4

SEDIMENTARY ROCK TEXTURE

4.1 Introduction

Sediment texture is concerned with the grain-size and its distribution, morphology and surface features of grains, and the fabric of the sediment. Induration and weathering (see Section 4.7), and colour (see Section 4.8) are also considered in this chapter.

Texture is an important aspect in the description of sedimentary rocks and can be useful in interpreting the mechanisms and environments of deposition. It is also a major control on the porosity and permeability of a sediment. The texture of many sedimentary rocks can only be studied adequately with a microscope and thin-sections. With sand and silt-sized sediments you cannot do much more in the field than estimate grain-size and comment on the sorting and roundness of grains. With conglomerates and breccias, the size, shape and orientation of grains can be measured accurately in the field; in addition, surface features of pebbles and the rock's fabric can be examined quite easily. A checklist for a sediment's texture is given in Table 4.1.

4.2 Sediment Grain-Size and Sorting

The most widely accepted and used grain-size scale is that of Udden–Wentworth (Table 4.2). For more detailed work, phi units (ϕ) are used; phi is a logarithmic transformation: $\phi = -\log_2 S$, where S is grain-size in millimetres.

For sediments composed of sand-sized particles, use a hand-lens to determine the dominant grain-size class present; it is usually possible to distinguish between very coarse, coarse, medium, fine and very fine sand classes. Comparison can be made with the sand-sizes depicted in Figure 4.1. For finer-grained sediments, chew a tiny piece of the rock; silt-grade material feels gritty between the teeth compared with clay-grade material, which feels smooth.

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Table 4.1 *Checklist for the field examination of sedimentary rock texture.*

1. **Grain-size, sorting and size-grading:** estimate in all lithologies: see Table 4.2 and Figures 4.1, 4.2 and 5.37. In conglomerates, measure maximum clast size and bed thickness; check for correlation
2. **Morphology of constituent grains:**
   Shape of grains: see Figure 4.4 (important for clasts in conglomerates); look for facets on pebbles, and striations (Figure 4.11)
   Roundness of grains: see Figure 4.5
3. **Fabric:**
   (a) Look for preferred orientation of elongate clasts in conglomerates and fossils in all lithologies (see Figures 4.6, 6.7 and 6.8); measure orientations and plot rose diagram (see Chapter 7)
   (b) Look for imbrication of clasts or fossils (see Figures 4.6 and 4.7)
   (c) Examine matrix-grain relationships, especially in conglomerates and coarse limestones; deduce whether the sediment is matrix-supported or grain-supported (see Figure 4.8)
   (d) Look for deformation of pebbles (compacted, fractured, split, pitted)

With chemical rocks such as evaporites, recrystallised limestones and dolomites, it is crystal size that is being estimated, rather than grain-size. Terms for crystal size are given in Table 4.3.

For accurate and detailed work, particularly on siliciclastic sediments, various laboratory techniques are available for grain-size analysis, including sieving of poorly cemented sedimentary rocks or modern sediments, point-counting of thin-sections of rocks and sedimentation methods (see Recommended Reading).

In the field only a rough estimate can be made of *sorting* in a sand-grade sediment. Examine the rock with a hand-lens and compare it with the sketches in Figure 4.2.

The grain-size of a sediment may fine- or coarsen-upwards through the bed to give a *graded bed*. Normal graded bedding is most common,

Table 4.2 *Terms for grain-size classes (after J.A. Udden and C.K. Wentworth) and siliciclastic rock types. For sand-silt-clay mixtures and gravel-sand-mud mixtures see Figure 3.1.*

ϕ			
		boulders	conglomerates (rounded clasts)
−8	256 mm		
		cobbles	and
−6	64		
		pebbles	breccias
−2	4		(angular clasts)
		granules	
−1	2 mm		
		v. coarse	
0	1		
		coarse	
1	500 μm		
		medium	SANDSTONE
2	250		
		fine	
3	125		
		v. fine	
4	63 microns		
		v. coarse	MUDROCKS
5	32		types:
		coarse	mudstone
6	16 SILT	SILTSTONE	shale
		medium	marl
7	8		argillite
		fine	slate
8	4 microns		
	CLAY	CLAYSTONE	

with the coarsest particles at the base, but inverse (or reverse) grading also occurs, with a coarsening up of grains. Often this is just in the lower part of a bed and then normal grading takes over. In some instances a bed may show no grain-size sorting at all. Composite graded bedding denotes a bed with several fining-upward units within it. See Section 5.3.4 for more information.

In a broad sense, the grain-size of siliciclastic sediments reflects the hydraulic energy of the environment: coarser sediments are transported

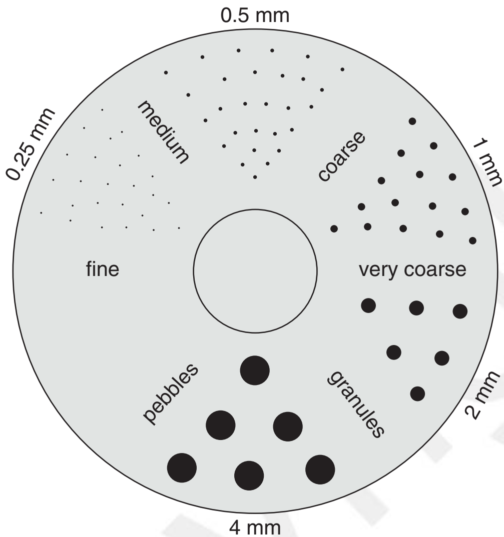

Figure 4.1 *Chart for estimating grain-size of sands: medium sand is 0.25–0.5 mm diameter, coarse sand is 0.5–1 mm diameter, and so on. Place a small piece of the rock or some grains scraped off the rock in the central circle and use a hand-lens to compare and deduce the size.*

Table 4.3 *Informal terms for describing crystalline rocks.*

	very coarsely crystalline
1.0 mm	
	coarsely crystalline
0.5 mm	
	medium crystalline
0.25 mm	
	finely crystalline
0.125 mm	
	very finely crystalline
0.063 mm	
	microcrystalline
0.004 mm	
	cryptocrystalline

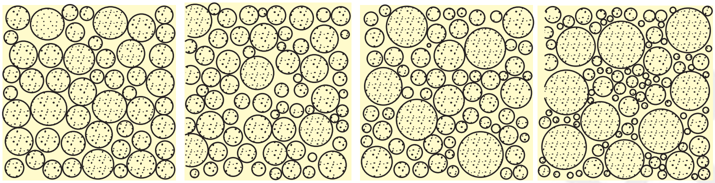

Figure 4.2 *Charts for visual estimation of sorting.*

and deposited by faster-flowing currents than those conveying finer sediments; mudrocks tend to accumulate in quieter water. The sorting of a sandstone reflects the depositional process, and this improves with increasing agitation and reworking. In contrast, the grain-size of carbonate sediments generally reflects the size of the organism skeletons and calcified hardparts that make up the sediment; these can also be affected by currents of course. Sorting terms can be applied to limestones, but bear in mind that some limestone types, oolitic and peloidal grainstones, for example, are well sorted anyway, so that the sorting terms do not necessarily reflect the depositional environment.

For grain-size and sorting of conglomerates and breccias see Section 4.6.

4.3 Grain Morphology

The morphology of grains has three aspects: *shape* (or form), determined by various ratios of the long, intermediate and short axes; *sphericity*, a measure of how closely the grain shape approaches that of a sphere; and *roundness*, concerned with the curvature of the corners of the grain.

For *shape*, four classes are recognised – spheres, discs, blades and rods, based on ratios involving the long (L), intermediate (I) and short axes (S) (Figures 4.3 and 4.4). These terms are useful for describing clast shape in conglomerates and breccias and can be applied with little difficulty in the field. The shape of pebbles is largely a reflection of the composition and any planes of weakness, such as bedding/lamination, cleavage or jointing in the rock. Rocks of a very uniform composition and structure, such as many granites, dolerites and thick sandstones, will give rise to equant/spherical pebbles; thin-bedded rocks will generally

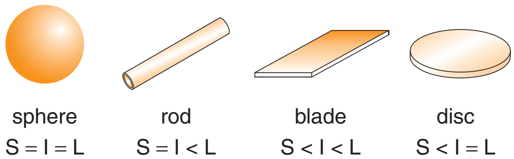

Figure 4.3 *The four common shapes of pebbles. S, I and L are the short, intermediate and long diameters, respectively.*

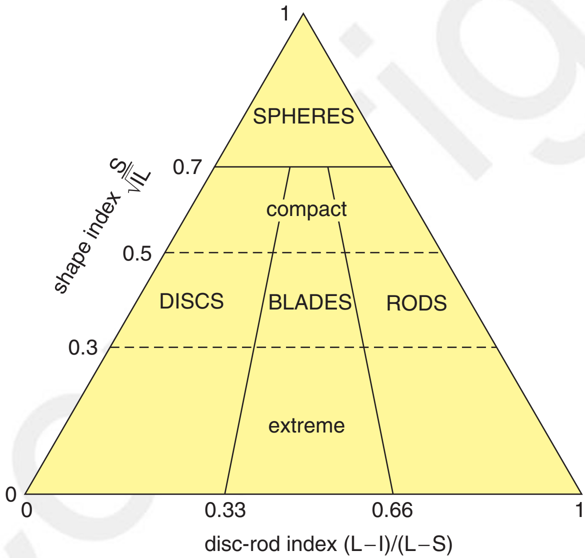

Figure 4.4 *The four classes of grain or clast shape based on the ratios of the long (L), intermediate (I) and short (S) diameters.*

form tabular and disc-shaped clasts; and highly cleaved or schistose rocks, such as slates, schists or some gneisses, will generally form bladed or rod-shaped pebbles.

Formulae are available for the calculation of sphericity and roundness (see Recommended Reading). *Roundness* is more significant than sphericity as a descriptive parameter and for most purposes the simple terms of Figure 4.5 are sufficient. These terms can be applied to

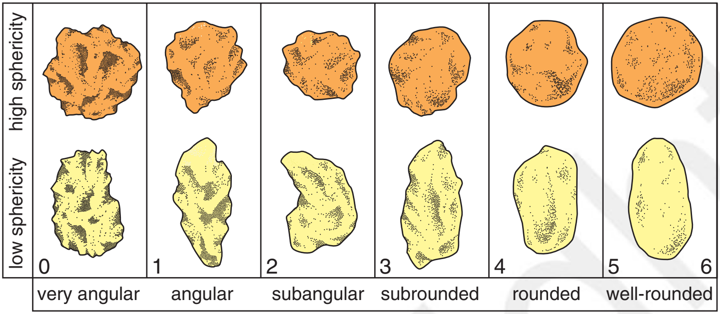

Figure 4.5 *Categories of roundness for sediment grains. For each category a grain of low and high sphericity is shown.*

grains in sandstones and to pebbles in conglomerates. In general, the roundness of grains and pebbles is a reflection of transport distance or degree of reworking.

The roundness terms are less environmentally meaningful for grains in a limestone since some, such as ooids and peloids, are well rounded to begin with. Skeletal grains in a limestone should be checked to see if they are broken or their shape has been modified by abrasion.

4.4 Sediment Fabric

Fabric refers to the mutual arrangements of grains in a sediment. It includes the *orientation* of grains and their *packing*. Fabrics may be produced during sedimentation or later during burial and through tectonic processes.

In many types of sedimentary rock a *preferred orientation* of elongate particles can be observed. This can be shown by prolate pebbles in a conglomerate or breccia, and fossils in a limestone (see, e.g., Figure 6.6), mudrock (see, e.g., Figure 6.7) or sandstone; such features are visible in the field. Many sandstones show a preferred orientation of elongate sand grains but microscopic examination is required to demonstrate this.

Preferred orientations of particles arise from interaction with the depositional medium (water, ice, wind), and can be both parallel to (the more common), and normal to, the flow direction (Figure 4.6).

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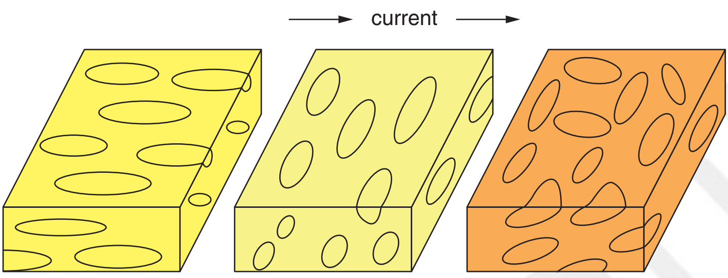

Figure 4.6 *Orientations of grains and pebbles: parallel to current, normal to current, and imbricated.*

Measurement of pebble, fossil or grain orientations can thus indicate the palaeocurrent direction (see Section 7.3.4). With pebbles it is best to measure clasts that have a clear elongation; a length-to-width ratio of more than 3:1 is acceptable. Preferred orientations can also be tectonically induced so if you are working in an area of moderate deformation, also measure fold axes, cleavage and lineations. Pebbles may be rotated into the tectonic direction. Look for pressure shadows and the development of fibrous minerals at the ends of the pebble.

Tabular and disc-shaped pebbles or fossils commonly show *imbrication*. In this fabric, they overlap each other (like a pack of cards), dipping in an upstream direction (Figures 4.6 and 4.7). This can be a useful texture for deducing the palaeocurrent direction (see Section 7.3.4).

The amount of fine-grained matrix and the matrix-grain relationship affect the *packing* and fabric of a sediment and are important in interpretations of depositional mechanism and environment. Where grains in a sediment are in contact, the sediment is *grain-supported*; matrix can occur between the grains, as can cement (Figures 4.7, 4.8 and 4.9). Where the grains are not in contact, the sediment is *matrix-supported* (Figures 4.8 and 4.10). Also look at the matrix between the large clasts in coarser sediments; this may be well-sorted or poorly sorted (i.e. the sediment as a whole may be bimodal or polymodal in grain-size; see Figure 4.8).

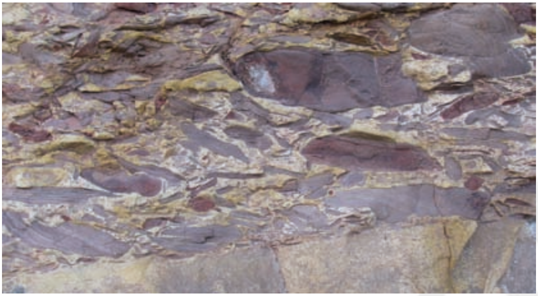

Figure 4.7 *Conglomerate with a sharp-base, clast-support fabric and well-developed imbrication (elongate, flat clasts dipping down to the right) indicating transport to the left. Clasts are mudstone fragments, and so intraformational. Fluvial facies, Upper Carboniferous, NE England.*

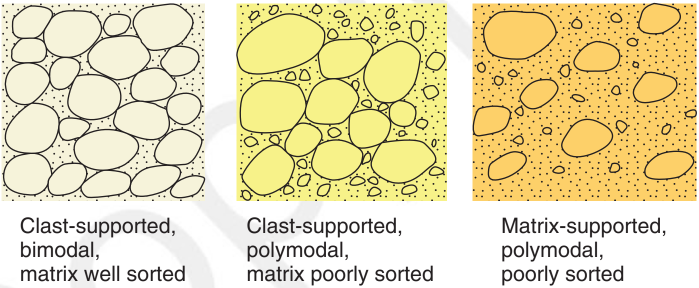

Figure 4.8 *Grain fabric and sorting: clast-support with well-sorted and poorly sorted matrix, and matrix support.*

With sandstones and limestones, a grain-support fabric with no mud generally indicates reworking by currents and/or waves/wind, or deposition from turbulent flows where suspended sediment (mud) is separated from coarser bed load. Limestones with a matrix-support fabric, such as a wackestone (see Table 3.3), mostly reflect quiet-water sedimentation. Rudstone and floatstone are coarse limestones with a grain-support and matrix-support fabric respectively (see Figures 3.10 and 4.8).

The fabric of conglomerates and breccias is discussed further in Section 4.6.

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Figure 4.9 *Conglomerate with a matrix-support fabric and subangular to subrounded pebbles. Tillite (ancient glacial deposit), Late Precambrian, Scotland.*

Figure 4.10 *Polymictic conglomerate with pebble-support fabric, occurring above a massive sandstone and overlain by a sandstone with scattered pebbles and then several other thin conglomerates. Thickness of section 2 m. Braided-stream fluvial facies. Devonian, Bungle Bungles, Western Australia.*

4.5 Textural Maturity

The degree of sorting, the roundness and the matrix content in a sandstone contribute towards the textural maturity of the sediment. Texturally immature sandstones are poorly sorted with angular grains and some matrix, whereas texturally supermature sandstones are well-sorted with well-rounded grains and no matrix. Textural maturity generally increases with the amount of reworking or distance travelled; for example, aeolian and beach sandstones are typically mature to supermature, whereas fluvial sandstones are less mature. Textural maturity is usually matched by a comparable compositional maturity (see Section 3.2). It should be remembered that diagenetic processes can modify depositional texture. An estimate of the textural maturity of a sandstone can be made in the field by close examination with a hand-lens.

4.6 Texture of Conglomerates and Breccias

There is no problem with measuring the grain-sizes of these coarser sediments in the field; a ruler or tape measure can be used. With conglomerates and breccias, it is the *maximum clast size* that is usually measured. There are several ways of doing this, but one method is to take the average of the 10 largest clasts in a rectangular area of 0.5×0.5 m. It can be useful to estimate modal size as well for a conglomerate bed. Measure the long axes of 20–30 pebbles; plot a histogram and determine the size of the dominant pebbles. Maximum clast size is used as a parameter since with many rudites this is a reflection of the competency of the flow.

It is also useful to measure the *bed thickness* of conglomerates. This may vary systematically up through a succession, increasing or decreasing upwards, reflecting an advance or retreat of the source area. With some transporting and depositing processes (e.g. mudflows and stream floods) there is a positive correlation between maximum particle size and bed thickness. With braided stream conglomerates there is no such relationship.

Maximum particle size and bed thickness generally decrease down the transport path. Measurements of maximum particle size and bed thickness from conglomerates over a wide area or from a thick vertical succession may reveal systematic variations, which could be due to changes in the environment and the amount and type of sediment being supplied, and these may reflect fundamental changes involving climate or tectonics.

For the *grain-size distribution* in coarse sediments the sorting terms of Figure 4.2 can be applied, but in many cases these terms are inappropriate since the distribution is not unimodal. Many conglomerates are bimodal or polymodal in their grain-size distribution if the matrix between pebbles is considered (see Figure 4.8). It is also important to check grain-size variations through a conglomerate bed. Normal size-grading of pebbles through a bed is common but inverse/reverse grading can also occur, particularly in the basal part (see Figure 5.37 and Section 5.3.4). In some rudites, such as those deposited by debris flows, large clasts occur towards the top of the bed; these were carried there by the upward buoyancy of the flow.

The *shape and roundness* of pebbles can be described by reference to Figures 4.3 and 4.4. Taken over a large area or up a thick succession, there may be significant changes in the degree of roundness of pebbles. This can be related to the length of the transport path. With regard to shape, some pebbles of desert and glacial environments possess flat surfaces, *facets* arising either from wind abrasion (such pebbles are known as ventifacts or dreikanters) or glacial abrasion. A characteristic feature of pebbles in a glacial deposit is the presence of *striations* (Figure 4.11), although they are not always present.

The shape of pebbles may be modified during burial and through tectonic deformation. Clasts of mudrock, especially those of intraformational origin, may be folded, bent, deformed and fractured during compaction. Where there is a lot of overburden, there may be sutured contacts (stylolites) between clasts as a result of pressure dissolution (see Section 5.5.7), or one pebble may be forced into another to produce a concave pit. During more intense deformation and metamorphism, pebbles may be flattened and stretched out.

Figure 4.11 *Pebble, 12 cm across, with striae from a glacial diamictite. Permian, Western Australia.*

Attention should be given to the *fabric* of the conglomerate; in particular, check for preferred orientations of elongate clasts (if possible measure several tens, or more, of long axes) and look for *imbrication* of prolate pebbles (long axes parallel to current and dipping upstream; see Figures 4.6 and 4.7). If exposures are very good then the dip angle of the long axis relative to the bedding can be measured to give the angle of imbrication. In fluvial and other conglomerates a normal-to-current orientation is produced by rolling of pebbles, while the parallel-to-current orientation arises from a sliding of pebbles. In glacial deposits, the orientation of clasts is mostly parallel to the direction of ice movement. Glacial diamictites that have been subjected to periglacial conditions of freeze and thaw may contain split boulders.

4.6.1 Limestone breccias

In limestones some breccias are the result of in situ brecciation processes; this is the case with some karstic breccias (see Section 5.4.1.5), brecciated hardgrounds and tepees (see Sections 5.4.3 and 5.4.4), brecciated soils (calcretes; see Section 5.5.6.2) and collapse breccias formed through dissolution of intrastratal evaporites (Section 3.6, see Figure 3.31).

Examine the pebble-matrix relationship (see Section 4.4). Pebble-support fabric (Figures 4.7, 4.8 and 4.10) is typical of fluvial and beach gravels; matrix-support fabric (Figure 4.8) is typical of debris-flow deposits (debrites), which may be subaerial (as in alluvial fans or in volcanic areas; see Section 3.11 and Figure 3.30) or submarine (as in slope aprons/fans). Glacial deposits, tills and tillites, deposited directly from glacial ice, are also generally matrix-supported (Figure 4.9) and debris-flow deposits are commonly associated (the terms diamict/diamicton and diamictite are often applied to muddy gravel/conglomerate with some glacial connection; see Section 3.3).

4.7 Induration and Degree of Weathering

The induration or hardness of a sedimentary rock cannot be quantified easily. It depends on the lithology, as well as the degree of cementation, the burial history, stratigraphic age, and so on. Induration is an important concept since it does affect the degree of weathering of a rock, along with topography, climate and vegetation. A well-indurated rock in the subsurface may be rendered very friable at the surface as a result of weathering. Calcite cements in a sandstone, for example are easily dissolved out at the surface, as are feldspar grains and calcareous fossils. Some sandstones at surface outcrop are friable and full of holes from decalcification. On the other hand, some rocks, such as limestones, become harder on surface exposure ('case hardening'). A qualitative scheme can be used for describing induration (Table 4.4).

4.7.1 Rock exposures and outcrops

The way in which sedimentary rocks appear at outcrop can give useful information on sediment lithology, in particular the vertical changes up the succession. Mudrocks are generally less well exposed than sandstones and limestones since they are usually less well indurated and soils develop more easily upon them. Thus in cliff and mountainside exposures, sandstones and limestones tend to stand out relative to mudrocks, which weather in or are covered in vegetation. Sandstones and limestones generally give rise to steeper slopes than mudrocks. Bedding-normal joints and fractures are more common in sandstones and limestones than in mudrocks and give rise to vertical cliffs in horizontal strata. The presence of cycles in a succession, and the fining-upward or coarsening-upward of sediments in a sequence may be revealed as a result of this differential response to weathering (see, e.g., Figure 8.1).

Look at a cliff or hillside carefully; the nature of the outcrop, even if poor, the slope profile and distribution of vegetation may all give important clues to the lithologies present and upward trends and changes.

Table 4.4 *A qualitative scheme for describing the induration of a sedimentary rock.*

Unconsolidated	Loose, no cement whatsoever
Very friable	Crumbles easily between fingers
Friable	Rubbing with fingers frees numerous grains
	Gentle blow with hammer disintegrates sample
Hard	Grains can be separated from sample with penknife
	Breaks easily when hit with hammer
Very hard	Grains are difficult to separate with a penknife
	Difficult to break with hammer
Extremely hard	Sharp, hard hammer blow required
	Sample breaks across most grains

4.7.2 Weathering and alteration of sediments and rocks

The state of weathering of sediments and rocks is an important aspect of description and can give useful information on climate, present and past, and length of exposure, as well as on the degree of alteration and loss of strength for engineering purposes (see British Standards Institute, 1981). All sediments and rocks are weathered to various extents when exposed to the elements at the Earth's surface, and eventually soils with A and B zones may develop with vegetation. The weathered zone of the rocks beneath the soil is zone C. The weathering of rocks leads to discoloration, decomposition and disintegration.

Weathering features can be looked for in present-day exposures as well as in the rock record beneath unconformities. The soils above weathered zones may well be removed by subsequent erosion and so not preserved. Weathering features and soils seen at outcrop today may not be currently forming but be relict, the result of processes in the past when climate was different.

Weathering of sediments and rocks takes place through both mechanical and chemical processes with climate mostly controlling the degree of each. Mechanical weathering (temperature changes, wetting-drying) results in the opening of fractures and discontinuities and creation of new ones, at both the rock and crystal scale. Chemical weathering causes discoloration of the rock, alteration of grains, as of many silicate minerals to clays, and dissolution of grains – especially carbonates (fossils and calcite cements), and even the rock itself, leading to potholes, caverns and karst (see Section 5.4.1.5). Dissolution of limestone may lead to the residue being left behind – a quartz sand or mud, as in terra rossa soil. A weathering scale, which can be adapted to your local situation, is shown in Table 4.5 and Figure 4.12. All degrees of weathering may occur in one profile, with the A and B horizons of the soil above, or a profile may just show the lower levels as a result of erosion. Figure 4.13 shows a well-developed weathering profile.

4.8 Colour of Sedimentary Rocks

Colour can give useful information about lithology, depositional environment and diagenesis. For many purposes a simple estimate of the colour is sufficient, although it is amazing how one person's subjective impression of colour can vary from another. For detailed work, a colour chart can be used; there are several widely available including

Table 4.5 Scale of weathering of sediment and rocks.

Term	Description	Grade
Fresh	No visible sign of rock weathering; perhaps slight discoloration on major discontinuity surfaces	I
Slightly weathered	Discoloration indicates weathering of rock material and discontinuity surfaces. All the rock may be discoloured by weathering	II
Moderately weathered	Less than half of the rock material is decomposed or disintegrated to a soil. Fresh or discoloured rock is present either as a continuous framework or as corestones	III
Highly weathered	More than half of the rock material is decomposed or disintegrated to a soil. Fresh or discoloured rock is present either as a discontinuous framework or as corestones	IV
Completely weathered	All rock material is decomposed and/or disintegrated to soil. The original structure is still largely intact	V
Residual soil	All rock material is converted to soil. The rock structure and material fabric are destroyed. There may be a change in volume, but the soil has not been transported significantly	VI

one from the Geological Society of America based on the 'Munsell Colour System'.

It is obviously best to measure the colour of a fresh rock surface, but if different, also note the colour of the weathered surface. The latter can give an indication of the rock's composition, for example in terms of iron content.

Two factors determine the colour of many sedimentary rocks: the oxidation state of iron and the content of organic matter. Iron exists

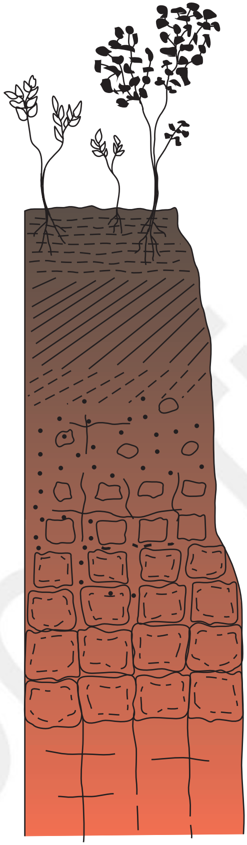

Figure 4.12 *Weathering zones of bedrock beneath a soil.*

in two oxidation states: ferric (Fe^{3+}) and ferrous (Fe^{2+}). Where ferric iron is present it is usually as the mineral hematite, and even in small concentrations of less than 1% this imparts a *red colour* to the rock. The formation of hematite requires oxidising conditions, and these are frequently present within sediments of semi-arid continental environments. Sandstones and mudrocks of these environments (deserts, playa lakes and rivers) are commonly reddened through hematite pigmentation

Figure 4.13 *Weathering profile upon a red calcareous mudrock. Northern Territories, Australia.*

(developed during early diagenesis) and such rocks are referred to as '*red beds*' (see, e.g., Figures 3.4b, 3.7, 4.10 and 5.3). However, red marine sedimentary rocks, for example some pelagic limestones (e.g. Ammonitic Rosso, see Figure 6.10), are also known.

Where the hydrated forms of ferric oxide, goethite or limonite are present the sediment has a *yellow-brown* or *buff colour*. In many cases, yellow-brown colours are the result of recent weathering and hydration-oxygenation of ferrous iron minerals such as pyrite or siderite, or ferroan calcite or ferroan dolomite.

Where reducing conditions prevailed within a sediment, the iron is present in a ferrous state and generally contained in clay minerals; ferrous iron imparts a *green colour* to the rock. Green colours can develop through reduction of an originally red sediment, and vice versa (see

Figure 3.14). With red- and green-coloured deposits see if one colour, usually the green, is restricted to, say, coarser horizons or is concentrated along joint and fault planes; this would indicate later formation through the passage of reducing waters through the more permeable layers or conduits.

Organic matter in a sedimentary rock gives rise to *grey colours*, and with increasing organic content to a *black colour*. Organic-rich sediments generally form in anoxic conditions. Finely disseminated pyrite also gives rise to a dark grey or black colour. *Black pebbles*, which are reworked out of soil horizons and may be the result of forest fires, are commonly associated with unconformities and exposure horizons.

Other colours such as olive and yellow can result from a mixing of the colour components. Some minerals have a particular colour and if present in abundance they can impart a strong colour to the rock; for example glauconite and berthierine-chamosite give rise to green-coloured sediments. Anhydrite, although not normally present at outcrop, may be a pale blue colour.

Some sediments, especially mudrocks, marls and fine-grained limestones, may be *mottled*, with subtle variations in grey, green, brown, yellow, pink or red colours. This may be due to bioturbation and the differential colouring of burrows and non-bioturbated sediment (*burrow mottling*), generating an *ichnofabric* (see Section 5.6.1), or it may be due to pedogenic processes: water moving through a soil causing an irregular distribution of iron oxides-hydroxides and/or carbonate, and/or the effect of roots and rhizoturbation (see Section 5.5.6.2). The term *marmorisation* has been applied to this process. Colour mottling

Table 4.6 *The colour of sedimentary rocks and probable cause.*

Colour	Probable cause
Red	hematite
Yellow/brown	hydrated iron oxide/hydroxide
Green	glauconite, chlorite
Grey	some organic matter
Black	much organic matter
Mottled	partly leached
White/no colour	leached

is common in lacustrine and floodplain muds and marls (see, e.g., Figure 5.3), especially sediments of palustrine facies (lake sediments strongly affected by pedogenesis).

Many sedimentary rocks show curious colour patterns (see Figure 4.14) that are similar to those produced in chromatography or loosely referred to as *Liesegang rings*: swirling, curved and cross-cutting patterns that are oblique to the bedding. The colours are usually shades of yellow and brown, even red, from variations in the contents of iron oxides and hydroxides. These may form at any time after deposition, although often related to weathering, and relate to the passage and diffusion of porewater through the sediment and precipitation or dissolution of minerals.

The common colours of sedimentary rocks and their cause are shown in Table 4.6.

Figure 4.14 *Iron-rich/iron-poor patterns (liesegang rings) in a fluvial sandstone resulting from pauses in the movement of groundwater through the sediment. The patterns usually have nothing to do with deposition or primary sedimentary structures. Carboniferous, Durham Castle, NE England.*