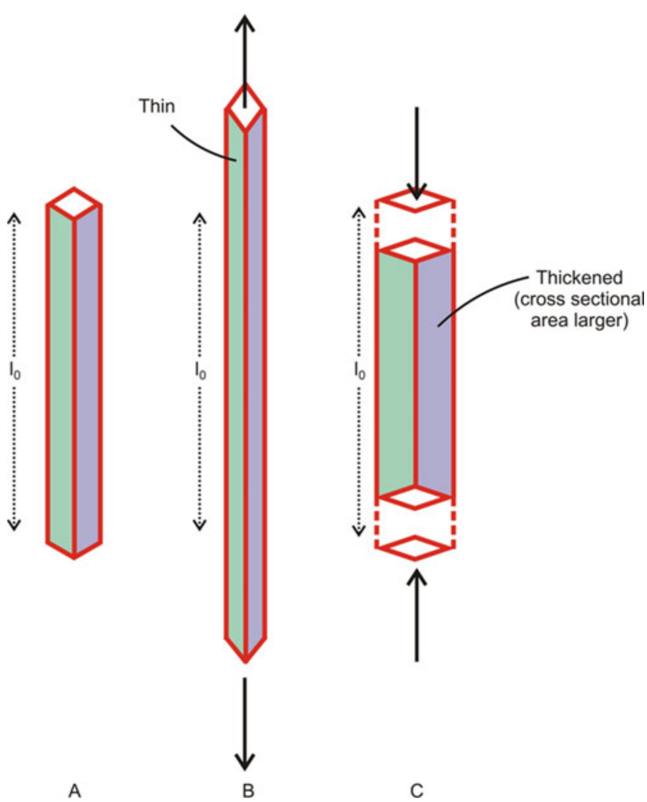

Fig. 3.5 Uniaxial stress showing stress acting along the length of a bar. (**a**) Original length (l_o). (**b**) The bar is under uniaxial tensile stress causing increase of its length. (**c**) The bar is under uniaxial compressive stress causing shortening of its length

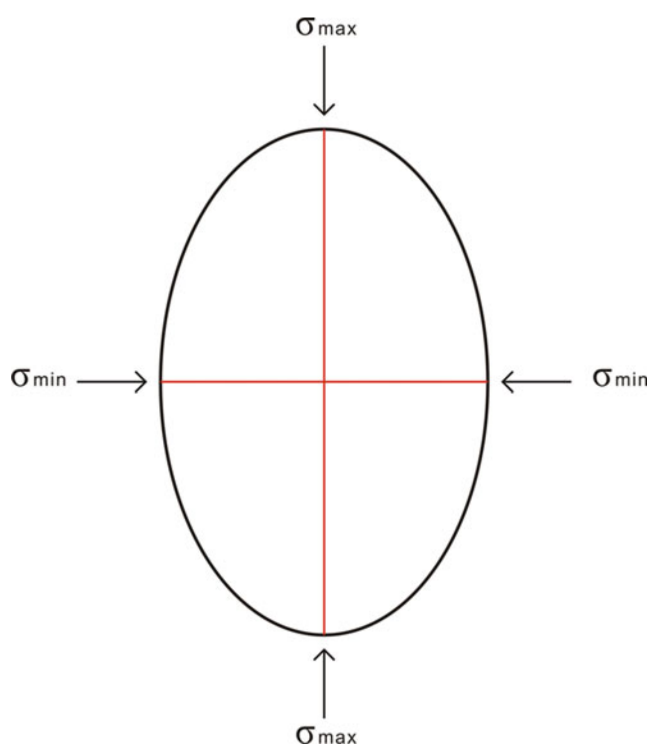

Fig. 3.6 Stress ellipse. The long and short axes represent the maximum stress and the minimum stress, respectively. The stresses are compressive in this case but can be tensile also

known. *A stress ellipse is thus a graphic method of showing relationships between the maximum and the minimum stresses*.

3.10 Biaxial Stress

When all the stresses act upon a body, the state of stress is known as *biaxial stress* or *two-dimensional stress*. A biaxial stress system is represented in two directions: a normal stress and a shear stress. Therefore, of the three principal stresses, one is equal to zero, and as such the biaxial stress is represented by

$$\sigma_1 \neq 0, \sigma_2 \neq 0, \sigma_3 = 0$$

A biaxial stress state in a body is thus developed when it is subjected to only two stresses σ_1 and σ_2 acting on its plane while the normal stress σ_3 is zero. Since all the stresses act in one plane, the biaxial stress is also called *plane stress*.

Application of biaxial stress in crustal rocks is as yet a matter of debate. As such, the concept of biaxial stress in structural geology is limited to geomechanics only, i.e. laboratory testing of materials mainly for engineering purposes.

3.11 Biaxial Stress on a Plane

We now consider a case when a force (F) acts at an angle to the plane (Fig. 3.7a). In this case, the force can be resolved into two components: (a) *direct* or *normal stress* that acts perpendicular to the plane and is denoted by σ (sigma) and (b) *shear stress* that acts parallel to the surface and is denoted by τ (tau) (Fig. 3.7b). Since the body is in equilibrium, we can imagine that σ is being affected by an equal and opposite component N. The shear stress τ can further be resolved into two components τ_1 and τ_2 at right angles to each other but in the same plane (Fig. 3.7c). Thus, we have resolved a force F acting on a plane at a point P into three stresses σ, τ_1 and τ_2.

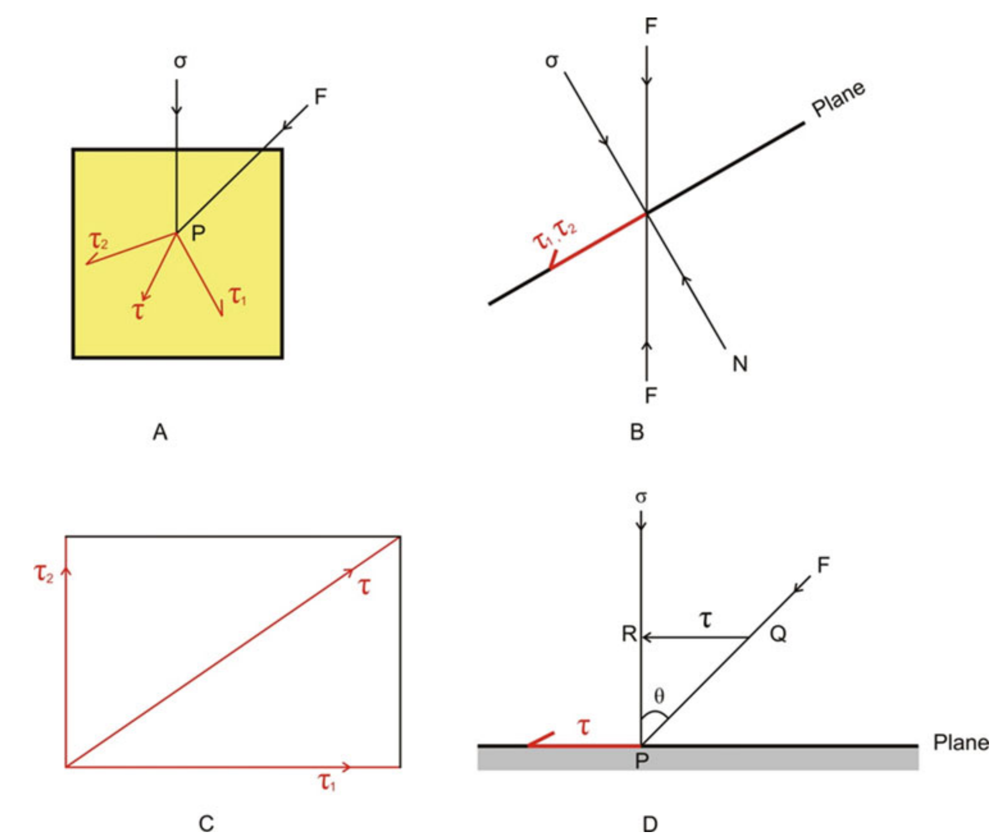

Fig. 3.7 Biaxial stress acting on a plane. (**a**) A force F acting at some angle to the plane is resolved into two stresses: one acting perpendicular to the plane, called normal or direct stress (σ), and the other acting parallel to the plane, called shear stress (τ). On the plane, τ can further be resolved into two mutually perpendicular stresses τ_1 and τ_2. (**b**) Another way to show the perpendicular position of the normal stress to the plane while the two shear stresses τ_1 and τ_2 remain along the plane. If the body is in equilibrium, the normal stress σ is subjected to an equal and opposite stress N acting at the same point. (**c**) The shear stress τ can be resolved into two mutually perpendicular stresses τ_1 and τ_2 by the law of parallelogram. Thus, the force F has been resolved into three mutually perpendicular stresses σ, τ_1 and τ_2. (**d**) The stress system shown in B is reproduced in a simple way on a horizontal plane. The force F acting on the plane is resolved in two mutually perpendicular stresses: normal stress (σ) and shear stress (τ)

Let us consider the above stress system in a rather simple way. The normal stress (σ), shear stress (τ) and applied force (F) are shown in Fig. 3.7d. From any point Q along the line of applied force, a perpendicular can be drawn on the line of stress to meet at R. In the triangle QRP, length QR then represents the shear stress (τ) while the lengths QP and RP represent the applied force (F) and the normal stress (σ), respectively. From the triangle QRP, we thus get the following:

$$\sin\theta = \frac{\tau}{F}$$
$$\cos\theta = \frac{\sigma}{F} \quad (3.4)$$
$$\tan\theta = \frac{\tau}{\sigma}$$

Geometrical representation of 2D stress on a plane thus helps in getting values of some unknown parameter if other two parameters are known.

In the coordinate system, the 2D stress can be represented by abscissa (x) and ordinate (y) as shown in Fig. 3.8. In the case of a normal stress (σ), each face of the plane can be considered to be represented by a set of two equal and opposite stresses, σ_{xx} and σ_{yy} (Fig. 3.8a), while in the case of a shear stress, the plane can be considered to be under the influence of two equal stresses, σ_{xy} and σ_{yx} (Fig. 3.8b), acting on the sides of the plane. The 2D stress σ_{ij} acting on the plane can then be represented by four components as shown by the following matrix:

$$\sigma_{ij} = \begin{bmatrix} \sigma_{xx} & \sigma_{xy} \\ \sigma_{yx} & \sigma_{yy} \end{bmatrix} \quad (3.5)$$

The state of two-dimensional stress acting on a plane such that there are only four components, as expressed above, and of which there are only three independent components, is called *plane stress.*

If, instead of designating the two orthogonal axes as x and y, we represent them by x_1 and x_2, the state of 2D stress at a point can be expressed by the following simple matrix:

$$\sigma_{ij} = \begin{bmatrix} \sigma_{11} & \sigma_{12} \\ \sigma_{21} & \sigma_{22} \end{bmatrix} \quad (3.6)$$

This matrix has only four components. Since $\sigma_{12} = \sigma_{21}$, the matrix has only three independent components.

3.12 Mohr Two-Dimensional Stress Diagram

Christian Otto Mohr, a German engineer, developed a graphical method in 1882 that represents in two dimensions the relationship between shear stress and normal stress. From the graph, it is possible to compute the values of shear stress, normal stress and angle of failure at the point when a crack is developed, i.e. when failure occurs. Mohr's theory can thus be applied to predict the failure of brittle materials.

In a *Mohr diagram*, the normal stress σ_n and shear stress τ_s are represented by the abscissa and ordinate of a graph. The known values of the greatest principal stress σ_1 and the least principal stress σ_3 are plotted on the axis of normal stress. According to Mohr's theory, it is possible to represent the state of failure by paired values of normal stress and shear stress on any plane with any orientation within the body by constructing a circle passing through σ_1 and σ_3. This is called *Mohr circle* (Fig. 3.9). It is also known as Mohr stress diagram. A circle is drawn passing through the values of the greatest principal stress (σ_1) and the least principal stress (σ_3) along the x-axis. The difference between the greatest and the least principal stresses, i.e. ($\sigma_1 - \sigma_3$), is called *differential*

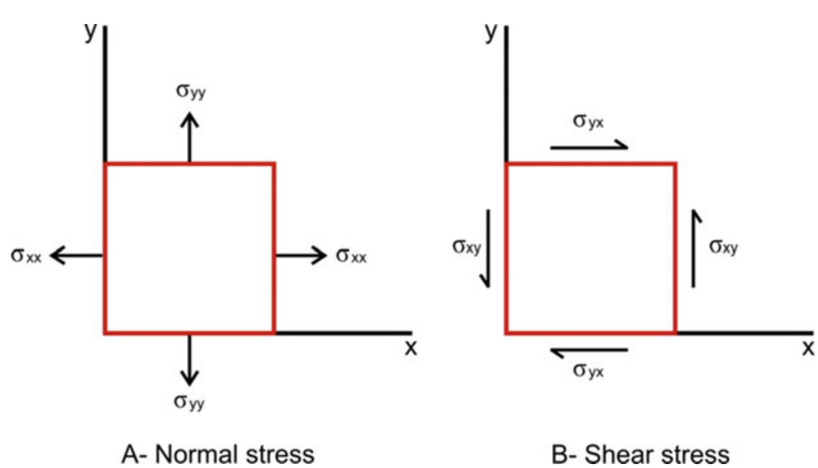

Fig. 3.8 Representation of two-dimensional stress on a plane in the coordinate system. (**a**) Normal stress, (**b**) shear stress (see text)

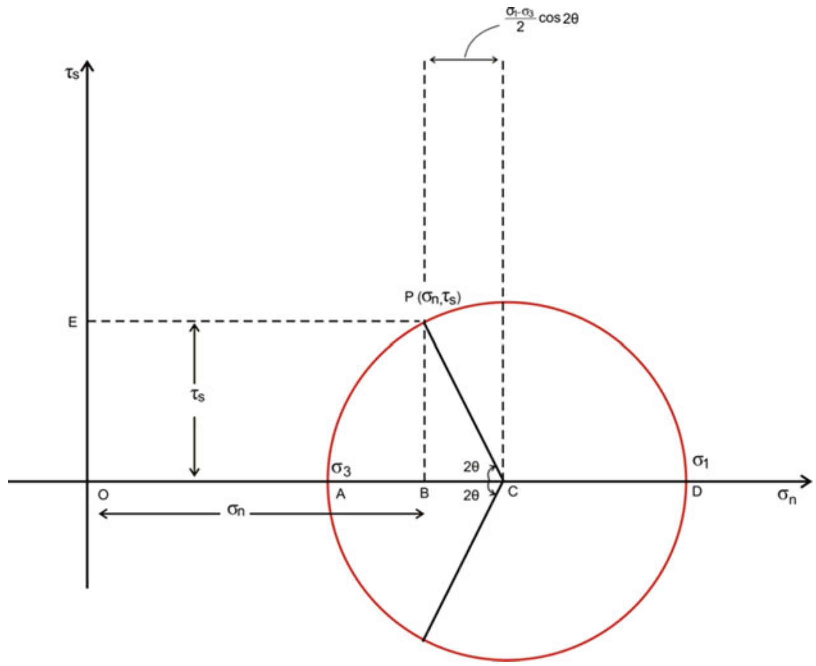

Fig. 3.9 The Mohr circle or the Mohr stress diagram. See text for details

stress and is represented by the diameter of the circle. If the angle between the shear fracture and the axis of normal stress is plotted on the circle, a point P where a fracture occurs is obtained on the circle. This is also known as the *fracture criterion*.

According to Mohr's theory, the coordinates of this point (σ_n, τ_s) should give the values of σ_3, normal stress and shear stress at the time of failure. 2θ is double the angle between σ_3 and the plane. From P, perpendiculars are drawn on the axes of τ_s and σ_n. Graphically, the values of normal stress and shear stress can be represented by lengths of PE and PB, which can be computed as below.

The diameter of the circle is ($\sigma_1 - \sigma_3$) so that its radius PC is 1/2 ($\sigma_1 - \sigma_3$). The centre C is {1/2 ($\sigma_1 - \sigma_3$), 0}. In the triangle PBC:

$$\mathrm{PB} = \tau_s = \mathrm{PC}\ \ \sin 2\theta = 1/2(\sigma_1 - \sigma_3)\sin 2\theta$$

Further, the *x*-coordinate of P is

$$\begin{aligned}\mathrm{OB} &= \mathrm{OC} - \mathrm{BC} = 1/2(\sigma_1 - \sigma_3) - \mathrm{PC}\ \ \cos 2\theta \\ &= 1/2(\sigma_1 + \sigma_3) - 1/2(\sigma_1 - \sigma_3)\cos 2\theta\end{aligned}$$

The *x*-coordinate of *C* should give the value of confining pressure at failure, i.e.

$$C = 1/2(\sigma_1 + \sigma_3) \tag{3.7}$$

Thus, at the point of failure, the values of shear stress, normal stress and confining pressure are given by

$$\tau_s = 1/2(\sigma_1 - \sigma_3)\sin 2\theta \tag{3.8}$$

$$\begin{gathered}\sigma_n = 1/2(\sigma_1 + \sigma_3) - 1/2(\sigma_1 - \sigma_3)\cos 2\theta \\ C = 1/2(\sigma_1 + \sigma_3)\end{gathered} \tag{3.9}$$

The Mohr circle, thus, gives a relationship between shear stress, normal stress and orientation of any plane. Accordingly, if the values of any two variables are known, the value of the third variable can easily be known.

From above, the radius *r* and the centre (*C*) of the circle σ_c are

$$r = 1/2(\sigma_1 - \sigma_3) \tag{3.10}$$

$$\sigma_c = 1/2(\sigma_1 + \sigma_3) \tag{3.11}$$

With these values, the equation for the Mohr circle is

$$\begin{gathered}(\sigma_n - \sigma_c)^2 + \tau_s^2 = r^2 \\ \{\sigma_n - 1/2(\sigma_1 + \sigma_3)\}^2 + \tau_s^2 = \{1/2(\sigma_1 - \sigma_3)\}^2\end{gathered} \tag{3.12}$$

On solving Eq. (3.12) for τ_s, we get

$$\tau_s = \pm\sqrt{\sigma_n(\sigma_1 + \sigma_3) - \sigma_n^2 - \sigma_1\sigma_3} \tag{3.13}$$

The positive and negative signs of Eq. (3.13) represent the top and bottom half of the Mohr circle, respectively; the top will represent right-lateral and the bottom left-lateral shear stress.

3.13 Three-Dimensional Stress

Stress that can be geometrically expressed and analysed in three dimensions is called *three-dimensional stress*. Since rocks exposed on the surface of the earth extend at depth, use of three-dimensional stress helps us interpret the forces acting at a particular point of the earth.

3.13.1 Stress Ellipsoid

A stress ellipsoid is a convenient way of graphically representing the state of three-dimensional stress acting at a point. If we assume that the normal stress acting upon the three planes of a body is represented by the lengths of the axes of a triaxial ellipsoid such that the shear stresses acting upon all the planes are zero, the ellipsoid thus constructed is called a *stress ellipsoid* (Fig. 3.10). The three mutually perpendicular planes in the ellipsoid are called the *principal greatest normal stress axis* (σ_1), *principal intermediate normal stress axis* (σ_2) and *principal least normal stress axis* (σ_3) such that $\sigma_1 \geq \sigma_2 \geq \sigma_3$. The state of stress acting at a point is then completely given by the directions and the sizes of the principal stresses.

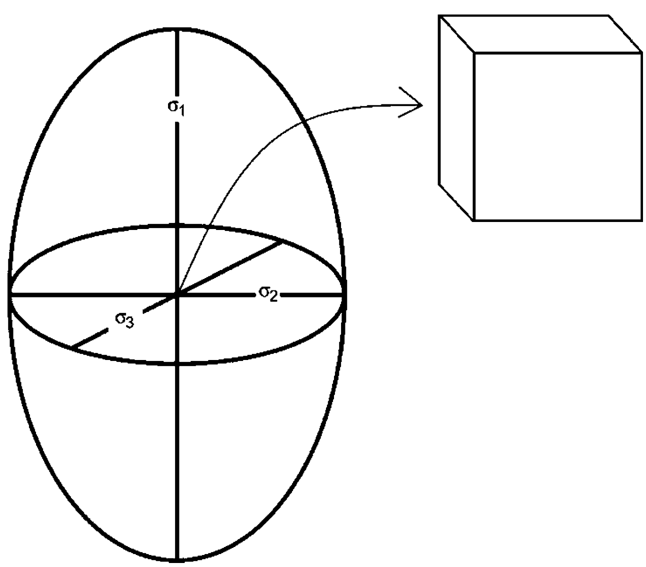

Fig. 3.10 Stress ellipsoid that represents the three-dimensional state of stress at a point that is represented by the infinitesimal cube located at the centre of the ellipsoid. σ_1, σ_2 and σ_3 are three mutually perpendicular surface stresses acting on the three principal planes. These are called the principal greatest normal stress axis, principal intermediate normal stress axis and principal least normal stress axis, respectively. Along these planes, normal stress has the greatest value while shear stress is zero. The stresses σ_1, σ_2 and σ_3 can be compressive (+ve sign) or tensile (−ve sign)

The plane that contains each of these axes is called a *principal plane*. Along the principal planes, normal stresses have greatest values while shear stress is zero. Each plane consists of two principal normal stresses. The difference between any two stresses gives the *differential stress* along that plane. Since σ_1 and σ_3 are the greatest and least stresses, ($\sigma_1 - \sigma_3$) gives the maximum differential stress. All planes other than the principal planes in the stress ellipsoid are *shear planes* because along such planes there is always a component of shear. Thus, we have only three planes along which the value of shear stress is zero.

3.13.2 Three-Dimensional Stress at a Point

We now consider three-dimensional stress acting at a point. For this, we imagine an infinitesimally small cube, i.e. one whose further smaller size is not visible. Let the cube be under stress by the action of a system of forces that can be represented by a single force F acting at the centre of the body. Let us now consider this cube on a three-dimensional coordinate system (Fig. 3.11) in which the three faces are parallel to each of the orthogonal axes x, y and z. Each of the six faces of the cube can thus be considered to be under the influence of the force. If we assume that the body is homogeneous and the effect of stress is also homogeneous, and also that the body is in equilibrium, the effect of each stress on a particular face is counterbalanced by an equal and opposite stress acting on the opposite face.

As we have explained earlier, a force acting at an angle to a plane would develop stresses that can be resolved into three components, one normal stress acting perpendicular to the plane and two shear stresses that are perpendicular to each other but act in the same plane. Thus, for the face, say ABCD, the force F will develop a normal stress σ_y and two shear stresses τ_{yx} and τ_{yz}. For another face CDEG, the stresses will be a normal stress σ_z and two shear stresses τ_{zy} and τ_{zx}. Likewise for the face ADEH, there will be a normal stress σ_x and shear stresses τ_{xz} and τ_{xy}. Since the body is in equilibrium, the stresses acting on each of these three faces will be counterbalanced by equal and opposite stresses acting on the opposite three faces of the cube. Thus, the state of stress σ_{ij} acting on the cube has been resolved into nine components and can be given by the following matrix or tensor form:

$$\sigma_{ij} = \begin{bmatrix} \sigma_x & \tau_{xy} & \tau_{xz} \\ \tau_{yx} & \sigma_y & \tau_{yz} \\ \tau_{zx} & \tau_{zy} & \sigma_z \end{bmatrix} \tag{3.14}$$

Each of the above stresses is called *stress components*. These nine components completely define the state of stress

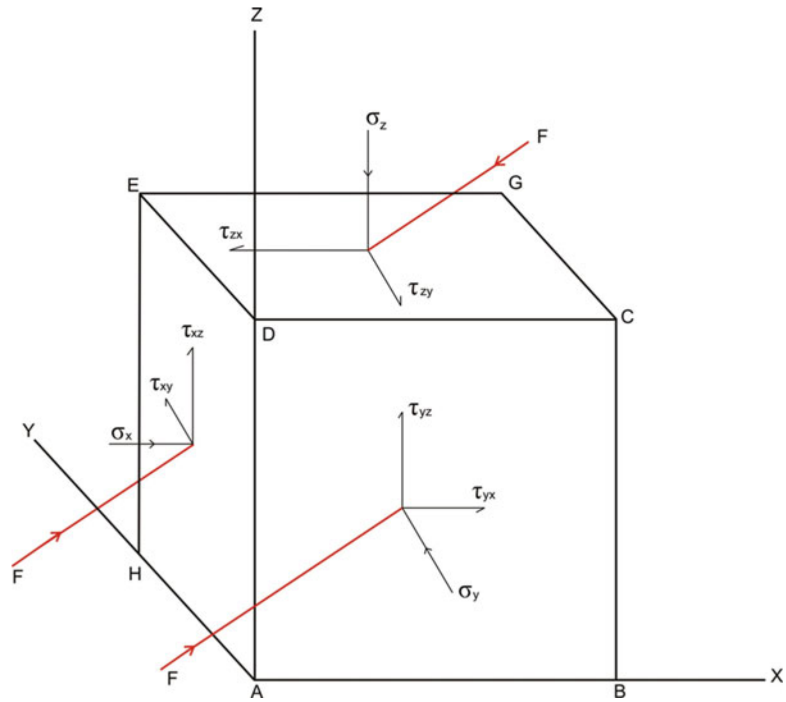

Fig. 3.11 Three-dimensional stress acting on an infinitesimally small cube. Note that on each face, a force *F* is resolved into three mutually perpendicular stresses of which one is a normal stress acting perpendicular to the plane while the other two are shear stresses acting parallel to the plane. Thus, with three stresses acting along each face, the three faces of the cube constitute a system of nine stresses (see text for further details)

in the cube in three dimensions. Since the cube is in equilibrium, each of the shear stresses acting on either side of an edge must balance each other and therefore there is no rotational movement, and therefore

$$\begin{aligned} \tau_{xy} &= \tau_{yx} \\ \tau_{yz} &= \tau_{zy} \\ \tau_{zx} &= \tau_{xz} \end{aligned} \tag{3.15}$$

Thus, there are only six independent components—three normal stresses σ_x, σ_y and σ_z and three shear stresses τ_{xy}, τ_{yz} and τ_{zx}—that completely define the three-dimensional stress acting at a point.

If we rotate the cube so that the coordinate system (*x*, *y* and *z* axes) gets aligned parallel, respectively, to the greatest, intermediate and least principal stresses, there will be no shear stress on the faces of the cube ($\tau = 0$ for all faces), and the stress tensor (i.e. Eq. 3.14) takes the form

$$\begin{bmatrix} \sigma_x & 0 & 0 \\ 0 & \sigma_y & 0 \\ 0 & 0 & \sigma_z \end{bmatrix} \tag{3.16}$$

Equation (3.16) represents a stress tensor with zero shear stresses and contains only three normal stresses, called *principal stresses*, acting perpendicular to the faces along which shear stresses are zero.

3.14 States of Stress

Depending upon the specific conditions, stress can be of various types; some common types are briefly described below.

3.14.1 Hydrostatic Stress

A state of *hydrostatic stress* develops when all the principal stresses are equal, i.e. $\sigma_1 = \sigma_2 = \sigma_3$. In this case, all the principal stresses are compressive (Fig. 3.12), and there is no shear stress as is the case with fluids. Since all the stresses are compressive, hydrostatic stress tends to reduce the volume.

3.14.2 Differential Stress

In a system with three unequal stresses, σ_1, σ_2 and σ_3, the difference between the greatest and the least stresses is called *differential stress*, σ_d, given by

$$\sigma_d = \sigma_1 - \sigma_3 \tag{3.17}$$

Crustal rocks are under the effect of differential stress, which plays a great role in rock deformation. A rock gets deformed when the amount of differential stress exceeds its strength.

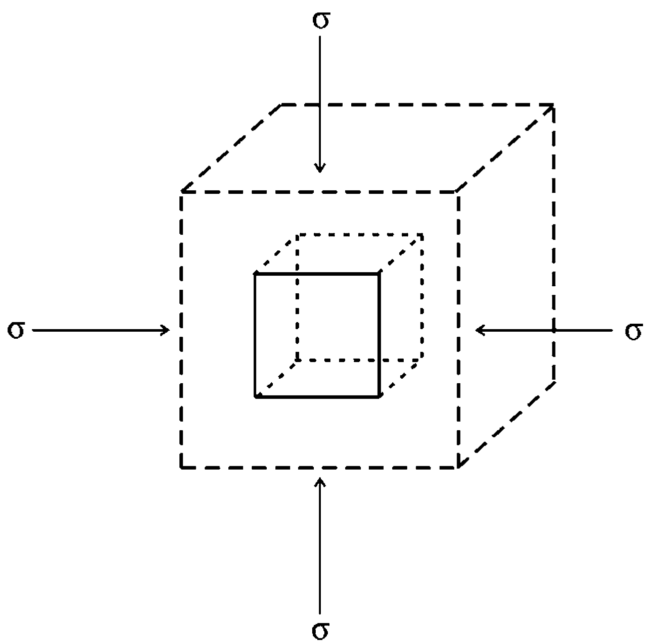

Fig. 3.12 Hydrostatic stress. All the principal stresses are equal and compressive that tend to reduce the volume of the body

3.14.3 Deviatoric Stress

Consider a system where all the principal stresses are unequal, i.e. $\sigma_1 \neq \sigma_2 \neq \sigma_3$, and all are compressive. In such a case, we can imagine a *mean stress*, σ_m, which is given by

$$\sigma_m = (\sigma_1 + \sigma_2 + \sigma_3)/3 \tag{3.18}$$

Now, the body can be considered to be under the influence of three unequal principal stresses (Fig. 3.13)—$(\sigma_1 - \sigma_m)$, $(\sigma_2 - \sigma_m)$ and $(\sigma_3 - \sigma_m)$—the average of which constitutes a mean stress (σ_m). The actual principal stresses thus mark departure under the influence of a system of these three stresses. Each of these is known as *deviatoric stress*. *A deviatoric stress is thus a component of a stress that expresses the difference between a normal stress and the mean stress*. It can be both a two-dimensional stress and a three-dimensional stress. Deviatoric stresses tend to change the shape of a body, which is then said to have undergone distortion.

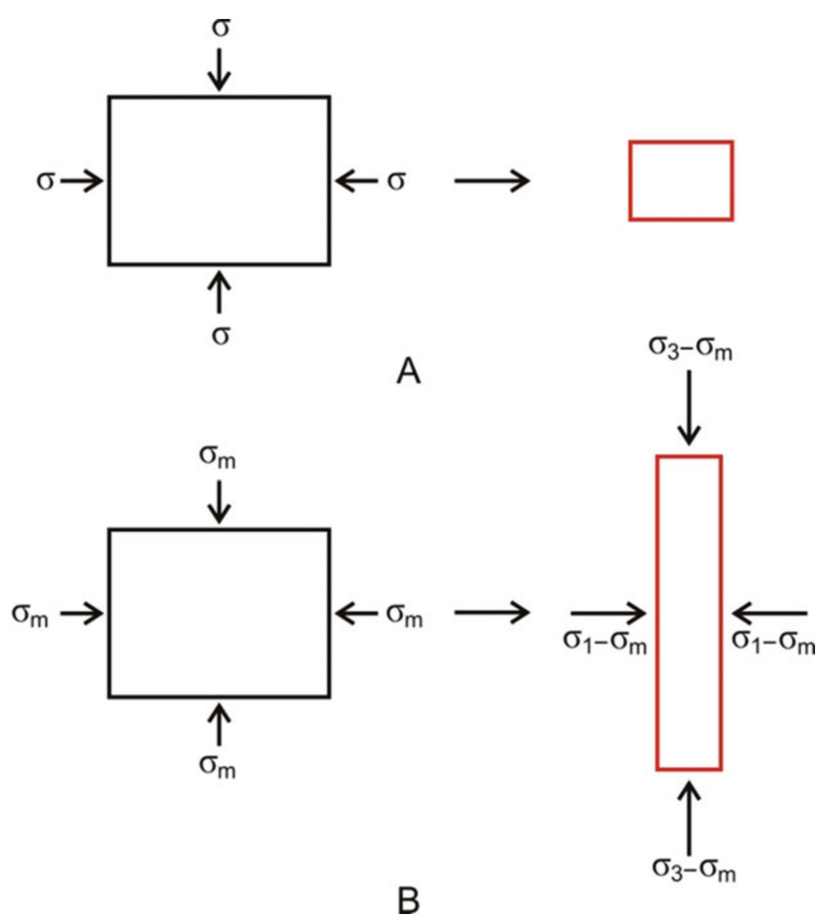

Fig. 3.13 Hydrostatic stress and deviatoric stress in two dimensions. (**a**) A hydrostatic stress (σ) causes a change (reduction) in volume but no change in shape of a body. (**b**) Deviatoric stresses $\sigma_1 - \sigma_m$ and $\sigma_3 - \sigma_m$ cause a change in shape of a body

Box 3.1 What Is a Tensor?

A *tensor* in general represents a set of numerical quantities that can describe the physical state of a material. A tensor relates the various geometrical vectors that, when related to a coordinate system, constitutes an organized frame for a set of physical properties of the body. Although tensors are individually independent of a coordinate system, they can be expressed in a coordinate system during a computational work. Tensors thus constitute geometric objects that relate the vector and scalar quantities of an object and can be represented as a multidimensional array of numbers. It may be noted that vectors and scalars are also tensors. Study of tensor is based on the use of a fixed reference frame, which is the coordinate system that uses indices for its notation. In the context of a tensor, a *dimension* represents the range of the indices.

In structural geology, tensors are important in providing a mathematical framework in relation to certain physical properties of rocks such as stress and strain in particular and rock mechanics in general.

Ranks of Tensors

Since a tensor represents a multidimensional array of numbers, it is often required to specify the number(s) by which this multidimensionality is represented. A *rank* (or *order* or *degree*) *of a tensor* represents the number of dimensions or indices required to define the multidimensionality of an array. A rank describes the tensor's dimensions. These numbers are called *components of the tensor*. The rank of a tensor is therefore equal to the total number of dimensions or indices that are needed to define each component. A *tensor of rank zero* is one that has only a single number of dimensions, e.g. a scalar quantity that is unrelated to any axes of reference. A scalar has no indices at all. A *tensor of rank one* is one that has only one dimension, e.g. a vector. A vector requires one axis of reference, and with each axis three numbers or components are associated. *A tensor of rank two* is one that needs two components to define the array of its multidimensionality, e.g. stress and strain. A stress requires two axes of reference, and with each axis nine numbers or components are associated. If the two components σ_{ij} of a stress are related to two vectors l_i

Box 3.1 (continued)

and k_j in a linear manner, then the components σ_{ij} form a matrix as given below:

$$\sigma_{ij} = \begin{bmatrix} \sigma_{11} & \sigma_{12} & \sigma_{13} \\ \sigma_{21} & \sigma_{22} & \sigma_{23} \\ \sigma_{31} & \sigma_{32} & \sigma_{33} \end{bmatrix} \quad (3.19)$$

In general, a rank of a tensor represents the number of indices required to determine the components. As we have indicated above, a rank of zero is represented by a scalar, a rank of one by a vector, a rank of two by matrix (stress), a rank of three by hyper-matrix and so on.

Box 3.2 Concept of Traction

Traction implies the distribution of forces acting on any specific surface of a solid substance, say a rock mass. The concept of traction appears to have not yet been widely used in structural geology, though there are several situations where the concept finds its utility. Pollard and Fletcher (2005) have used this concept to elucidate deformation in rocks. They highlighted the role of a *traction vector* that gives the distribution of forces, or lack of it, acting on any arbitrary surface within a rock mass. Traction vector can explain several geological situations efficiently such as what were the forces distributed on the surfaces of a fault that would cause it to slip or what were the forces distributed on the surfaces of a dike that would cause it to open. In the concept of traction, the earth's surface has been recognized as a traction-free surface.

Let us come back to a rock which represents a material continuum as its space is filled up with some material. A rock can be considered heterogeneous as well as homogeneous depending upon the scale of our observation. On a smaller scale, it may show several constituents, say rock fragments, large grains or grain aggregates, so that it looks heterogeneous. On a larger, say outcrop, scale, the same rock may look homogeneous apparently with no discontinuities, thus reflecting a material continuity. The concept of traction vector, which also presupposes a material continuum, can thus be extended to rocks also.

3.14.4 Lithostatic Stress

Rocks in the earth's crust are always under stress caused by the weight of the overlying burden that gives rise to *lithostatic stress*. Since these rocks are in equilibrium, i.e. physically stable, we can assume that at each point the stress is acting uniformly from all sides. In other words, the principal stresses are equal, all the principal stresses are compressive and there is no shear stress.

3.15 Palaeostress

3.15.1 Nature of Palaeostress

Palaeostress refers to the stress locked in the rocks in the geologic past. It is a branch of structural geology whose target is characterizing stress systems acting in the past from their record in deformation structures, singularly from fault-slip data (Simón 2019, p. 124). Palaeostress analysis is concerned with the directions along which stresses were active during rock deformation and, in certain cases, gives the magnitude of stress of the geologic past. With a beginning in France in the 1970s, this branch is nowadays practised all over the world. Excerpts from a review by Simón (2019) on 40 years of palaeostress analysis are highlighted below.

Palaeostress analysis is based on four basic models of relationships between brittle structures and stress/strain axes (Fig. 3.14): (a) conjugate faults, (b) overall discontinuous deformation, (c) Wallace-Bott's principle and (d) orthorhombic or 'biconjugate' fault pattern.

The *conjugate fault model* (Fig. 3.14a) is based on the conceptual relationship between conjugate fault systems and stress axes, as was established by Anderson (1951). The *overall discontinuous deformation model* (Fig. 3.14b) is based on inferring the strain axes from an assembly of faults, stylolites and tension gashes that jointly represent the bulk of deformation of a rock body (Arthaud 1969). *Wallace-Bott's principle* assumes that fault slip parallels the maximum resolved shear stress, which depends on the attitude of the fault plane and the stress ratio $R = (\sigma_z - \sigma_x)/(\sigma_y - \sigma_x)$, where σ_z is the vertical principal stress and $\sigma_y > \sigma_x$ (Fig. 3.14c). From slip data on a minimum of four independent faults, the four parameters that define the reduced stress tensor (three Euler angles for the orientation of stress axes, and R) can be obtained. *Orthorhombic or 'biconjugate' fault pattern* (Fig. 3.14d) is based on the analysis of faulting in three-dimensional strain fields (Reches 1978). Considering all the above models, palaeostress analysis is based on a number of assumptions (Simón 2019, p. 126):

(a) The stress state is homogeneous within the studied rock body.
(b) All fault slips are related to a single stress tensor.
(c) All deformation is accommodated by discrete slip on faults, i.e. blocks bounded by the fault planes are rigid and show no significant rotation.
(d) Slip accumulates on randomly oriented pre-existing fractures.
(e) The studied rock volume is large compared to the scale of each individual fault, and fault displacements are small with respect to fault dimensions.
(f) Movement on each fault is independent of the other ones, i.e. either they are not coeval or they do not interact to each other.

Simón's review emphasizes that palaeostress is a powerful tool for understanding ancient stress fields recorded in rocks, and as such it has implications for lithospheric dynamics and present-day plate kinematics.

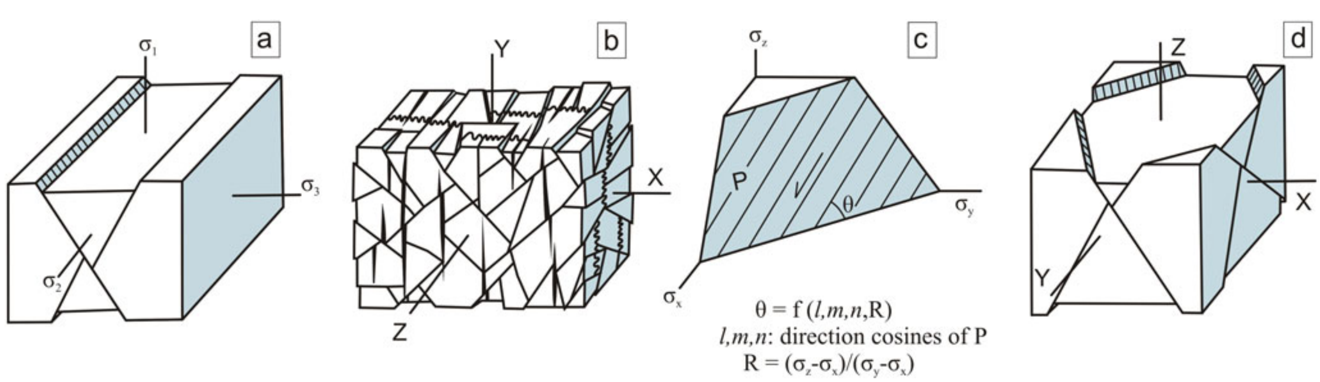

Fig. 3.14 Relationships between brittle structures and strain/stress axes have been explained by four models. From left to right, these are (**a**) conjugate faults, (**b**) overall discontinuous deformation, (**c**) Wallace-Bott's principle and (**d**) orthorhombic or 'biconjugate' fault pattern. See text for details. (Reproduced from Simón 2019, Fig. 1 with permission from Elsevier Copyrights Coordinator, Edlington, U.K. Submission ID: 1193025)

3.15.2 Estimation of Palaeostress

Estimation of palaeostress is sometimes also called *palaeopiezometry*. The features/signatures in rocks that help in estimating the palaeostress are called *palaeopiezometers*. Palaeostress has been estimated by various workers by devising several methods. The various methods use both macroscopic (folds, faults, fractures, joints, veins) and microscopic (grain size in dynamically recrystallized rocks, calcite twins, deformation lamellae in quartz, pressure solution, microfractures, etc.) features of rocks. Estimation of palaeostress has been done by several workers (see Srivastava et al. 1998; Simón 2019, and the references therein) using different methods. Although description of all the methods is not possible here, some common tools or features used in palaeostress estimation are highlighted.

3.15.2.1 Faults

Faults are common structures used in palaeostress analysis. Movement on fault planes occasionally gives rise to slickenside lineations that help in reconstructing the orientation of the fault slip. The stress-slip relationship for the fault could be established by applying the *Wallace-Bott hypothesis* that states that slip along a fault plane should be parallel to the direction of the greatest shear stress. It is thus possible to establish the orientation of all the principal stress axes. It is assumed that most large faults have signatures on small scales or outcrop scales. The above hypothesis can be applied for a population of smaller faults, and the results thus obtained can be generalized for the larger faults. Although the Wallace-Bott hypothesis provides an important tool for palaeostress analysis, it has some limitations. For example, the hypothesis requires that the fault has undergone only one phase of deformation and that it has undergone no rotation. Field evidences, however, reveal that many fault planes show more than one set of slickensides and evidence of rotation. Despite this, the hypothesis stands as a reliable tool for investigating the stress-slip relation of faults.

3.15.2.2 Grain Size

Deformation of rocks under certain conditions promotes recrystallization. When recrystallization takes place during concurrent deformation, it is called *dynamic recrystallization*. Recrystallization in the absence of deformation is called *static recrystallization*. During dynamic recrystallization, a strained grain releases its internal strain energy by the formation of smaller, strain-free grains. As deformation continues, smaller, strain-free grains form more in number and the rock ultimately becomes finer grained. The process of recrystallization is accentuated by increase of temperature.

Dynamic recrystallization involves appearance of new dislocation densities, and the degree of recrystallization depends much on the differences in dislocation density across the grain boundaries. Thus, dynamic recrystallization is a stress-sensitive process. In fact, it is the differential stress that dominates during concurrent deformation. This in turn suggests that the size of the subgrains formed due to dynamic recrystallization is a function of the differential stress. The size of the dynamically recrystallized subgrains is related to the flow stress by the following equation:

$$\sigma = Ad^{-m} \tag{3.20}$$

where σ is the differential stress in megapascals, d is the grain size in μm and A and m are constants for the material (Mercier et al. 1977; Twiss 1977; Ord and Christie 1984).

The above relation implies that during dynamic recrystallization, the average grain size undergoes reduction with increase in differential stress. Thus, the grain size of a dynamically recrystallized rock can serve as a means to estimate palaeostress in such rocks. The method as such demands further refinement. For example, the values of the flow stress for a given grain size have been found to be different by different workers. Also, we are not sure whether the grain size reduction is due to differential stress only; several other factors may also play their part. Despite all these shortcomings, the method is significant especially in the absence of any other technique available at hand.

3.15.2.3 Calcite Twins

Some minerals such as calcite develop twins due to deformation. If simple shearing takes place parallel to twin planes in a calcite crystal, the mechanism is called *twin gliding*. The calcite twins thus developed are the result of stress that produces bends in the crystal lattice structure. Twin gliding can take place at very low shear stress of 10 MPa (Twiss and Moores 2007, p. 500). By geometric construction of the twin plane and the shear direction, it is possible to infer the orientation of the principal stresses that gave rise to twin gliding.

Yamaji (2015) pointed out that although mechanical twinning along calcite *e*-planes is used for palaeostress analyses, the orientations of twinned and untwinned *e*-planes are known to constrain not only stress axes but also differential stress, *D*. The orientations lose the resolution of *D* if the twin lamellae were formed at *D* greater than 50–100 MPa. This can have distortive effects on palaeostress analysis.

3.15.2.4 Veins

Veins are commonly used for palaeostress estimation. Veins represent hydrothermal fluids, which under high fluid pressure result in dilation of pre-existing anisotropy/flaws in a rock, thus forming hydrothermal veins (Yardley 1986; GDH Simpson 1998). Veins are often associated with mineral deposits, some of which are of high commercial value such as gold.

Recently, Lahiri et al. (2020) used quartz veins for estimating palaeostress in mineralized and non-mineralized zones of Gadag, falling in a part of Dharwar craton of South India. The mineralized zones show occurrences of gold. The authors collected data on orientation and spacing of veins. By plotting the orientation data for every outcrop, the authors obtained two types of distributions: cluster distribution (when $P_f < \sigma_2$) and girdle distribution (when $P_f > \sigma_2$), where P_f is fluid pressure. Their palaeostress analysis reveals that the veins were emplaced when fluid pressure (P_f) varied from $P_f > \sigma_2$ (intermediate principal stress) to $P_f < \sigma_2$.

3.16 Stress Tensor

Use of tensor (see also Box 3.1) can be extended to our understanding of stress (see Means 1976; Oertel 1996, for details). Stress is a tensor of rank 2. Since stress is a force per unit area, we can assume that a unit cube of material is affected by forces acting on it in three dimensions (Fig. 3.15). We know the surface area of each face. If we know the value of the force acting on the body, we can calculate the amount of stress acting on the cube in three dimensions. The state of stress acting on the cube can be resolved into nine components that form a tensor, i.e. stress tensor.

If we use eigenvalues σ_{ij} for stress notation, then σ_1, σ_2 and σ_3 represent the principal axes of stress representing the maximum, intermediate and minimum principal stress axes. The 3D stress tensor (in the coordinate system x, y, z) can then be given by the following expression:

$$\sigma_{ij} = \begin{bmatrix} \sigma_{xx} & \sigma_{xy} & \sigma_{xz} \\ \sigma_{yx} & \sigma_{yy} & \sigma_{yz} \\ \sigma_{zx} & \sigma_{zy} & \sigma_{zz} \end{bmatrix} \tag{3.21}$$

If we choose coordinates x_1, x_2 and x_3, the state of stress acting on the cube can be resolved into nine components as shown in Fig. 3.16. The 3D stress matrix in this case takes the following form:

$$\sigma_{ij} = \begin{bmatrix} \sigma_{11} & \sigma_{12} & \sigma_{13} \\ \sigma_{21} & \sigma_{22} & \sigma_{23} \\ \sigma_{31} & \sigma_{32} & \sigma_{33} \end{bmatrix} \tag{3.22}$$

Since the body is in equilibrium and there is no shear stress on principal planes,

$$\sigma_{12} = \sigma_{21}, \sigma_{13} = \sigma_{31} \text{ and } \sigma_{23} = \sigma_{32},$$

and therefore the number of independent stress components is six.

If $\sigma_{13} = \sigma_{31} = \sigma_{23} = \sigma_{32} = \sigma_{33} = 0$, the stress tensor can then be expressed in 2D in the following form:

$$\sigma_{ij} = \begin{bmatrix} \sigma_{11} & \sigma_{12} \\ \sigma_{21} & \sigma_{22} \end{bmatrix} \tag{3.23}$$

A stress tensor is taken to be symmetric, i.e.

$$\sigma_{ij} = \sigma_{ji}.$$

Further, a symmetric matrix can be diagonalized, i.e.

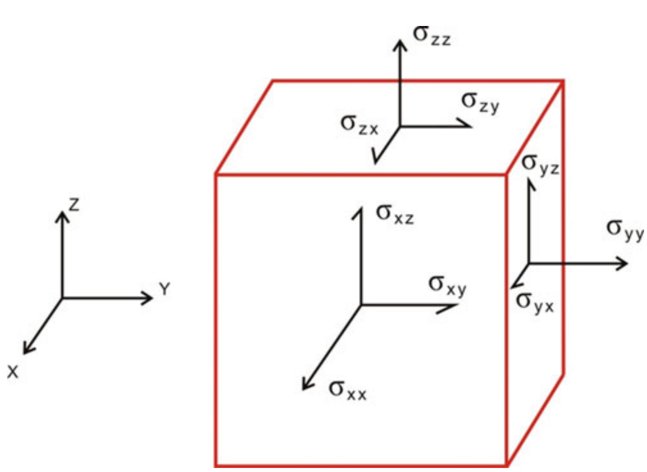

Fig. 3.15 Representation of stress tensor acting on a cube with eigenvectors on Cartesian coordinates X, Y, Z

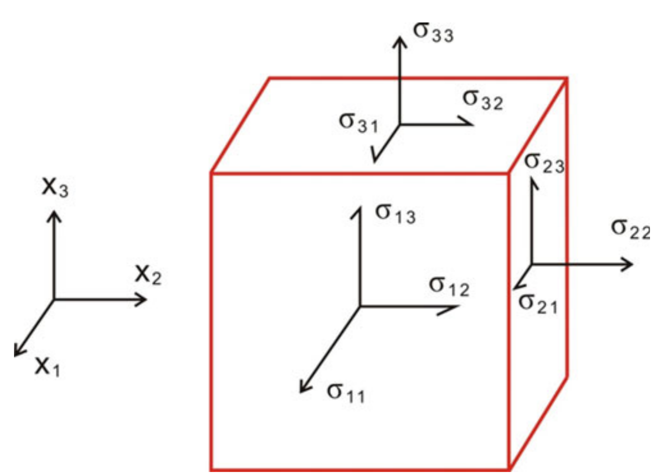

Fig. 3.16 The state of stress acting on a cube as represented in coordinate system x_1, x_2, x_3

$$\sigma_{ij} = \begin{bmatrix} \sigma_1 & 0 & 0 \\ 0 & \sigma_2 & 0 \\ 0 & 0 & \sigma_3 \end{bmatrix} \quad (3.24)$$

In this matrix, the off-diagonal terms of stress tensor are zero and the diagonal elements of the tensor, i.e. σ_1, σ_2 and σ_3, represent the three principal stresses (note that shear stresses are equal to zero).

3.17 Stress Field

As highlighted earlier, the state of stress in a rock mass inside the earth's surface is compressive and by nature it is hydrostatic, i.e. equal in all directions. This seems to be an oversimplification as the stresses are not always of equal amount in all directions. A generalized state of stress in rocks is called *stress field*. In fact, it is the stress field that determines what types of deformation the rock mass should show. Knowledge on stress field thus throws light on the deformation behaviour of the crust at any particular time.

Several workers have attempted estimation of stress field in some selected parts of the crust. Here, we present an example of the regional stress field of the Indo-Australian plate highlighting the results of the work of Cloetingh and Wortel (1986). The authors considered the Indian plate as elastic with a nominal thickness of 100 km and divided the spherical surface of the Indian plate in grids with the maximum grid size of 5 degrees. The results suggest a high level of regional stress field in the Indo-Australian plate. There is a concentration of compressive stresses of the order of 3–5 kbr in the Ninetyeast Ridge area. The maximum compression acts horizontally in NW-SE-oriented direction. The areas west of the Indian peninsula have an approximately N-S-directed tensional and E-W-oriented compressional stress field. There is more symmetric distribution of stress and deformation in the oceanic lithosphere with respect to the Indian peninsula. The Indian plate is undergoing a considerable net resistance at the Himalayan collision zone, and the exceptionally high level of the present-day regional stress field of the Indian plate is a transient feature that results from the unique dynamic situation in which the Indian plate now finds itself.

3.18 Stress History

In any sedimentary terrain, rocks have a long history from the time of sedimentation and burial to the time when they are seen exposed on the surface of the earth as a lithified, and may be deformed, rock. Accordingly, the stress condition also changes all through the long history of the rock from burial to the present-day form, and all this is called *stress history*. Of all, the joints are possibly the commonest structures that are formed at various stages from burial of sediments till uplift and erosion. However, the joints formed in such basins are formed under different stress conditions as the host rock continued to change its state (from unconsolidated mass to a lithified mass) as well as its location. For example, during sedimentation and diagenesis, the area witnessed increase of stresses, while during uplift and erosion, the stresses were gradually released. It is thus possible to understand stress history of any sedimentary basin or any sedimentary terrain by careful study of joints.

3.19 Stress Inside the Earth

3.19.1 Nature of Stress Inside the Earth

Earth below the surface is a storehouse of stress. Our knowledge of underground stress is important for certain engineering aspects. During excavation, for example, knowledge of pre-existing in situ stresses is necessary, especially for the stability of the excavation (Jaeger et al. 2007, p. 399). However, precise estimation of stress from below the earth's surface is rather difficult. Here, all the stresses are compressive; other types of stresses are considered negligible. Anderson (1951) suggested that below the earth the state of stress is everywhere equal to vertical. Also, estimation of horizontal stress is difficult, and therefore our knowledge on complete information on the state of stress on the earth's surface is as yet far from being satisfactory (Zoback and Zoback 1989).

In general, the in situ stresses below the earth vary as a function of depth, but the nature of the stresses and their variations are difficult to ascertain. The underground stresses are influenced by the topography, tectonic forces, constitutive behaviour of the rocks and local geological history (Jaeger et al. 2007, p. 399).

3.19.2 Basic Stress Types Inside the Earth

Our understanding of the basic stress types in the lithosphere comes from laboratory experiments. Engelder (1993, p. 22) suggested that the governors for earth stress include three general types of mechanisms that lead to failure of intact rock: *crack propagation*, *shear rupture* and *ductile flow* (Fig. 3.17). *Friction* is an additional governor for lithospheric stress if joints or shear fractures are reactivated to slip due to large shear traction. In the middle portion of the brittle intracontinental crust, the least compressive stress, σ_3, is

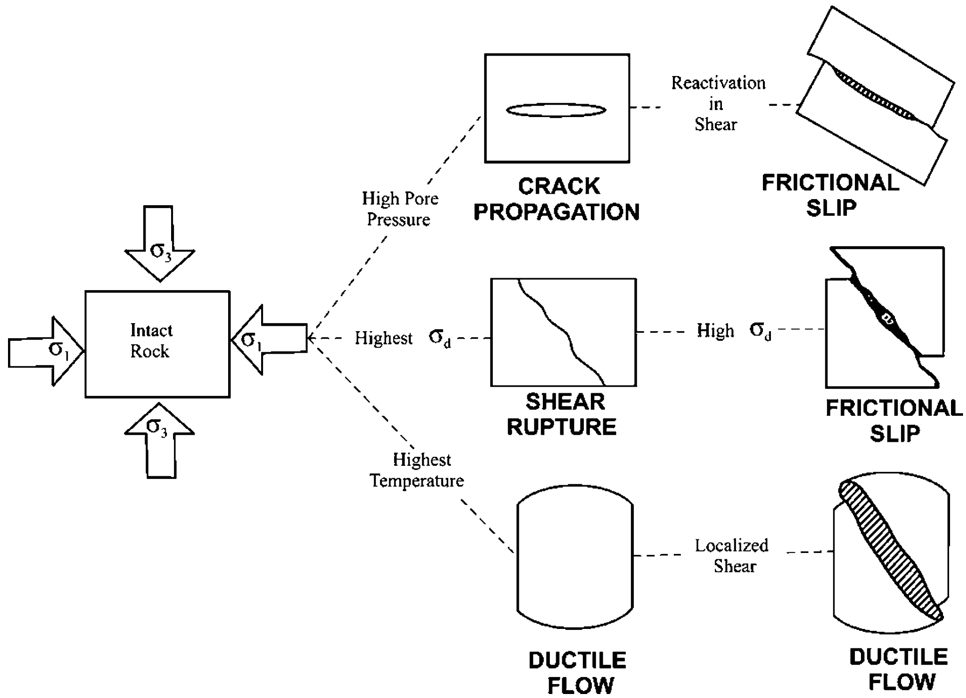

Fig. 3.17 Basic three types of stress inside the earth as revealed from the laboratory test. See text for details. (Slightly simplified from Engelder T, 1993, Stress Regimes in the Lithosphere Fig. 1–6. Reproduced with permission from Princeton University Press. Request ID: 600067225)

commonly horizontal but equally likely vertical in the upper kilometre or two (Nadan and Engelder 2009).

On the basis of rheological properties, Scholz (1990) divides lithosphere into two parts: the *schizosphere* or brittle region and the *plastosphere* or ductile region. Since earthquakes are a manifestation of brittle behaviour, the extent of schizosphere is demarcated by earthquakes. Since schizosphere shows evidences of some ductile deformation mechanisms such as stress solution, the boundary between schizosphere and plastosphere is gradational or blurred.

3.19.3 Causes of Stress Inside the Earth

Stresses inside the earth arise due to several causes of which the following are important.

3.19.3.1 Overburden

The weight of the overlying rocks and soils is broadly called overburden at any point inside the earth. By far, it is the most important factor to develop stress at any point. Obviously, the magnitude of stress inside the earth caused due to overburden increases with depth.

3.19.3.2 Pore Fluid Pressure

Presence of water or any fluid or gas has great effect on the physical-mechanical properties of rocks located inside the earth. Water constitutes the most common fluid inside the earth. Presence of water strongly affects the strength of rocks. Water is believed to be present in the pore space of the rocks. Water reduces the stress that is locked in the fractures, and this in turn proportionately reduces the pore fluid pressure, i.e. the hydrostatic pressure; as such, the normal stress of the rock is reduced (Ranalli 1987, p. 94). In such cases, failure is controlled by the *effective stress* (σ'), which is the principal stresses (considered positive for compressive stresses) minus the pore pressure (P) (Jaeger et al. 2007, p. 98), i.e.

$$\sigma_1' = \sigma_1 - P, \sigma_2' = \sigma_2 - P, \sigma_3' = \sigma_3 - P \qquad (3.25)$$

Effective stress significantly affects the deformational processes such as initiation of fractures or initiation of sliding of a pre-existing fault, and all these, on a grand scale, control various geodynamic processes such as faulting, thrust movement and seismic activity.

3.19.3.3 Thermal Stresses

Rocks are susceptible to changes in temperature in several ways. Rocks have variable thermal expansion, and because of this some rocks undergo change in volume to accommodate the change in temperature while others do not. Rocks that undergo change in volume can be considered to have accommodated the imposed thermal stress; such rocks practically do not induce additional stress. Rocks that do not undergo change in volume, on the other hand, can be considered to store thermal stress within them. In the rocks of the upper crust down to a depth of 15–20 km, thermal stresses arise from resistance to the temperature-induced volume change (Ranalli 1987, p. 146). In general, increase of temperature generates a compression, and it has been estimated that a change in temperature by 100 K generates stresses of the order of 100 MPa (Ranalli 1987, p. 148).

The amount of thermal stresses generated in the upper crust is generally very low as compared to stresses generated by some other processes. However, it is believed that thermal stresses can in some cases produce fractures, if not large-scale signatures of deformation.

3.19.3.4 Plate Motion

The state of stress in the lithosphere constitutes a primary cause of plate motion. High degree of stress is generated at plate boundaries where seismic and tectonic processes are active. The state of stress in a plate depends much on the type of relative plate motion, i.e. whether the boundary of two plates is convergent, divergent or transform. A major part of stress in a plate also arises from the plate-mantle interaction, which in turn varies depending upon whether the plate is oceanic or continental. The forces acting on the subducting slab are the largest and control the velocity of oceanic plates (Forsyth and Uyeda 1975). All this also implies that the state of lithospheric stress controls, directly or indirectly, the thermal convection system of the underlying mantle.

3.19.3.5 Burial

Below sea level, burial of sediments is a common process that generates a good deal of stress (burial stress) below the surface. Burial stress acts in vertical direction due to gravity. Maltman (1994) suggested that there is a proportional increase of burial stress with a depth at least up to about 1000 m below seawater. Here, sedimentation and tectonism also commonly go together, and it is the tectonic forces that induce dewatering to generate overpressures. Thus, in regions where burial is active, the crust continues to add stress.

3.20 Significance of Stress

3.20.1 Academic Significance

- Our knowledge of stress in rocks is necessary because it helps in giving an idea of directions in which the stresses had acted upon the rocks and also the amount of deformation, i.e. strain in rocks.
- Knowledge of stress field throws light on the deformation behaviour of the crust at a given time.
- Estimation of stress of the crust has implications for the direction of relative plate motions and seismicity distribution.

3.20.2 Engineering and Economic Significance

- The stresses inside the earth not only are heterogeneously distributed both horizontally and vertically but also keep changing with time. Despite all these, our understanding of the state of stress inside the earth is important, especially for several utility and practical aspects in addition to academic interests.
- In the construction of *dams*, knowledge of the state of stress is helpful in ensuring that the weight of the superstructure should be much below the affordable stress of the earth at that point.
- In *mining*, information on stress is necessary for the construction of tunnels, pits and related engineering work.
- During drilling for *petroleum*, artificial fractures are created at depth to increase the permeability of the petroleum-bearing horizon; this enhances the production of petroleum.
- In *civic construction* work, knowledge of in situ stress of the foundation ensures longer safety of the structures and buildings to be constructed.

3.21 Summary

- A rock mass under the influence of externally applied forces develops stress within it. Stress is thus caused by external forces and is given by the forces acting per unit area of a body.
- Every rock has strength (i.e. the value of stress that causes failure), small or large, due to which it is able to bear some or a large quantity of stress. If the stress developed in a rock exceeds its strength, the rock accommodates the stress by changing its shape or size, which is called *strain*. Thus, if stress is a cause, strain is an effect.
- Considering a rock mass as a deformable solid or fluid, the forces acting upon it can be of two types, *body forces* and *surface forces*. Body forces act upon a unit volume of the body and are therefore given by the volume or size of the body such as gravity or magnetic force. Surface forces act upon the boundaries of a body and thus act along the area where the forces are in contact with the body. Force acting on a unit area of the body is known as *stress*.
- Stress can be tensile, compressive or shear stress. *Tensile stress* develops when the stress acts along the plane in opposite directions. *Compressive stress* develops when the stress acts perpendicular to the plane. *Shear stress* develops when the stress acts in opposite directions but not along the same plane or line; as a result, the body shows angular changes of its original planes and thus to its shape.
- If all the stresses acting upon a body are represented by only three stress components acting parallel to each of the coordinate axes x, y and z, the normal stress acting parallel to each axis is called a *principal stress* and is designated as X, Y and Z, respectively, such that $X > Y > Z$. The coordinate axes along which the stresses act are called *principal axes of stress*.

- If we represent all the stresses acting on a body by lines, the longest and shortest lines trace an ellipse called *stress ellipse* that represents 2D stress. Likewise, the normal stress acting upon three planes can be represented by the lengths of the axes of a triaxial ellipsoid, and this gives a *stress ellipsoid*.
- When all the stresses act in the plane of a body, the state of stress is known as *biaxial stress*. It is a 2D stress that is represented in two directions: a normal stress acting normal to the plane and a shear stress acting along the plane.
- 2D stress can also be determined by *Mohr diagram* that can estimate the values of shear stress, normal stress and angle of failure at the point when a crack is developed, i.e. when failure occurs.
- Stress can be hydrostatic stress, differential stress, deviatoric stress and lithostatic stress.
- Stress locked in rocks in geologic past is called *palaeostress*. Basic principles of estimating palaeostress have been described.
- A generalized state of stress in rocks inside the earth's surface is called *stress field*. Knowledge of stress field throws light on the deformation behaviour of the crust at any particular time.
- The stress condition of a rock mass changes all through the long history of the rock from burial to the present-day form, and all this is called *stress history*.
- The state of stress inside the earth varies as a function of depth and is influenced by the topography, tectonic forces, constitutive behaviour of the rocks and local geological history.

Questions

1. What is stress? Explain why does a body develop stress.
2. What are the differences between body forces and surface forces?
3. What are normal stress and shear stress?
4. What is the difference between tensile stress and shear stress?
5. Outline salient features of uniaxial stress and biaxial stress.
6. What is Mohr diagram? Give its utility in structural studies.
7. With the help of a cube, describe the stress at a point and the state of stress acting on the cube.
8. What is deviatoric stress? What does it signify in the deformation of a rock?
9. Describe how the concept of stress tensor can be extended to structural geology.
10. Describe the causes of stress inside the earth.