Description, measurement and analysis of glacitectonically deformed sequences

Emrys Phillips and Jonathan R. Lee

1. Introduction

In this chapter of the guide, we show how glacitectonic structures can be described and classified in terms of their shape or morphology. The range of structures (folds, faults, foliations) developed within proglacially to subglacially deformed sediments provides important information on the character of these glacier-induced deformation events so their accurate description and measurement is key to understanding the stresses involved during their formation. The approach used is that routinely employed by structural geologists for recording the orientation and, where applicable, sense of movement recorded by faults, folds, foliations and lineations. In structural geology a distinction is made between **primary structures**, ones that develop during the formation/deposition of the sediment (e.g. bedding), and **secondary structures**, created later when the bedrock or sediment suffers strain in response to changes in the stress conditions, i.e. when it is deformed. This principle can, with care, be applied to glacial sediments. However, sedimentation and deformation of glacial sediments may occur almost simultaneously, for example during the formation of a subglacial traction till, so that a separation between the two processes can be difficult to establish.

The following sections are intended to provide a brief introduction to the description and classification of deformation structures. For a more detailed, comprehensive account the reader is referred to the relevant chapters in the structural geology text books included in the 'suggested further reading' section at the end of this chapter.

2. Primary or sedimentary structures

Primary structures, as defined above, develop during the formation or deposition of the sediment and include a wide range of sedimentary structures, for example bedding, cross-lamination, parallel-lamination...etc.

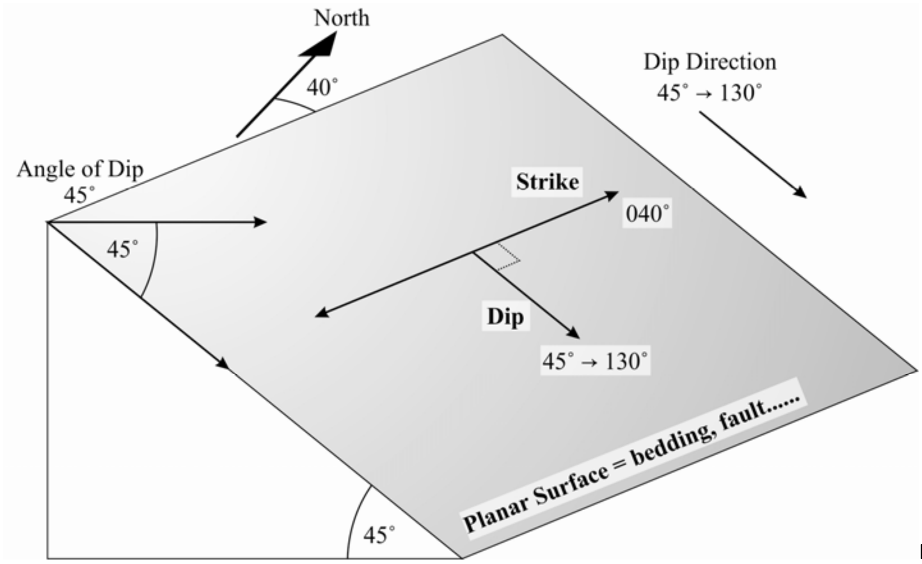


Figure 1.1. Instructions on measuring and recording dip and strike measurements using the right-hand-rule method

The orientation of planar sedimentary structures, such as bedding, can be determined by measuring its **dip** and **strike** (Figure 1.1). The **strike** is defined as the compass direction (0˚ to 360˚) of a horizontal line in the plane. The **dip** is the direction (0˚ to 360˚; or N, NW, SE...etc) and angle (0˚ to

90˚) of maximum inclination of the surface and is always measured perpendicular (90˚) to strike (Figure 1.1). The convention is to write this data in the form 220˚/36˚SE. However, an alternative method is to record the **dip** and **dip direction** of planar structures; for example, bedding 45˚ → 130˚, i.e. a dip of 45˚ in the direction of 130˚ from north (Figure 1.1). The systematic recording of the changes in the orientation of bedding between individual exposures or along the length of a cliff section, or quarry face, may reveal the presence of large-scale (10's or 100's m in amplitude) folds, or the presence of unexposed faults which result in the reorientation of bedding.

3. Secondary or deformation structures

3.1. Faults and fractures

Fractures are a common form of brittle deformation structure in geological materials and represent planar discontinuities where the rock or sediment has lost cohesion (failed or fractured). Where there is a measurable displacement across the fracture plane, i.e. one side has moved relative to

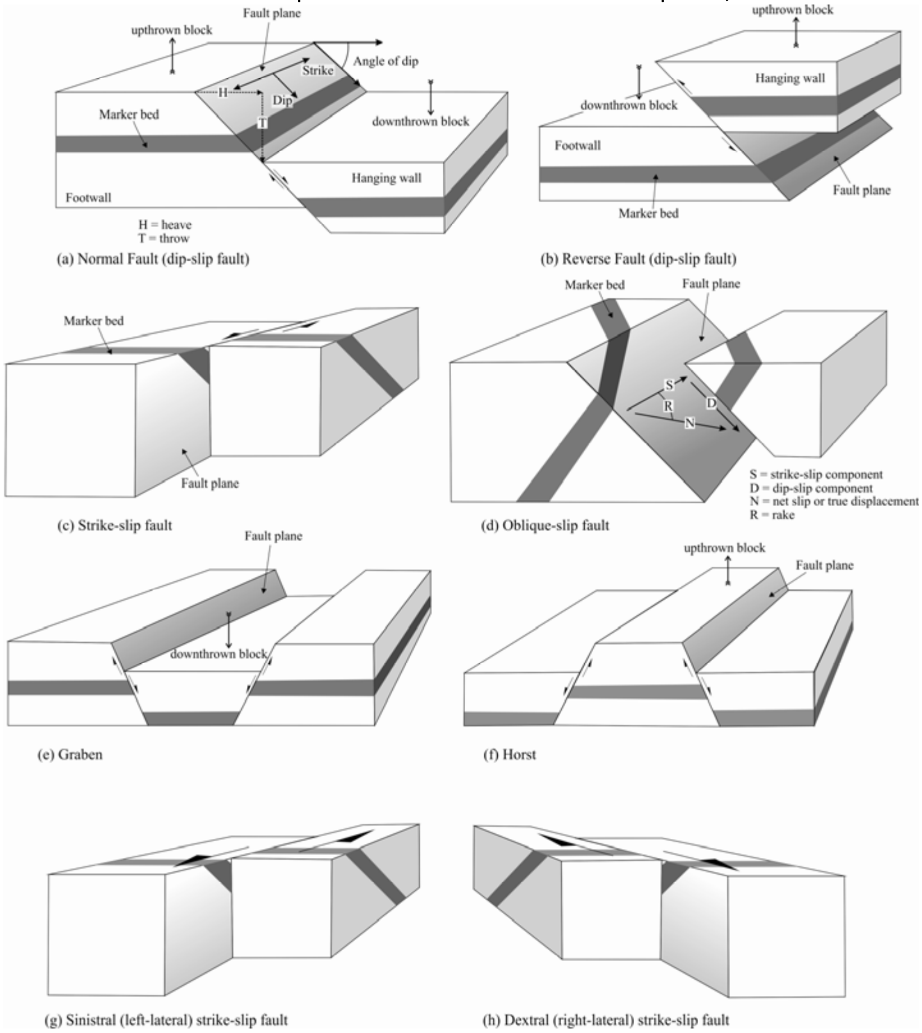


Figure 1.2. Nomenclature and classification of different types of faults: (a) Normal dip-slip fault; (b) reverse dip-slip fault; (c) strike-slip fault; (d) oblique dip-slip fault; (e) graben; (f) horst; (g) sinistral strike-slip fault; (h) dextral strike slip fault.

the other, the fracture is termed a **fault**. Conversely, where there is no visible displacement, the fracture is termed a **joint**. Fractures are important as their presence significantly affects the strength of a rock or sediment. Furthermore, open fractures formed under extensional stresses may represent important fluid pathways facilitating the migration of groundwater or meltwater through the glacial hydrogeological system.

3.1.1. Fault geometry and nomenclature

A **fault** is a planar fracture which has accommodated significant displaced/movement parallel to the fracture plane. The orientation of the **fault 'plane'** (fault planes are commonly non-planar) can be determined by measuring its dip and strike, or dip and dip direction. The main elements of faults are shown in Figure 1.2. Where the fault plane is not vertical the rock or sediment mass resting on the fault plane is referred to as the **hanging wall**, and the mass beneath the fault plane is called the **footwall** (the terms originated in mining).

The **slip** along a fault describes the movement parallel to the fault plane and can be used to divide faults into three main types: *(i)* **dip-slip** where movement is up or down parallel to the dip direction of the fault; *(ii)* **strike-slip** where movement is parallel to the strike of the fault plane; and *(iii)* **oblique-slip**, where the movement is a combination of both (i) and (ii). The **net slip**, or **true displacement**, is the total amount of motion measured parallel to the direction of movement. The term **separation**, the amount of apparent offset of a faulted surface (such as a sand bed) measured in a specified direction can be described as **strike separation**, **dip separation**, and the total or **net separation** of a fault. The terms **heave** (measured perpendicular to the strike of the fault) and **throw** (measured in the vertical plane containing the dip) have also been used to describe the horizontal and vertical components of dip separation, respectively (Figure 1.2a).

One feature commonly found on fault surfaces are grooves and growth of fibrous minerals, both aligned parallel to the movement direction and arranged in a series of steps, facing the movement direction of the opposing fault block. Polished fault surfaces are called **slickensides** and the striations on them **slickenlines**.

3.1.2. Classification of faults

Faults are divided into three main types (Figure 1.2a): *(i)* **Normal faults** are dip-slip faults in which the hanging wall has moved down (**downthrown**) relative to the footwall. **Lag faults** or **lags** are a type of normal fault where the fault plane dips at an angle of 30˚ or less. A **graben** comprises a downthrown block that has dropped down between two sub-parallel normal faults that dip towards one another (Figures 1.2e). The opposite is referred to as a **horst** (Figure 1.2f), comprising an upthrown block between two sub-parallel normal faults that dip away from each other; *(ii)* **Reverse faults** are dip-slip faults where the hanging wall has moved up (**upthrown**) relative to the footwall (Figure 1.2b). **Thrust faults** or **thrusts** are a type of reverse fault where the fault plane dips at an angle of 30˚ or less; and *(iii)* **Strike-slip** or **wrench faults** where movement occurred parallel to the strike of the fault plane (Figure 1.2c,d). The sense of movement on strike-slip faults can be describes as **sinistral** (or left lateral; Figure 1.2g) if the opposite block has moved to the left, or **dextral** (or right lateral; Figure 1.2h) if the opposite block has moved to the right, as viewed by an observer standing on one side of the fault looking across the fault plane.

3.1.3. Evidence of faulting

Evidence for the existence of a fault at a particular locality may include: *(i)* The displacement or offset of a recognisable marker horizon; *(ii)* The truncation of earlier formed deformation or primary sedimentary structures such as bedding, or stratigraphical and/or sedimentary units. This type of evidence, however, must be used with care as major erosion surfaces and unconformities may also lead to truncation; *(iii)* The repetition or omission of stratigraphic units; *(iv)* Localised deformation along the fault plane leading to the reorientation of pebbles of similar clasts, the formation of a glacitectonic foliation, or fragmentation of the sediments forming a **fault breccia**; *(v)* **Drag** of bedding along the fault plane resulting in a marked thinning of bedding or even folding (**drag folds**); and *(vi)* The formation of a **fault scarp** where the topographic surface is offset by dip-slip movement along a fault.

3.1.4. Joints and shear fractures

As defined above, a **joint** is a planar fracture which does not show any significant displaced/movement and is the most common of all the deformation structures. The orientation of the joint can be determined by measuring its dip and strike, or dip and dip direction. Sub-parallel, regularly spaced joints are referred to as a **joint set**, with two or more joint sets developed in a particular area constituting a **joint system**. The surface of the joint may be either smooth, or irregular, with others being marked by a series of low concentric ridges or feathered pattern (**plumose structure**) that record the direction of propagation of the fracture. Fractures may also occur as paired or **conjugate** sets.

The type of fracture formed depends on the orientation and magnitude of the principal stresses at the instant failure occurs, as well as the mechanical properties of the sediment or rock. Three types of fractures have been identified in structural geology, each formed by a separate kind of motion (Figure 1.3):

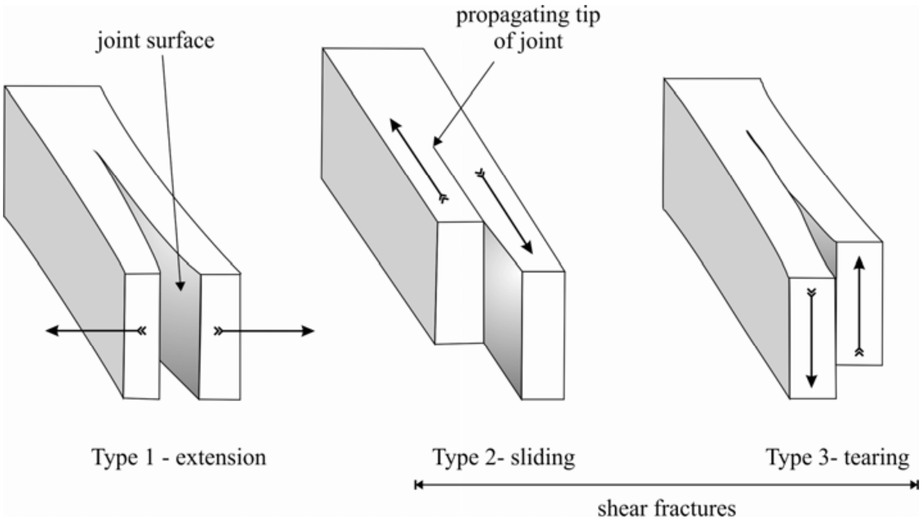


Figure 1.3. Types of joint and shear fractures

(i) Type 1 fractures are joints formed by extension leading to the opening of the fracture; *(ii)* Type 2 fractures form by sliding; and *(iii)* Type 3 fractures formed by a tearing motion. Types 2 and 3 are referred to as **shear fractures** and are broadly analogous to strike-slip and dip-slip faults.

Joints may also be characterised as tectonic, hydraulic, unloading or release joints based upon the environment and mechanism of their formation. **Tectonic** and **hydraulic joints** develop in response to fracturing caused by over-pressurised fluids (**hydrofracturing**). In the glacial environment these hydrofractures may be filled with fluidised and remobilised sediment. In contrast, **unloading** and **release joints** form as erosion removes overburden and contraction occurs. In the glacial environment unloading joints may develop during retreat as the load imposed on the bed is removed by the wasting or active removal of the overlying ice. Unloading joints are thought to begin to form when more than half the original overburden has been removed. For vertical unloading joints to form, the effective stress in the horizontal plane must be tensile (extensional). In glacial sediments, joints may also form in response to the dewatering or drying out of originally wet or saturated sediment, for example subglacial traction tills. This process may be analogous to the development of cooling joints in igneous rocks where thermal gradients and contraction processes control the orientation of the joints.

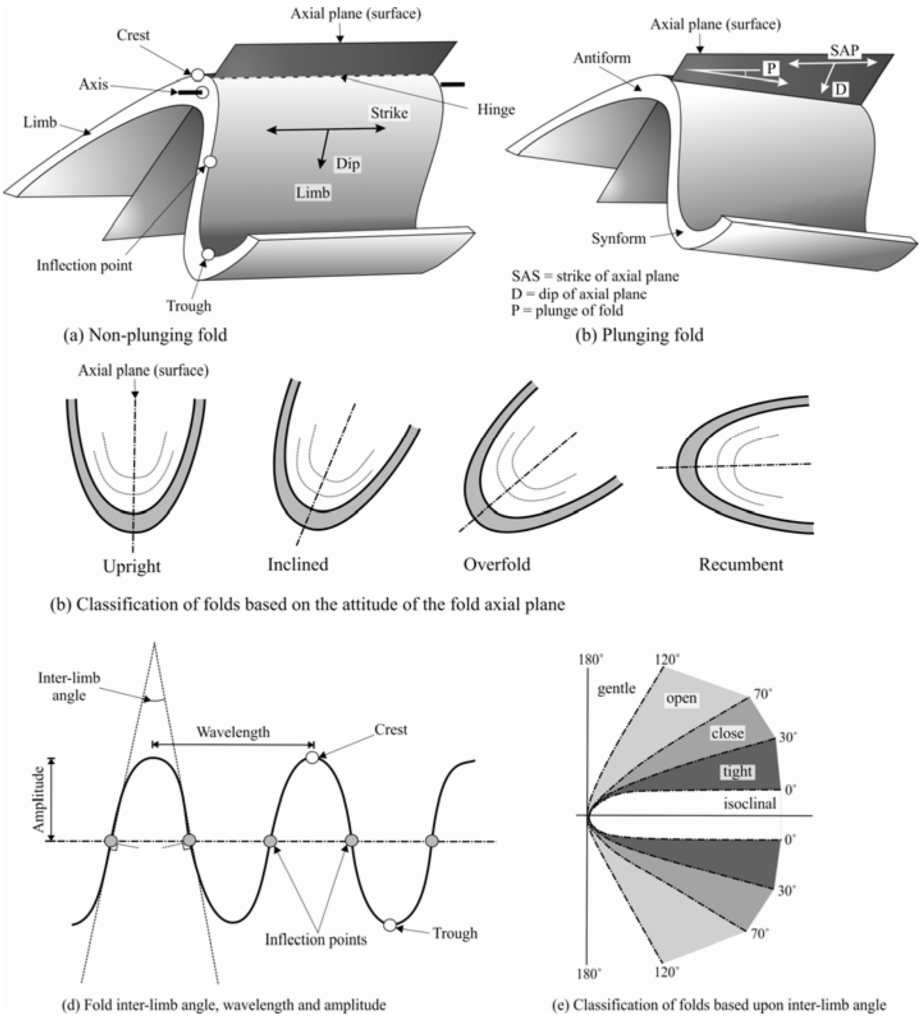


Figure 1.4. Classification and description of fold geometry. (a) non-plunging fold; (b) plunging fold; (c) classification of folds based on the attitude of the fold axial plane; (d) fold inter-limb angle, wavelength and amplitude; (e) classification of folds based upon inter-limb angles

3.2. Folds and folding

A **Fold** is a curved or wave-like structure that developed as a result of the ductile deformation of bedding, foliation, or other pre-existing planar structure present within a rock or sediment. They occur on a variety of scales, ranging from **micro-scale** structures only visible under the microscope (also referred to as **crenulations** or **microfolds**), to **meso-scale** folds which are clearly seen in hand specimen or outcrop, through to larger **macro-scale** features up to several tens of metres or even kilometres across.

3.2.1. Fold geometry and nomenclature

The main elements of the geometry of simple fold pair are shown in Figure 1.4a,b. The highest point (elevation) on a cross section through a fold is referred to as its **crest**, with the lowest point being the **trough**. The zone of maximum curvature on the fold is known as the **hinge** or **hinge zone**, with the **limbs** being the straighter or least-curved parts of the fold. The **inflection point** is a point on a fold limb separating the convex curved part of one fold, from the concave curved section of the adjacent fold. The **hinge line** is a line joining the points of maximum curvature on a folded surface. The hinge lines on a series of adjoining folded surfaces can be connected to define the folds **axial plane** (or **axial surface**). This imaginary plane cuts the hinge of the fold along a line known as the **fold axis**. It is important to note that where the axial plane is inclined the hinge and crest of a fold may not coincide. The angle made by the fold limbs is referred to as the **inter-limb angle** or **fold**

angle (Figure 1.4e). The tightness (small inter-limb angle) or openness (large inter-limb angle) of a fold is used by structural geologists as a way of classifying folds (see section 3.2.3) and may reflect the relative intensity of deformation. **Asymmetrical folds** possess one limb that dips more steeply and is shorter that the other, as a result these folds appear to overturned or 'lean' in a particular direction, known as the fold **vergence**. Conversely, in **symmetrical folds**, the limbs are approximately of equal length.

The **amplitude** and **wavelength** are a measurement of the size of a fold. The wavelength is determined by measuring the distance between adjacent crests (or troughs), whereas, the amplitude is half the perpendicular distance between the crest and the trough on a fold pair (Figure 1.4d).

3.2.2. Fold orientation

The orientation of a fold can be simply obtained by measuring the dip and strike (or dip and dip direction) of both limbs. Similarly the orientation of the axial surface of a fold may be measured in a similar manner. The plunge of a fold (or **fold plunge**) is the angle between the fold axis and the horizontal, and is recorded as the angle of plunge and direction from north; for example, fold plunge 25˚ → 110˚, i.e. an angle of plunge of 25˚ in the direction of 110˚. The direction from north of the vergence of a fold may also be recorded; for example, fold vergence towards 020˚. The systematic recording of the orientation of the axial surfaces and plunge of small, or meso-scale folds can provide important information on the orientation of larger macro-scale structures, as well as the orientation of the principal stresses responsible for the folding.

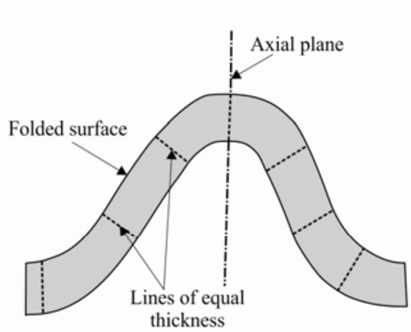
3.2.3. Classification of folds

The classification of folds used by structural geologists is based on four criteria: *(i)* direction of closing of the fold; *(ii)* the attitude of the axial plane; *(iii)* the size of the inter-limb angle; and *(iv)* the nature of the profile of the fold. These classification schemes are illustrated on Figure 1.4.

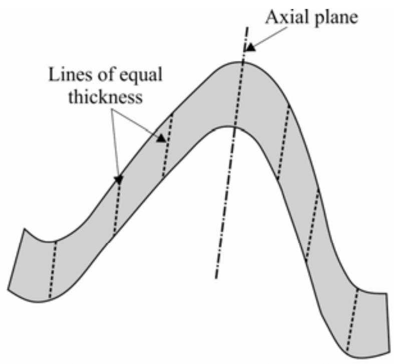
Antiforms are folds which close upwards and their limbs dip away from the hinge (Figure 1.4b). In the opposite situation, where the fold closes downwards and the limbs dipping towards the hinge, it is known as a **synform** (Figure 1.4b). Folds which close sideways are referred to as **neutral folds**. In the simplest situation, the sequence of bedded sediments being folded will become progressively younger upwards. In which case the oldest sediments occur within the core of the antiform and it is referred to as an **anticline**. Conversely, the younger sediments will occur within the core of the adjacent synform, or **syncline**. In areas of more complex folding it is possible that part of the sedimentary sequence may become locally overturned or inverted (i.e. becomes younger downwards). If this inverted part of the sequence is folded again during a later phase of deformation it is possible to get downward-closing (**downward-facing**) anticlines and upward-closing (**upward-facing**) synclines.

Folds can be divided into three groups based upon the attitude or dip of the axial plane (Figure 1.4d): *(i)* **upright folds** which possess a steeply dipping of vertical axial plane; *(ii)* **inclined folds** with moderate to gently dipping axial planes; and *(iii)* **recumbent folds** where the axial plane is sub-horizontal. The division between these classes is not rigidly defined. In some asymmetrical, inclined folds the sediments on one limb may be inverted (i.e. young downwards). These folds are sometimes referred to as **overfolds**.

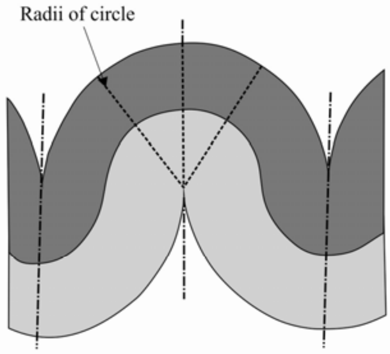
The size of the inter-limb angle is a measure of the tightness of a fold which reflects the amount of compression or shortening of the sedimentary layer. Figure 1.4e shows the classification scheme based upon inter-limb angle, diving folds into: *(i)* **gentle** (180˚ – 120˚); *(ii)* **open** (120˚ – 70˚); *(iii)* **close** (70˚ – 30˚); *(iv)* **tight** (30˚ – 0˚); and *(v)* **isoclinal** (0˚).



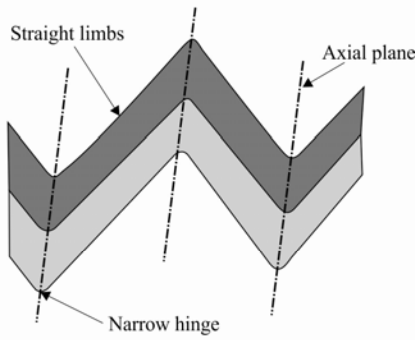
(a) Parallel folds - constant layer thickness measured perpendicular to fold surface



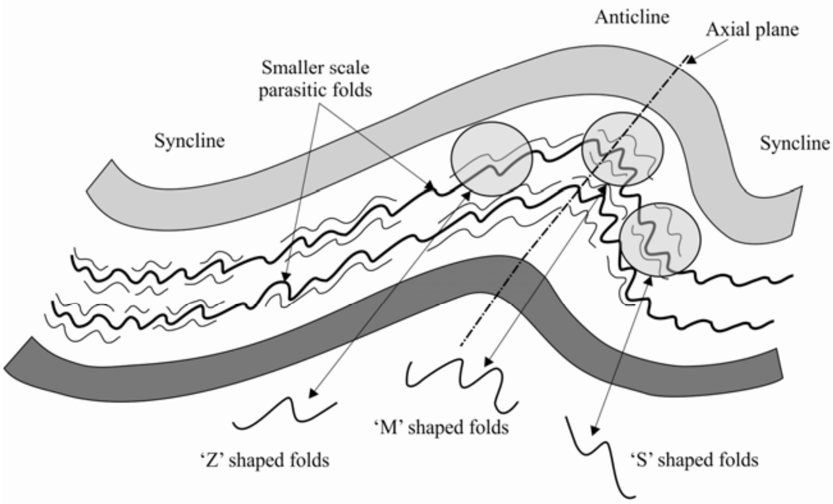
(b) Similar folds - constant layer thickness measured parallel to fold axial plane



(c) Concentric folds



(d) Chevron folds - narrow hinges separating straight fold limbs (kink folds)



(e) Disharmonic and parasitic folds

Figure 1.5 (previous page). The classification of folds based upon their profile, including parallel (a) and similar (b) folds; concentric (e) and chevron (d) folds; and disharmonic and parasitic folds (e)

The classification scheme based upon the profile of the fold is shown in Figure 1.5. **Parallel folds** are the simplest and are characterised by the fact that the orthogonal thickness of the folded layer remains constant (Figure 1.5a). A **concentric fold** is a special type of parallel fold where the folded surfaces are arcs of a circle with a common centre, known as the centre of curvature of the fold (Figure 1.5c). In **similar folds** the orthogonal thickness of the folded layer changes in a systematic manner so that the thickness measured parallel to the axial plane remains constant (Figure 1.5b). Folds which possess sharp hinges and planar limbs are known as **chevron folds** (Figure 1.5d). Where typically small-scale chevron folds are asymmetrical, the superimposed short limbs can give the appearance of narrow bands running across the sediment or rock. These bands are contained within the adjacent axial planes and are known as **kink bands** or **kink folds**.

3.2.4. Description of fold systems

Folds can occur as single isolated structures or as extensive **fold trains** or **fold systems** comprising a number of structures of different sizes. **Pumpelly's rule** states that small-scale structures, in general, mimic larger-scale structures that formed at the same time. Consequently, micro-scale and meso-scale folds may have the same shape and orientation as the larger, macro-scale structures. Smaller wavelength folds, also known as **parasitic folds** (Figure 1.5e), are superimposed on the limbs of larger wavelength structures. In many cases there is systematic relationship between the asymmetry of the parasitic folds and their position on the larger fold which formed at the same time. The sense of asymmetry of the parasitic folds changes from Z-shaped on the left-hand limb of an anticline, to M-shaped in the hinge zone, through to S-shaped on the right-hand limb of the anticline when viewed in cross section. This relationship can be used to establish the existence of very large scale folds within a deformed sequence and locate the hinge zones of these macro-scale antiforms and synforms.

Harmonic folds occur where the folds deforming adjacent layers correspond with each other in terms of their wavelength, symmetry and overall shape. Conversely, folds where the wavelength and shape changes from one layer to the next are referred to as **disharmonic** (Figure 1.5e).

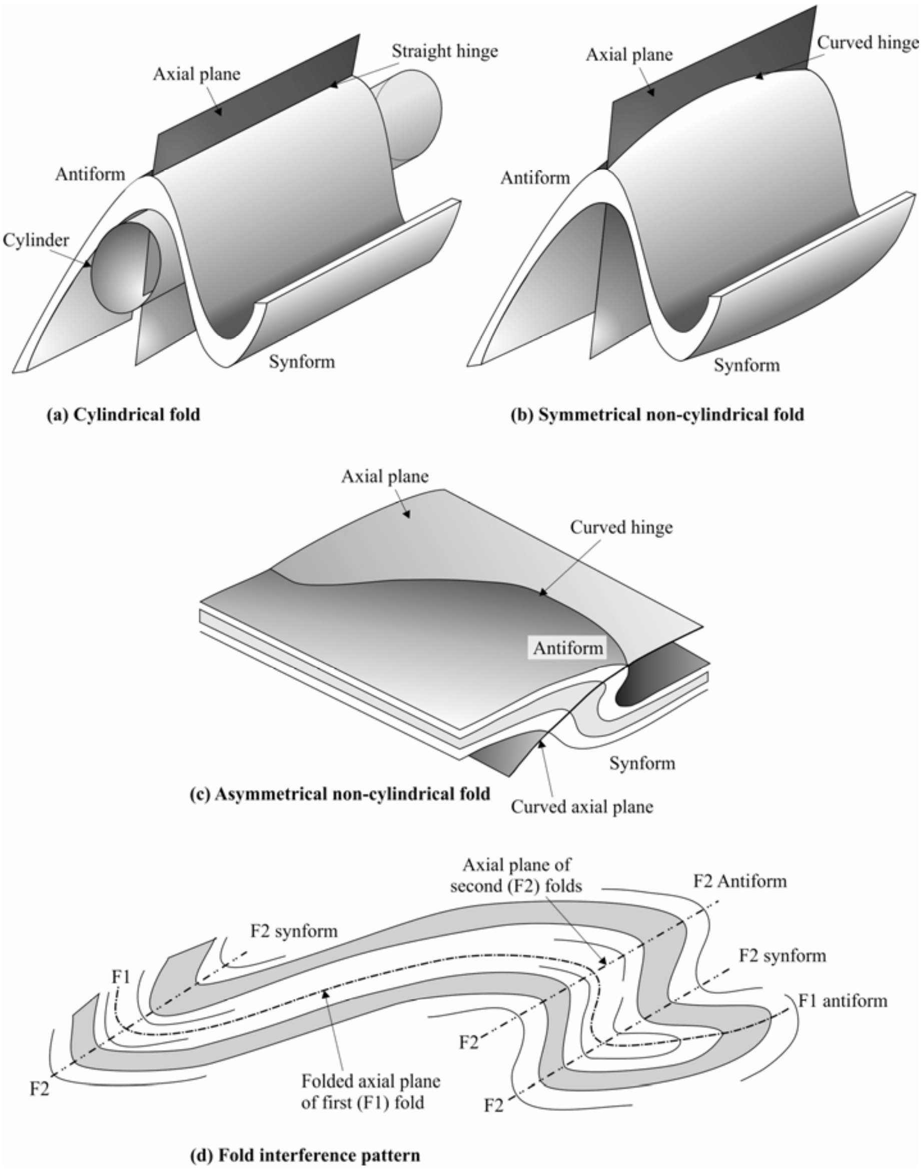


Figure 1.6. Folds in three dimensions including cylindrical (a), symmetrical non-cylindrical (b), asymmetrical non-cylindrical folds, and fold interference patterns (d)

Disharmonic folds often develop due to marked differences in the physical properties (e.g. lithology, water content and/or pressure) or thickness between the layers.

3.2.5. Folds in three dimensions

The previous sections deal with the description and classification of folds in 2D, however, the following section contains the terminology used by structural geologists to describe these structures in 3D. **Cylindrical folds** (or more correctly **cylindroidal**, or cylinder-like) are defined as *"folds that can be generated by moving a line (e.g. the fold axis) parallel to itself"* and are generally folds were the hinges are everywhere parallel on successive folds (Figure 1.6a). More simply, cylindrical folds look as if they have been formed by bending the layers around a cylinder. On the other hand, the hinges of **non-cylindrical folds** are curved and not parallel on successive folds, or the hinges of successive folds converge towards a point rather than being parallel (Figure 1.6b,c). **Sheath folds** are strongly non-cylindrical and close at one end, furthermore the fold hinges curve within the axial surfaces. These folds are commonly associated with shear zones where the rocks or sediments have been deformed by a strong component of inhomogeneous simple shear. In bedrock geology, the presence of sheath folds has often been used as an indicator of high cumulative strains. However, the unconsolidated nature of glacial sediments, coupled with the potentially high water contents/pressures encountered in subglacial to ice marginal environments means that sheath folds in glacially deformed sequences may develop at considerably lower shear strains.

The term **pericline** is typically only applied to large-scale folds where the amplitude of the fold decreases systematically to zero in both directions resulting in a humped or whale-back like profile to the fold. A **dome** is a unique kind of antiform where the layering dips in all directions away from a central point. A **basin** is a type of synform where the layering dips inwards towards a central point.

In highly deformed glacial sequences the fold systems may be complex and formed by the interference between two or more fold sets of much simpler geometry. An example of a **fold interference pattern** is shown in Figure 1.6d and formed as successive generations of folds are superimposed on the bedrock or sediment. Successive generations of structures should be distinguished chronologically using the nomenclature commonly used by structural geologists with F1 being the earliest folds and Fn the latest, with the later folds deforming the earlier formed structures.

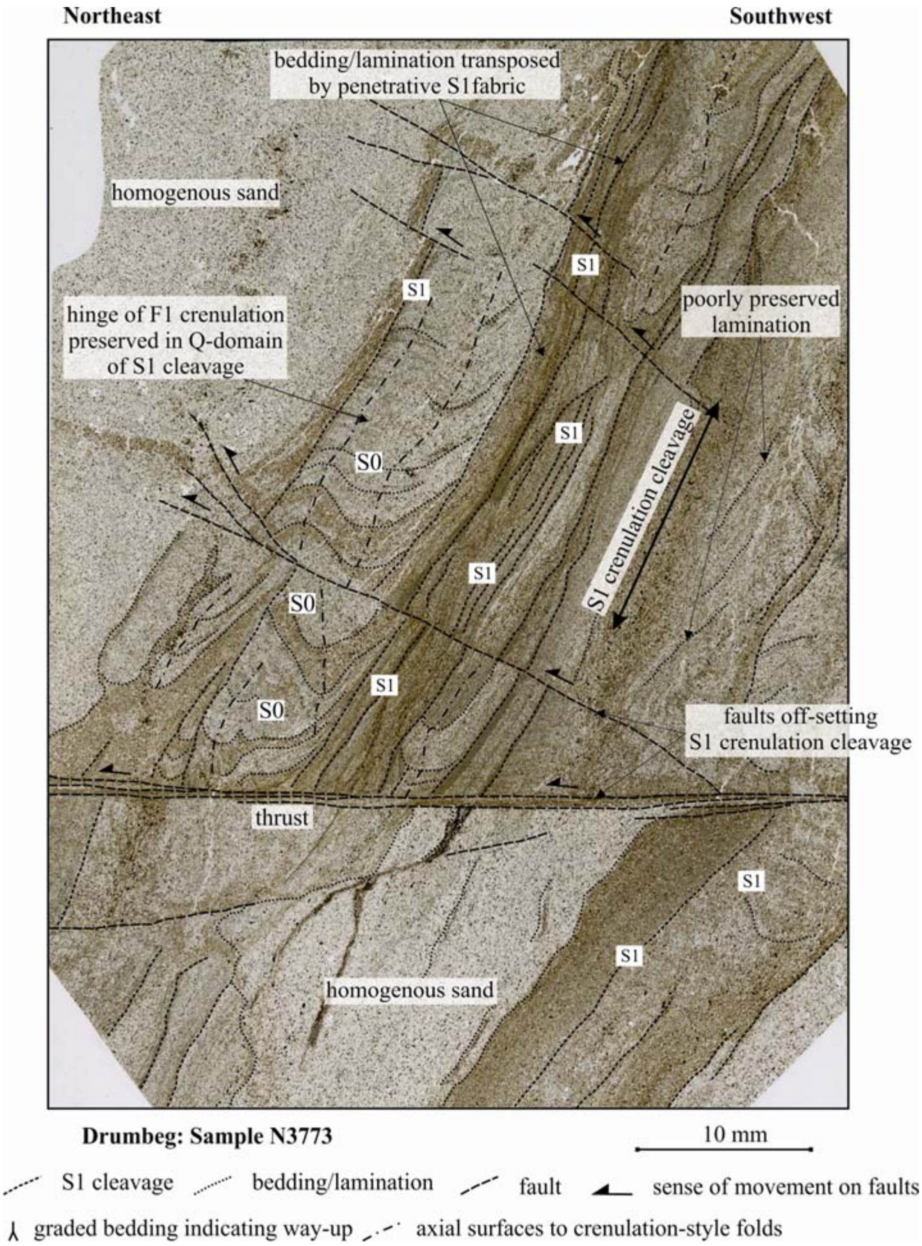


Figure 1.7. Thin section of tectonised sands from Drumbeg Quarry, Scotland (Phillips *et al*., 2007), showing a range of foliations including a crenulation cleavage

Establishing the relative age relationships displayed between successive generations of folds is critical to determining the sequential stages or phases of the **deformation history** recorded by the sediments or bedrock.

3.3. Foliations and lineations

3.3.1. Foliations

A **foliation** can be defined as a new set of planar surfaces that are penetratively developed in a body of rock or sediment as a result of deformation (Figure 1.7). In glacial sediments the foliation may be defined by a spatial variation in composition leading to a banding or layering, by a preferred alignment of elongate clasts (e.g. pebbles forming a **shape fabric foliation**) or aggregates of grains (e.g. micro-scale plasmic fabrics; see Chapter 2), or by sets of closely spaced fractures or micro-fractures (e.g. **fracture cleavage**), or any combination of these elements. The bedrock or sediment may exhibit several generations of foliation which should be distinguished chronologically using the nomenclature commonly used by structural geologists with S1 being the earliest foliation and Sn the latest, with earlier foliations being either deformed, cross-cut, or even overprinted by the later generations. A foliation formed due to the micro-scale folding of a pre-existing planar surface (e.g. bedding or foliation) is referred to as a **crenulation cleavage** (Figure 1.7). Establishing the relative age relationships displayed between successive generations of foliations is critical to determining the sequential stages or phases of the deformation history recorded by the sediments or bedrock. Bedding is usually the first recognisable planar surface in most sediments and sedimentary rocks and is designated S0. However, bedding may be **overprinted** or **transposed** by the development of a pervasive foliation. Elsewhere, a foliation may have been superimposed or imprinted onto bedding to form a **bedding-parallel foliation**. This may simply occur in response to the load exerted by the pressure of the overlying strata with the foliation forming due to compaction and dewatering, or alternatively as a result of loading by ice.

In deformed sequences a foliation may show a simple geometrical relationship to a particular set of folds, indicating that the deformation responsible for the formation of the foliation was also responsible for the folding. Consequently, a foliation, which corresponds to the plane of maximum compression or flattening, may occur parallel to sub-parallel to the axial surface of the fold (**axial planar foliation**), or form a fan-shaped structure (**foliation fan**) symmetrically arranged about the axial surface.

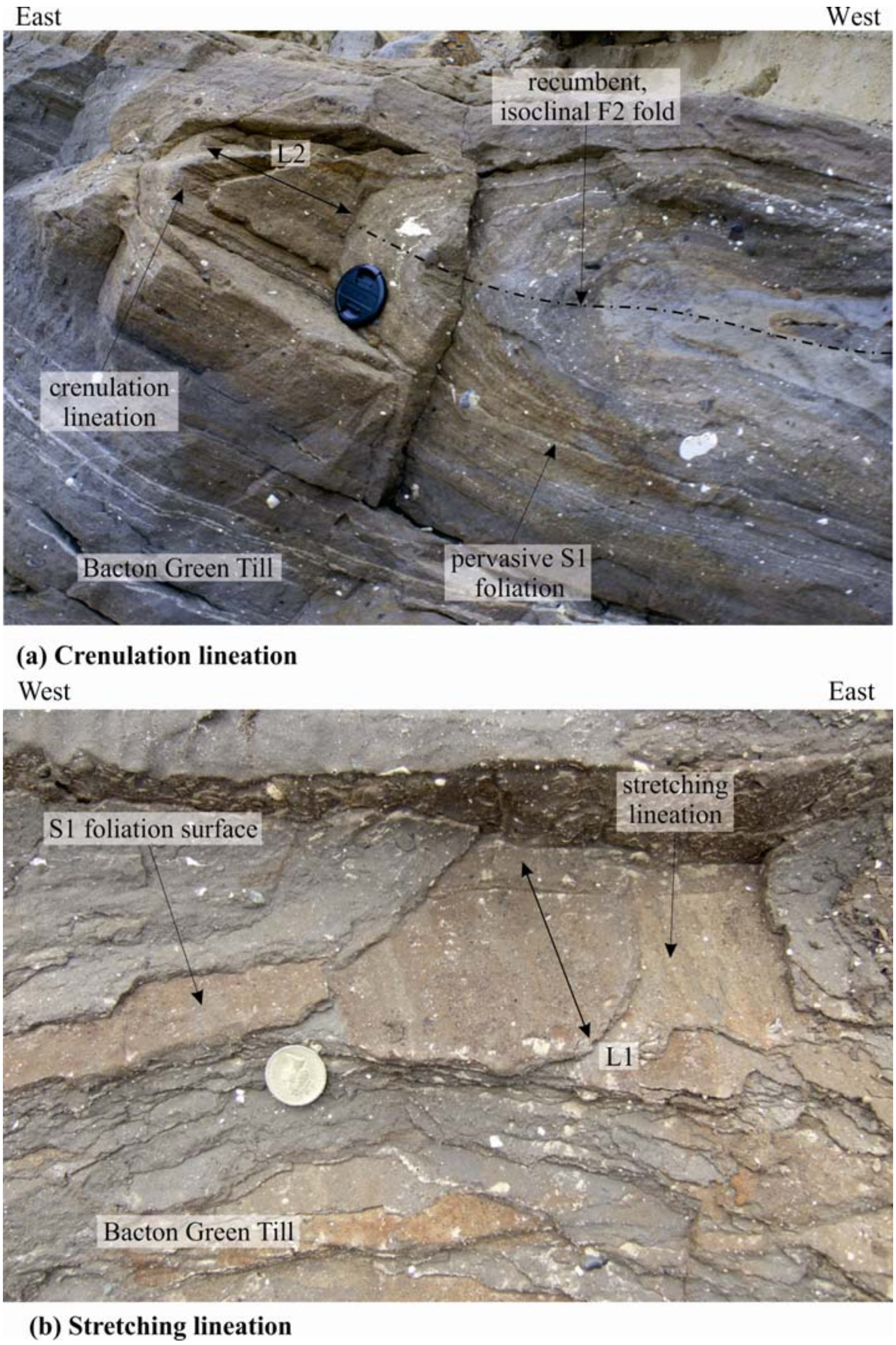


Figure 1.8 (previous page). Photographs of deformed till at West Runton, Norfolk (Phillips *et al*., 2008), showing crenulation (a) and stretching (b) lineations

Where the rock or sediment has undergone relatively intense deformation, tight to isoclinal folds may possess highly attenuated (thinned) limbs which may eventually become indistinguishable from the foliation when only the fold hinges are clearly visible. These folds are referred to as **intrafolial folds**, or where the limbs are completely detached, **rootless intrafolial folds**.

The orientation of any planar foliations can be determined by measuring its **dip** and **strike** and/or **dip** and **dip direction**, and is written in the form 220˚/22˚SE, or 22˚ → 130˚, i.e. a dip of 45˚ in the direction of 130˚ from north, respectively. The relative age (S1, S2, S3...Sn) of the foliation being measured should also be noted.

3.3.2. Lineations

A **lineation** is the linear counterpart of a foliation, and is defined as any linear feature that occurs penetratively within a body of sediment or rock. The five most important types of lineation are: *(i)* **intersection lineations** formed by intersecting foliations, e.g. a foliation intersecting a bedding surface; *(ii)* **crenulation lineations** defined by the hinge lines of micro-scale folds in a foliation plane (Figure 1.8a); *(iii)* **stretching lineations** defined by deformed linear aggregates of grains (Figure 1.8b), minerals, or deformed (stretched) objects such as mud clasts or pebbles; *(iv)* **mineral lineations** defined by the preferred shape alignment of elongate mineral grains or clasts; and *(v)* **slickenside striations** on fault surfaces showing the direction of movement.

In three dimensions, a foliation imposed on a body of rock or sediments as a result of deformation may contain an associated linear component. Consequently, several generations of lineation may be present and can be distinguished chronologically by using the nomenclature L1, L2.....to Ln, where L1 is the earliest lineation and Ln the latest. The presence of linear and/or planar foliations within highly deformed glacial sediments can potentially result in a complete spectrum from examples where only a foliation being present (**S-tectonites**), to cases where both a foliation and lineation are present (**LS-tectonites**), through to instances where only a lineation is developed (**L-tectonites**).

The orientation of a lineation can be determined by measuring its **pitch** and **pitch direction**, and is written in the form 10˚ → 110˚, i.e. a dip of 10˚ in the direction of 110˚ from north, respectively. The relative age (L1, L2, L3...Ln) and potential origin (e.g. intersection) of the lineation being measured should also be noted.

3.4. Boudinage

The process known as **boudinage** occurs when relative competent layers of rock or sediment become stretched and elongated during deformation. As a result these layers may become separated into blocks, or lens-shaped structures known as **boudins**. Where the process is incomplete the layers show a narrowing or necking, and the structure has commonly been referred to as **pinch-and-swell**. This type of structure is useful as an indicator of the extension direction within deformed sequences.

4. Shear zones and shear sense indicators

The application of the concept of 'soft deforming beds' beneath glaciers (Boulton, 1986; Boulton and Hindmarsh, 1987; Murray, 1997) and ice sheets, and the predominantly ductile nature of this deformation led directly to the concept of subglacial deformation being equated with the development of a **subglacial shear zone** (e.g. van der Wateren *et al*., 2000). In bedrock structural studies **shear zones** are defined as planar zones of relatively intense deformation produced both by homogeneous and inhomogeneous non-coaxial deformation or **simple shear**. These high strain zones occur on all scales (from a micro-scale to regional-scale crustal shear zones, 10's if not 100's kilometres across) and includes those dominated by brittle faulting (**brittle shear zones**), a combination of both brittle and ductile deformation (**brittle-ductile shear zones**), through to those resulting from only ductile deformation (**ductile shear zones**). Deformation within these zones results in the development of suite of characteristic structures (folds, fabrics) that reflect the pressure and temperature conditions, flow type, sense of movement and deformation history recorded by the shear zone. The depth of the transition between dominantly brittle and ductile deformation within the shear zone varies depending upon a variety of factors, including lithology, strain rate, thermal gradient, grain size, fluid pressure, orientation of the stress field and the existence of pre-existing planar foliations. Although most, if not all, of these factors can be directly applied to the development of a subglacial shear zone, two important factors must be borne in mind when attempting a more direct comparison with the processes occurring in ductile shear zones formed within bedrock terrains:

(i) The strains encountered in shear zones within bedrock terrains are significantly higher than those in deformed glacial sediments. Deformation structures typically equated with high shear strains (e.g. sheath folds) may be developed at much lower values of cumulative strain in glacial deposits due to elevated pore water pressures/contents encountered in these potentially water saturated sediments;

(ii) Deformation within the bedrock shear zones is accompanied by metamorphism, with ductile shearing being accompanied by a prolonged period of elevated temperatures and pressures. This leads to dynamic recrystallisation and new mineral growth, and the formation of the distinctive ductile fault rocks known as mylonites. New mineral/grain growth is not common in glacial sediments and, where present, is typically equated with the diagenetic (very low temperature) growth of carbonate minerals from solution within subglacial tills (e.g. van der Meer *et al*., 2003).

4.1. Shear-sense or kinematic indicators

Determining the direction of movement or **shear sense** on a shear zone is important as it is helps define the overall stress regime responsible for deformation. The sense of shear can be described as **sinistral** (or left-lateral), the sense of movement across the shear zone is to the left, or **dextral** (or right-lateral), where the movement is to the right. In bedrock terrains this has traditionally been established by the offset of distinct marker horizons, such as dykes, or the deflection of a well-developed layering or foliation across the shear zone. Additionally the geometry of a range of macroscopic (field observations) and microscopic (also see Chapter 3) deformation structures present within the shear zone can also be used to determine the shear sense. The most commonly used shear sense indicators are listed below and illustrated in Figure 1.9.

4.1.1. Displacement and deflection of markers

The simplest and most well known shear sense indicator is the deflection and displacement of markers such as bedding or dykes, across the shear zone (Figure 1.9a).

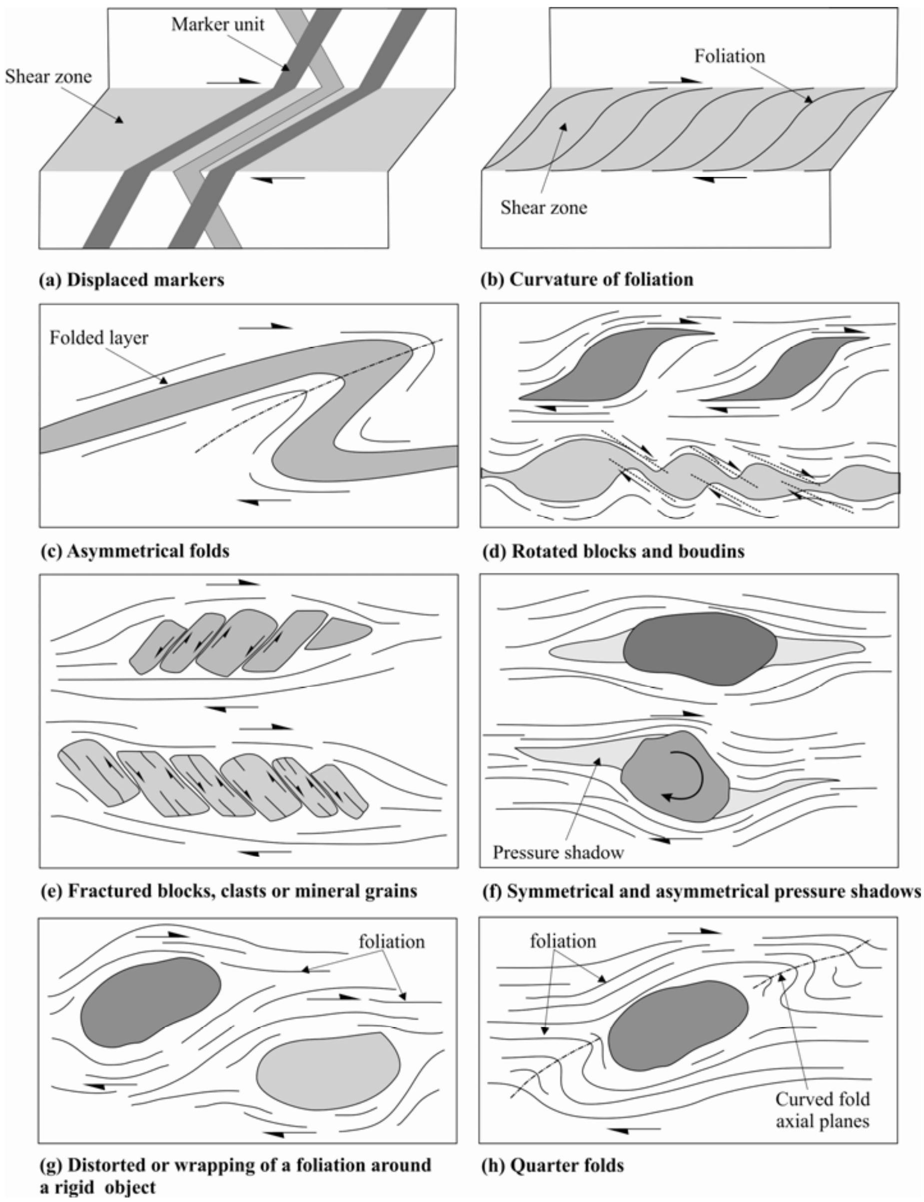


Figure 1.9. Sense of displacement indicators. (a) displaced markers; (b) curvature of foliation; (c) asymmetrical folds; (d) rotated blocks and boudinage; (e) fractured blocks, clasts or mineral grains; (f) symmetrical and asymmetrical pressure shadows; (g) distorted or wrapping foliation; (h) quarter folds

4.1.2. *Curvature of a foliation*

Foliated ductile shear zones may show a gradient in foliation development from the adjacent undeformed hanging or footwalls towards the centre of the zone. The foliation typically possess a characteristic curved or sigmoidal shape which can be used to determine shear sense (Figure 1.9b) and reflects a gradient in finite strain from a peak in the centre of the shear zone outwards towards its margins.

4.1.3. *Asymmetrical folds*

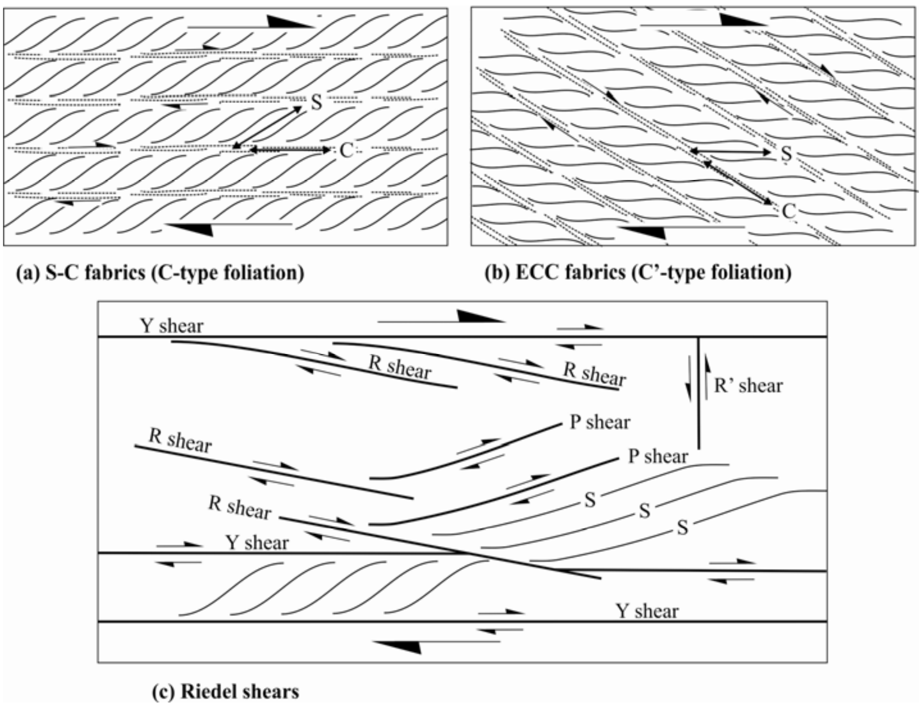
Folds developed within ductile shear zones are typically strongly asymmetrical, close, tight, or even isoclinal structures which comprise a short steeply inclined to overturned limb, and much longer more gently inclined limb. The asymmetry or shape of these folds can, with care, be used to determine the sense of shear as shown in Figure 1.9c.

4.1.4. *Rotated blocks or boudins*

Intensely deformed glacial sediments may become highly disrupted to form a **glacitectonite** in which more competent layers, such as beds of sand and/or gravel, become broken into a series of blocks or boudins which are sometimes rotated during deformation. The size of these blocks can vary from a few tens of centimetres to several 10's of metres in length, and display a variety of morphologies ranging from elongate slab-like bodies with variably rounded to tapered terminations, to more rounded 'eye' or 'augen' shaped pods with variably developed 'tails'. The sense of asymmetry of these tails and tapered terminations to the boudins may be used to establish the sense of shear (Figure 1.9d), Pressure shadows and/or the wrapping of a foliation around variably rotated boudins may also be used to determine the sense of rotation during deformation. The reorientation (tilting) of bedding and other primary sedimentary structures, such as graded bedding and cross-lamination, preserved within the boudins may also be used to determine the sense of rotation.

4.1.5. *Faulted or fractured blocks, clasts and mineral grains*

Blocks of sediment (e.g. sand) or bedrock may continue to undergone brittle deformation even when they are enclosed within a ductilely deforming matrix (e.g. till). Consequently these relatively more competent blocks may fracture and become offset as shown in Figure 1.9e. Similar features may also develop on a smaller scale as a result of the brittle deformation of pebbles or similar clastic grains, or on a micro-scale the fracturing of large mineral grains.



(c) Riedel shears

Figure 1.10. Composite foliations as indicators of shear displacement. (a) S-C fabrics; (b) ECC fabrics; (c) Riedel shears

4.1.6. Asymmetrical pressure shadows

Asymmetrical **pressure shadows** or **strain shadows** are low strain areas formed adjacent to more rigid objects (e.g. boulders, pebbles or on a micro-scale sand grains) set within a more ductile matrix (e.g. clay, silt). The asymmetry of these shadows may be used to determine the sense of shear as shown in Figure 1.9f.

4.1.7. Distorted layering or foliation

The geometry of distorted layering or foliation as it wraps around included rigid clasts or boudins within a more ductile deforming matrix can be used to establish the sense of shear (Figure 1.9h). Asymmetrical folds that deform the layering may also occur adjacent to more rigid clasts, formed as a result of the mechanical instability caused by the presence of the clast, or in response to localised deformation due to the rotation of the included block, can similarly be use to determine the sense of shear.

4.1.8. Composite foliations

Several types of foliation may be developed within shear zones as a result of progressive simple shear (Figure 1.10). These foliations can be recognised in the field, but are more commonly described in thin sections made from oriented samples collected during fieldwork (see Chapter 3). The first type of foliation is termed the **C-surface foliation** or **shear band foliation**. Two types of shear band foliations have been recognised (Passchier and Trouw 1996):

(i) C′-type foliation is related to the shearing and initially forms at an angle of 15˚ to 35˚ to the margins of the shear zone, becoming progressively rotated into parallelism with the shear zone margin as deformation continues. Any older foliation is cut by the shear bands and deflected in the same way as the foliation curvature in a larger scale shear zone. This type of foliation may superficially resemble a crenulation cleavage, but rather than being formed as a result of compression (**compressional crenulation cleavage**), forms in response to extensional deformation of the pre-existing foliation and is, therefore, often referred to as an **extensional crenulation cleavage** or **ECC fabric** (Figure 1.10b). New C′-surfaces continue to developed, are rotated and may be progressively overprinted as deformation within the shear zone continues. This process may explain the complexity of deformation observed in some subglacial traction tills and glacitectonites.

(ii) C-type foliations form part of the so-called **S-C** or **C-S fabrics** (Figure 1.10a). Pre-existing or earlier formed foliations, termed **S-surface** (S = *schistosité*) **foliations**, within the sediment or rock are crosscut and deformed by the developing **C-surface** foliation or **C planes** (C = *cisaillement*, French for shear) which occur parallel to the margins of the shear zone. The S-surfaces preserved are preserved as a distinctive sigmoidal or S-shaped foliation between the developing C-surfaces and can be used to determine the sense of shear (Figure 1.10). S-C foliations are thought to reflect inhomogeneous simple shear with both the S and C surfaces continuing to develop during deformation. S-C foliations may be crosscut or even overprinted by C′-type shear band foliation.

4.1.9. Porphyroclast systems

Porphyroclasts occur within mylonites and similar ductile fault rocks where they are composed of relatively large, single crystals (e.g. feldspar) or small, polycrystalline rock fragments set within a finer grained matrix. They are

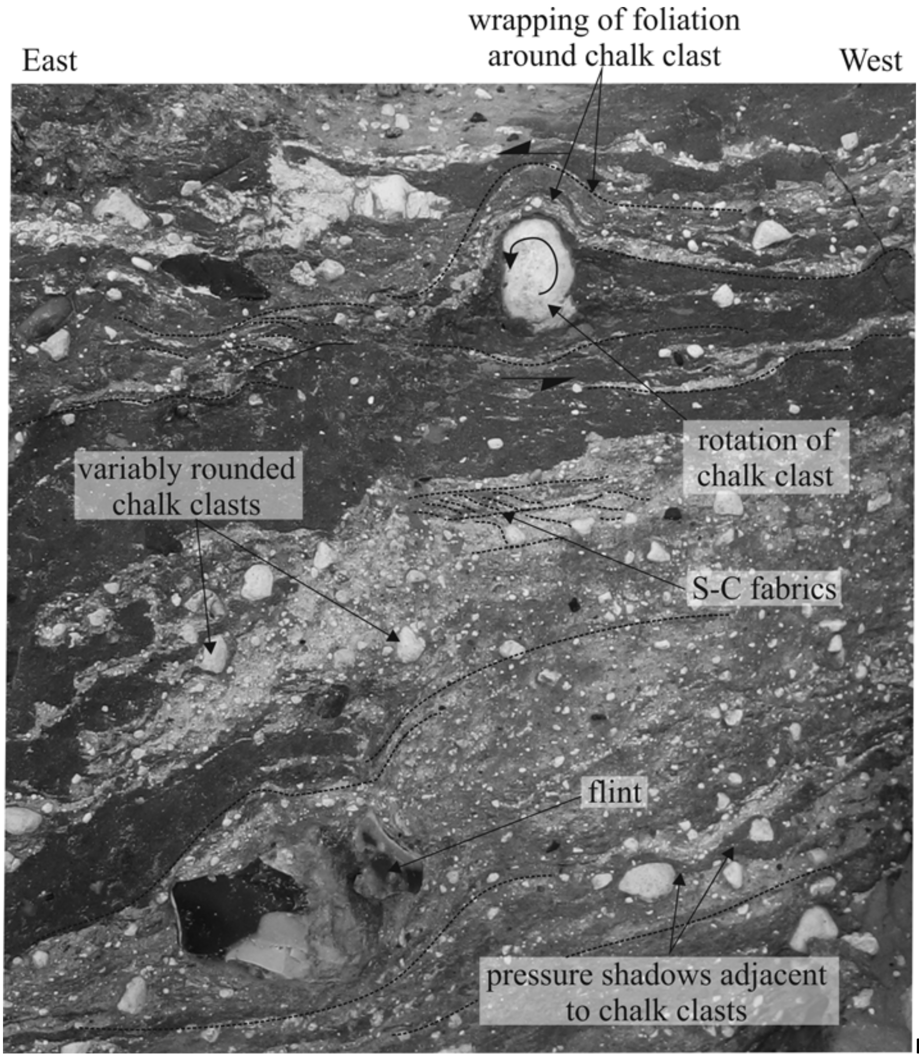


Figure 1.11. Photograph of a highly deformed chalk-rich till at East Runton, Norfolk (Burke *et al*., 2008), showing a well-developed compositional layering, rotated chalk clasts, S-C fabrics and pressure shadows developed adjacent to the variably rounded chalk fragments

inferred to have been formed by the breaking ('clasis') of larger pre-existing grains leading to a reduction of the grain size of the rock as a result of deformation. These microstructures are commonly enclosed within tapering tails or wings composed of finer grained material known as **mantles** with the whole structure being referred to as a **mantled porphyroclast** or **porphyroclast system**. The tails are stretched out into the plane of main foliation within the shear zone, with their sense of asymmetry being used as a shear sense indicator. The difference in elevation of the tails on either side of the porphyroclast is known as **stair-stepping**. Five types of porphyroclast system, based upon the morphology and distribution of the porphyroclast tails, have been recognised in mylonitic rocks (see Passchier and Trouw, 1996).

Although the cumulative strains recorded by subglacially deformed sediments are much lower than those responsible for the formation of porphyroclast systems within mylonites, similar looking structures do occur. For example, Figure 1.11 shows a fallen block of highly deformed clay-rich till containing rounded to elliptical chalk fragments enclosed within thin, highly attenuated tails of fine-grained chalk. These tails are stretched out into the plane of a well-developed foliation defined by pale coloured, chalk-rich layers. The asymmetrical tails developed on the chalk clasts can be used to establish the sense of shear as shown in Figure 1.11.

4.1.10. *Riedel Shears*

The term **Riedel shears** is used to describe several sets of shear fractures developed within near surface brittle faults. They are divided into R, R′, P and Y shears, each set possesses a characteristic orientation, relative to bounding surfaces of the fault or shear zone, and sense of shear (Figure 1.10c). **Y shears** occur parallel to the boundaries of the shear zone or may form the bounding brittle faults. Riedel shear superficially resemble shear bands developed in ductile shear zones, but are formed in response to brittle fracturing and possess a slightly different geometry and orientation. These brittle structures can be used to establish the sense of shear, developing during the later stages of deformation within the subglacial shear zone as the systems begins to lock-up in response to decreasing porewater content and/or pressure.

5. Graphical representation of field structural data

The **lower hemisphere stereographic projection** (Figure 1. 12), or more simply **stereonet**, is a method used by structural geologists to represent and analysing 3D orientation data of lines and planes collected in the field (e.g. strike and dip of bedding...etc) in a 2D graphical form. It can be used to determine the angular relationships between planar features such as bedding and a foliation in 3D space, or establish the plunge and trend of large scale folds from orientation data of bedding recorded from the fold

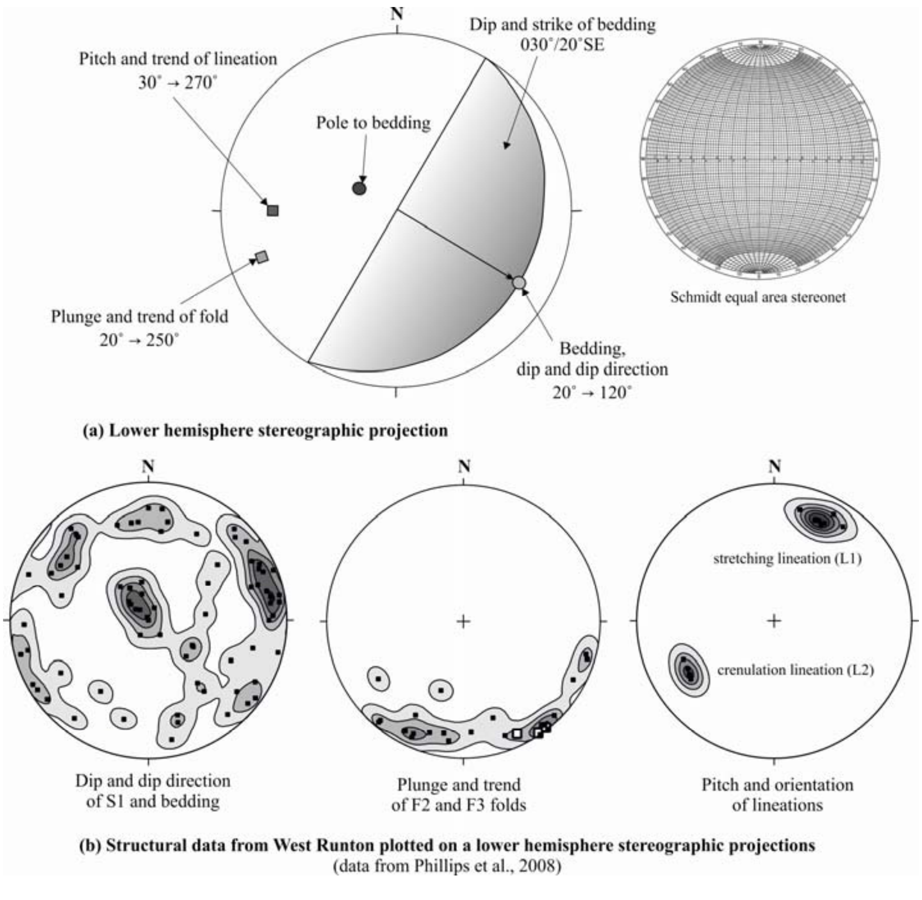


Figure 1.12. Methods for presenting structural data. (a) Lower hemisphere stereographic projection; (b) structural data from West Runton plotted on a lower hemisphere stereographic projection

limbs in the field. The process of plotting data onto a stereonet is covered in detail in several of the structural geology textbooks (e.g. Ragan, 1985; Hatcher, 1995) listed in the further reading section at the end of this chapter. Consequently, this section only provides a basic introduction to this graphical method of representing field structural data. Examples of a stereonets plotted for a range of structural data collected in the field are shown in Figure 1.12b.

Two types of stereographic projection are commonly used: *(i)* the **Wulff** stereonet or equal angle net that can be used to solve angular relationships; and *(ii)* the **Schmidt** stereonet or equal area net, which can be used to solve both angular relationships and statistically analyse larger orientation datasets using contoured stereographic projections. Plotting lines and planes on either stereonet projection is carried out in exactly the same way. Typically linear structures such as fold axes, are plotted as points, and planar features, such as bedding, foliations or faults, plotted as arcs. To simplify the process of plotting planes it is common to plot the pole to each plane. The pole is an imaginary line projected perpendicular to the plane running through the centre of the projection sphere, and occurs at 90˚ to both the strike line and dip line (Figure 1.12a). Consequently the pole to a plane always plots in the opposite quadrant of the stereonet from the dip of the plane, which can be confusing for students new to the field of glacitectonic research. An alternative approach is to plot the dip and dip direction of planar features as a point making it easier to visualise the orientation of the planes (see Figure 1.12b).

Several computer packages are available which allow large numbers of readings to be plotted relatively quickly, and for calculating geometrical relationships, contouring of large data sets as an aid to determining statistical distribution of features. Stereonets can be added to annotated field photographs, section diagrams, cross sections, or geological maps to provide a 3D basis for the interpretation of glacitectonic structures.

6. Suggested further reading

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