Chapter 1

THE NATURE, DESCRIPTION, AND CLASSIFICATION OF SEDIMENTS

1. SIZE

1.1 Introduction

1.1.1 If all sedimentary particles had the shape of regular geometrical solids, like a sphere or a triaxial ellipsoid or a square, both the concept and the measurement of grain size would be straightforward. But *sedimentary particles are almost always irregularly shaped*.

1.1.2 Some sand grains are almost perfect spheres, and many well rounded pebbles can be approximated well by triaxial ellipsoids, which are described by the lengths of the longest (*a*), shortest (*c*), and intermediate (*b*) major axes (Figure 1-1). But most sediment particles are so irregular that at best they can be represented only approximately in this way. The problem is worst with very fine particles. We thus run immediately into the difficulty of *what we mean by the size of a representative sedimentary particle*.

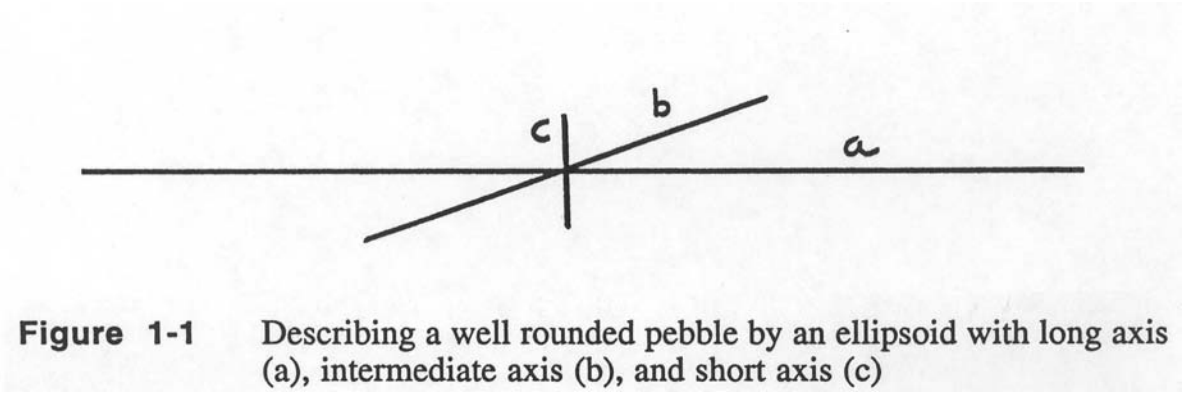

Figure 1-1 Describing a well rounded pebble by an ellipsoid with long axis (a), intermediate axis (b), and short axis (c)

1.1.3 In theory, a good way out of this problem is to characterize the size of the particle by its ***nominal diameter***, *the size of the sphere with volume equal to that of the given particle*. This is a nice concept, but it's hard to put into practice. (How would you measure the nominal diameter of a given sedimentary particle?)

1.1.4 So there's *no single unique dimension we can measure readily that would suffice to characterize the size of the particle*. We get around this in two ways, principally:

1.7.2 In many river systems it's observed that *sediment size decreases progressively downstream*. (This is called ***downstream fining***.) Why? Possibilities:

• *permanent storage* (sometimes certainly the case, sometimes certainly not the case), together with *size-selective deposition*.. Laboratory experiments have shown that this can be a substantial effect, at least under certain conditions. It makes sense, too, from the standpoint that the coarser particles are likely to be the ones that are preferentially extracted from the through-going sediment to be deposited.

• All particles are transported start to finish, all are progressively abraded. This certainly happens to some extent in gravel-size material, but there must be other factors involved too; it's unlikely that a boulder is continuously reduced to a sand grain. Probably much more important is:

• *fracturing* of coarser particles to form lots of fine particles. Also probably important in some if not most cases:

• *weathering* of sedimentary material as it resides in temporary alluvial deposits awaiting further downstream transport (but this probably doesn't affect quartz debris very much). Also important in many cases:

• *dilution* by fine material from tributaries.

1.7.3 A case could be made for nature providing us with three dominant sediment sizes: gravel, sand and coarse silt, and clay, resulting respectively from breaking along joints and bedding planes, granular disintegration and abrasion, and chemical decomposition. Many have proposed that there are indeed two distinct gaps or dearths in natural size distributions, *around 2 mm and at several microns*.

1.7.4 Mean size is seldom in the range 1–4 mm, and the feeling is that there's an absolute volumetric deficiency of this size (but this isn't easy to prove). Also, some have claimed to recognize a break around the medium to coarse silt size. What might be the reasons for the suspected gaps?

- not produced by rock disintegration or decomposition;
- less stable chemically or mechanically in weathering and transportation.

It's also possible that hydraulic effects cause some sizes to be scarce as modal sizes, although volumetrically no less abundant overall.

1.7.5 In summary, I think it's fair to say that although sediment size must have great potential for interpreting the origin and deposition of sediments, it has not yet become useful, because we don't yet fully understand its controls.

2. SHAPE

2.1 Introduction

2.1.1 The other important textural aspect of individual grains is *shape*. Sedimentologists have for a long time made a point of drawing a fundamental distinction between two different aspects of shape:

shape (*sensu stricto*): the *whole-particle* aspect of shape; this is usually viewed with reference to a sphere, so it's termed ***sphericity***.

roundness: the *local* aspect of shape, involving the sharpness, or lack thereof, of edges and corners.

2.2 Sphericity

2.2.1 It would be nice to have a unique and well-defined shape parameter that can be assigned to any sediment particle. One good way of doing this is to work from some reference shape. A sphere is often considered to be the best such reference shape, because it's claimed that *the limiting shape assumed by homogeneous and isotropic rock and mineral grains upon prolonged abrasion is a sphere*. (But I'm not really sure this is true.)

2.2.2 ***Sphericity*** is defined as *the ratio of the nominal diameter of the particle to the diameter of the circumscribed sphere*. A sphere has a sphericity of one, and particles of all other shapes have sphericities less than one.

2.2.3 You can imagine, however, the difficulties of measuring the sphericity of a large number of sedimentary particles according to this definition—especially tiny sand grains!

2.2.4 Another trouble with sphericity is that it doesn't distinguish among different nonspherical shapes. A more specific thing to do is compare the shape of the particle to that of a triaxial ellipsoid. This might still seem very artificial to you, but lots of fairly well rounded sediment particles don't deviate greatly from triaxial ellipsoids.

2.2.5 Figure 1-17 shows the standard and time-honored way of classifying particle shapes with reference to triaxial ellipsoids (or their equivalents in terms of rectangular solids). It was first proposed by Zingg, so the graph is called a ***Zingg diagram***. It's based on the three ratios of the principal axes of the triaxial ellipsoid:

L (or *a*): Long axis
I (or *b*): Intermediate axis
S (or *c*): Short axis

angular: little evidence of wear; edges and corners sharp

subangular: some evidence of wear; edges and corners rounded to some extent; original faces only slightly modified

subrounded: considerable wear; edges and corners rounded; area or original faces reduced

rounded: original edges and corners smoothed to broad curves; original faces almost gone; original shape still apparent

well rounded: no original edges, corners or faces; original shape only suggested by present shape

But nobody actually applies these descriptions directly. What one does is use an outline comparison chart. They even sell little plastic cards, called *grain-shape comparators*, which you can carry with you in the field to use when you're examining grains in a hand specimen with a hand lens, or when you're looking at a loose sediment with a hand lens or through a microscope.

2.3.5 What might shape and roundness of a sedimentary particle depend on?

- *Original shape and roundness* upon production (crystallization; fracture)
- *Inherent nature of material* (crystal habit; mineral/rock cleavage; abrasional anisotropy; fracture anisotropy)
- *Nature of weathering*
- *Transport processes*: distance, mode, medium (abrasion; corrosion; breakage)

2.3.6 And don't forget about the possibility of *shape-selective deposition*, mentioned earlier in connection with downstream fining. That would affect the aggregate shape and roundness properties of transported aggregates of grains, even though the shapes of individual sediment particles didn't change during transport.

2.3.7 The shape and roundness of clasts (mainly sand and gravel particles) have always been thought to hold great potential value for interpreting the *source, transportation, and deposition* of sediments. But there aren't many definitely established facts. One would like to know the answers to such important questions as:

- Do beach pebbles differ in shape from river pebbles?
- Does wind round grains more effectively than water?
- What is the lower size limit (if any) for rounding by wind or water?
- Can quartz grains be rounded in one sedimentary cycle (and if so, how)?

Despite lots of effort, so far there are only mixed results on the answers. There's a large literature on these matters, but I won't summarize it here. Suffice it to say that shape studies have so far not been any more useful than size studies in interpreting sediments.

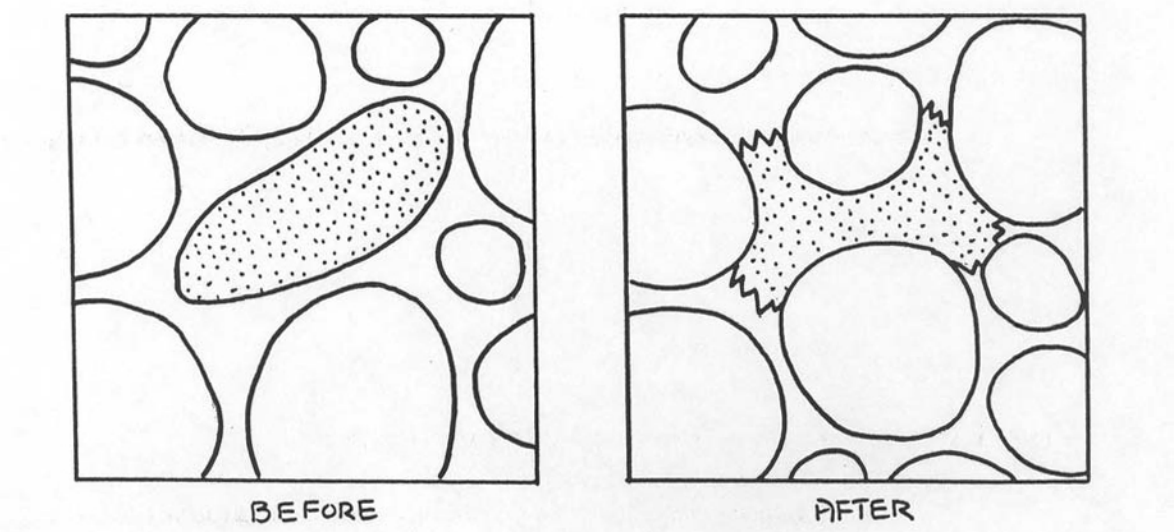

Figure 1-21 Squashing of relatively soft grains into unrecognizability during burial

pressure suturing (growing together of grains by solution at contact points and reprecipitation nearby; see the chapter on sedimentary minerals)

recrystallization, a word that gets used for two different processes, both important: (i) reorganization of crystal geometry of a given mineral (think in terms of what happens to snow as it sits in a snowbank at temperature not far below freezing); (ii) growth of crystals of certain newly stable minerals at the expense of crystals of the original detrital minerals which now find themselves out of equilibrium

3.7 All these things, both physical and chemical, that happen to a sediment during and after lithification, but before what's considered to be metamorphic, when the original sedimentary nature of the rock is largely obscured, is called ***diagenesis***. It covers a great variety of processes that operate from "early" to "late". There will be more on diagenesis later in the course.

3.8 The practical outcome of all this high-sounding discussion should be: *how do you look at the geometry of a sediment or a rock by hand lens or in thin section?* Figure 1-22 shows the standard way of viewing the granular organization of a detrital sediment or rock. The ***framework grains*** are *the larger detrital grains, which are in contact with one another*, leaving ***voids*** among them. The voids are filled with various kinds of ***void filler***. The classic distinction is between *cement* and *matrix*. The term ***matrix*** is used for *fine detritus that is deposited in the interstices among the framework grains*. Figure 1-22A shows a sediment with framework grains surrounded by precipitated mineral cement; Figure 1-22B shows a sediment with framework grains set in a matrix of fine detrital sediment deposited at the same time as the framework grains.

3.9 But here are four problems with this approach, in order of increasing severity:

- *Cement and matrix can be present together in the same sediment*.

- *If the void-filling material is fine-grained you often have a hard time telling matrix from cement*. You have to look at some thin sections to appreciate this.

- *The distinction between framework and matrix becomes unrealistic when the sediment is poorly sorted* (Figure 1-22C), especially if it's unimodal rather than bimodal. You just have to live with that. The distinction between framework and matrix then becomes just conventional or arbitrary.

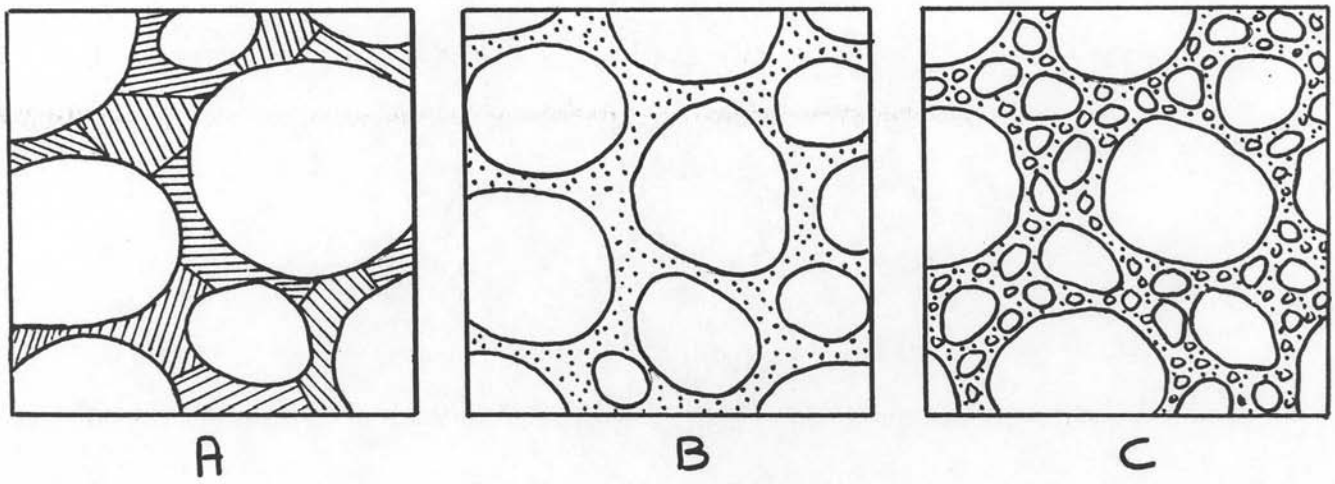

Figure 1-22 The granular organization of a clastic sediment. A: Framework grains are surrounded by precipitated mineral cement. B: Framework grains are set in a matrix of fine detrital sediment. C: In poorly sorted sediment, the distinction between framework and matrix is less natural.

• *Soft rock fragments can undergo a combination of deformation and recrystallization to the point where their material surrounds the more durable framework grains as a uniform "paste" that looks just like matrix.* There's been a lot of controversy over the years about the importance of this effect, but it's generally agreed to be important. You can see a whole spectrum of intermediate cases. So what you call matrix in a well diagenetized rock may or may not have anything to do with recrystallized original fine detritus.

4. SEDIMENTARY MINERALS

4.1 Introduction

4.1.1 What are the most important sedimentary minerals (or, more accurately, *sedimentary components*, because polycrystalline aggregates of one kind or another, rock fragments being the most important, are also common constituents of sediments and sedimentary rocks)? It's easy to list the common components:

quartz
feldspars
rock fragments
calcite + aragonite
dolomite
clay minerals
micas
heavy minerals

(Not necessarily in order of importance.)

4.1.2 Here I'll give some attention to the *terrigenous* or *siliciclastic* minerals of this list; the *carbonate* minerals will come later, in the section on carbonate sediments and rocks, and clay minerals in dealing with muds and shales.

4.1.3 Some miscellaneous general points about sedimentary minerals:

• *Any mineral can be present* in a siliciclastic sedimentary rock, but *not many are important*.

• There are *fewer important terrigenous sedimentary minerals* than there are important igneous and metamorphic minerals: for terrigenous rocks, weathering tends to get rid of them; the same is not true, however, for chemically precipitated rocks, because the range of physicochemical conditions in evaporating environments is very wide.

• Because of the very wide range of temperature and pressure conditions under which metamorphic rocks can be generated, and the restrictive regularities of crystallization in cooling silicate melts, *most minerals in terrigenous sedimentary rocks are metamorphic* (but this is not saying anything about volumes; igneous minerals would certainly be more abundant).

• For terrigenous rocks, *the same minerals* don't *have the same significance in a sedimentary rock as in igneous and metamorphic rocks*. The minerals in this kind of sedimentary rock are *not necessarily in equilibrium with each other or with their environment*. Each mineral grain has its own history, and each is an independent variable in the total characterization of the rock.

• A further complication, which you will inevitably run into when you look at clastic sedimentary rocks in thin section, is that *the same mineral can be present in a rock for both equilibrium and disequilibrium reasons*, i.e., the same mineral can be present in a rock in both detrital and authigenic form. (How do you figure this out?)

4.2 Quartz

4.2.1 *Quartz is probably the single most important sedimentary mineral.* It constitutes something like 25% of all sediments in the continental crust. The figure is much higher for most coarse-grained rocks, but very low in fine-grained

rocks and carbonate rocks. Here we're dealing with the relative proportions of (nonweatherable) quartz vs. (weathering-product) clay minerals from the source rocks; we shouldn't expect the proportions to differ too much from those in the source rocks, in terms of quartz versus clay minerals. *As a group, the clay minerals are more abundant than quartz.* The clay minerals are a group that's fairly homogeneous in form, occurrence, and general mineralogy, but no single mineral is dominant; more on clay minerals later.

4.2.2 Two factors contribute to the predominance of quartz as a sedimentary mineral:

• It's *moderately abundant in crystalline source rocks*; it forms perhaps 10–20% of the average igneous or metamorphic rock.

• It's *virtually indestructible by weathering (at least mechanical) and transportation*, whereas all other important source minerals are much more easily destroyed by weathering and/or transportation (mainly the former). All other highly stable igneous and metamorphic minerals, like zircon, are negligible volumetrically in source rocks, so even though they pass right through the sedimentary cycle unscathed, we don't see much of them.

4.2.3 There are five naturally occurring silica minerals:

stishovite: ultra-high T and P form, dense (4.3 g/cm^3), 1200° C, 10^5 atm. Present in meteor craters.

coesite: a high-P, high-T form of SiO_2, several 100° C, a few 1000 atm P. Producible in the laboratory, and formed under conditions like meteorite impact.

tridymite, **cristobalite**: high-T forms of SiO_2 found in various volcanic rocks (even abundantly, in some cases), but of negligible sedimentological importance (because of low resistance to chemical weathering).

quartz: two forms, *high quartz* and *low quartz*, related by a nonreplacive ("flip-over") inversion at 573°C. Whether the quartz originally crystallized as high or low, all quartz in sedimentary rocks is of course now low quartz. There's a slight change in *c*/*a* axial ratio in the high-low inversion, about 0.3%; this promotes fracturing. The two forms have a different crystal habit, but this is sedimentologically unimportant.

4.2.4 Miscellaneous mineralogical points about quartz:

• *Quartz is virtually a pure substance*; there isn't even limited isomorphous substitution or solid solution. All quartz contains trace elements, however, and this is beginning to be used in provenance studies.

• *Quartz is stable at surface T and P, and under all T and P conditions of sedimentary processes, as well as under almost all igneous and metamorphic conditions.* The reason we don't see quartz in things like basalts is compositional, not T and P stability. Quartz is *almost universally stable*; it's not around just

around 1–4 mm. Because quartz is the predominant clastic component in those sizes, this shortage must largely be tied up with *the overall size distribution of sedimentary quartz grains*. The following line of reasoning seems to be on firm ground:

• *The size of quartz grains in sediments is determined largely by the size of the quartz grains in the source rock*s, as modified by breakage, wear, and solution. (I'm unsure in detail about the relative or absolute importance of each of these.)

• *The quartz-bearing plutonic rocks (mainly granitic igneous rocks, gneisses, and schists) are the ultimate source of almost all detrital quartz.* But what is meant by *ultimate*? Most of the quartz from metamorphic rocks started out as sedimentary quartz. But the distinction between sedimentary and metamorphic quartz is significant for provenance studies.

• *Granitic source rocks produce only a small proportion of grains larger than 1 mm*: although perhaps one-quarter of quartz grains in the average granitic rock are larger than 1 mm, it's been estimated that so many of these are fractured that only about 10% of quartz grains from granitic rocks are larger than 1 mm (and only about 20% larger than 0.6 mm).

4.2.9 But how about *silt-size quartz in the sedimentary cycle*? It could be either *source-dependent* (from breakdown of fine-grained quartz-bearing crystalline rocks, phyllites being the only obvious possibility), or *process-dependent* (formed from coarser quartz particles by some process or processes). Here are some possibilities for processes of silt-size quartz production:

- *Spalling* of silt-size quartz grains from sand-size quartz grains upon impact during wind transport;
- *Glacial grinding and crushing* of quartz sand;
- Production somehow in other environments?

4.2.10 Because quartz is such a stable and abundant sedimentary mineral (to the extent that in many sedimentary rocks it is the only detrital mineral present in appreciable proportions), much effort has been expended in trying to *develop and apply criteria based on features of detrital quartz grains for determining the nature of the source rocks*. Here I'll discuss, briefly, four major features of this kind: *shape*, *inclusions*, *strain shadows*, and *polycrystallinity*.

4.2.11 Quartz grains obviously vary a lot in shape, but they tend to be *equigranular* or *spheroidal* (which term seems better sort of depends on how well rounded the grain is). They always show more or less *elongation*, however, and this tends to be along the sixfold crystallographic axis of the quartz grain. Why? Two possibilities:

- anisotropic abrasion properties caused by crystallographic anisotropy;
- quartz grains of source rocks crystallized elongated in this way.

The latter possibility seems to be the case; quartz seems to be *virtually isotropic with respect to abrasion* even though it is anisotropic with respect to crystal structure.

4.2.12 Quartz forms *anhedral grains* in most crystalline rocks. And quartz has absolutely *no mineral cleavage*. So almost all quartz grains are very irregular in shape and roundness from the beginning (neither highly angular nor highly inequigranular). When quartz crystallizes freely, it forms doubly terminated pyramids—but such crystals are extremely uncommon in sedimentary rocks; *you rarely see a euhedral detrital quartz grain*. Crushed quartz looks different; but we almost never have to deal with crushed quartz in natural situations.

4.2.13 There's conflicting evidence on whether quartz grains from schists and gneisses tend to be more elongated than those from coarse-grained igneous rocks. For a long time this was assumed to be so, without any careful and extensive and systematic measurements. But more recent studies have shown that this is not strongly the case, if at all. It seems to be generally thought now that *nothing much can be said about either provenance or transport history by studying the shapes of quartz grains* in a detrital sediment.

4.3 Feldspars

4.3.1 *Feldspar is the second-most-important coarse detrital mineral in sedimentary rocks.* It's subordinate to quartz in sedimentary rocks, and on the average it's *sharply* subordinate: the average sandstone contains something like 10-12% feldspar (mainly detrital). (Fine-grained terrigenous rocks contain at least this much on the average, but it's hard to tell how much is detrital and how much is authigenic; probably more authigenic than detrital.) Another important fact is that *feldspar content is much more variable than quartz content*: some detrital rocks contain almost no feldspar, whereas others contain more feldspar than quartz.

4.3.2 Although it's hard to sample representatively, *there's a real increase in the average feldspar content of terrigenous rocks from the early Paleozoic to the present*: Pre-Pennsylvanian, about 2%, and Cenozoic, about 25%. WHY? There are only four logical possibilities for why there's less feldspar in older sandstones:

- *source*: less got in, because typical source rocks had less feldspar. [seems unlikely]
- *weathering*: less got in, because typical relief and/or climate was more conducive to elimination of feldspar. [seems unlikely]
- *preservation*: the same amount got in, but rocks with more feldspar have been preferentially destroyed. [seems reasonable]
- *dissolution*: the same amount got in, but the rocks have had more chance for feldspar to be dissolved out or altered by pore waters. [seems reasonable]

4.3.3 With regard to composition, every known variety of feldspar in crystalline rocks is found also in sedimentary rocks, but:

- certain kinds of dominant, and others are rare;
- the proportions of the various feldspars in sedimentary rocks are markedly different from in crystalline source rocks.

Plagioclase is an important constituent of crystalline source rocks, but it's *not abundant* in sedimentary rocks; it's strongly subordinate to Kspars, which are the dominant feldspars in sedimentary rocks. Also, *the more sodic plagioclases are more common than the more calcic plagioclases*. Of the Kspars, *microcline is somewhat more common than orthoclase, but both are very common*. Sanidine, on the other hand, is rare.

4.3.4 What's the *provenance* of feldspar in sedimentary rocks? The Kspar comes mainly from coarse-grained felsic (granitic) igneous rocks and gneisses. Plagioclase comes mainly from intermediate to mafic igneous rocks, as well as plagioclase-bearing metamorphic rocks. Sedimentary and volcanic sources seem to be unimportant.

4.3.5 *The percentage of detrital feldspar in sedimentary rocks is significantly less than in the source rocks*—something like one-fifth as much. This is certainly caused at least in part by chemical weathering of feldspar at the source. It's observed that feldspars are unstable during weathering in soils, and that the various feldspars vary in their resistance to weathering, and that these differing resistances correspond to the differing abundances of feldspar in sedimentary rocks. You can further feel confident that if you find a sedimentary rock with a lot of potentially unstable feldspar, like calcic plagioclase, there must have been *some special conditions such that weathering did not have enough chance to operate* (either not enough time or not enough intensity, or both).

4.3.6 What about the possibility that the feldspar is eliminated also in part by mechanical abrasion during transport (or this plus chemical weathering while in temporary storage)? Careful abrasion experiments have shown that *abrasion of feldspar is twice as fast as of quartz but still insignificant in subaqueous transport*. So it seems that the percentage of sand-size feldspar in a river can be reduced only by dilution or by chemical weathering along the way.

4.3.7 It was long believed that for feldspar to survive weathering and transportation and be deposited in substantial quantities in a sediment requires very special climatic conditions: very arid (not much water to facilitate chemical decomposition) and/or very cold (low temperatures retard chemical decomposition). But it gradually became clear that *highly feldspathic sediments accumulate in very wet and hot climates*. Another piece of evidence is the presence of tropical flora in feldspar-rich sandstones. And the mere presence of altered feldspar along with fresh feldspar in the same rock, which is very common, indicates something is wrong with "cold dry climate" theory.

4.3.8 Probably more important than climate is *the relief of the source area*. Climate controls the *intensity* of weathering, but relief controls the *time available* for weathering. Even in a climate with intense weathering, if the relief is very great the rocks are eroded very rapidly, so there's not enough time for weathering to destroy feldspar. But if relief is low, the rate of erosion is so slow that if the climate is favorable, all the feldspar is destroyed. *The presence or absence of feldspar is thus controlled by the balance between rate of weathering and rate of erosion*. Detrital feldspar is an index of both climate and relief. What causes high relief? Uplift. Thus, *feldspar in sediments should indicate tectonic activity in the source area*.

4.4 Rock Fragments

4.4.1 Petrologically, a ***rock fragment*** is *a detrital component or clast that is large enough relative to the grain size of the source rock so as to be petrologically representative of the source rock from which it was derived*. The size limitations on rock fragments should be obvious: a source rock can't yield rock fragments finer than the constituent mineral grains of the rock! Thus, coarse-grained plutonic rocks can produce gravel-size rock fragments but not sand-size rock fragments.

4.4.2 The nice thing about rock fragments is that they are a great way of *telling provenance*: it's like having little chunks of source-rock outcrops delivered to your door. The trouble is that the rock fragments that are delivered may not (and are usually almost certainly not) representative of all the rocks in the source area, because of differing resistance to weathering and/or scale of fragmentation. Also, for sandstones, the only rock fragments you will see are very fine-grained, and it's difficult to do careful petrographic work on such fine-grained rock fragments.

4.4.3 Rock fragments are *moderately abundant*: perhaps 10–15% of average sandstone. But percentages vary widely, from almost zero to almost all the framework. Here are the most important *kinds* of rock fragments of sand size:

- **volcanic igneous**. Usually so fine-grained that they are hard to identify. Usually felsic rocks, because they are more resistant to chemical weathering than mafic rocks. They are often hard to distinguish from impure chert. Mafic volcanic rock fragments are easier to identify but are less common in most sandstones.
- **metamorphic**. The whole range of foliated fragments from the series shale–phyllite–schist are common. Usually they show a high percentage of micaceous material with preferred orientation, and are therefore fairly easy to identify.

- **sedimentary**. Chert is probably the most common, and is usually easy to identify. Shale and mudstone clasts are also common, and grade over into metamorphic rock fragments.

4.4.4 Rock fragments are usually counterposed to feldspars in interpreting provenance of sedimentary rocks: both have about the same resistance to weathering (to first order), and both are abundant in source rocks (potentially produced in large quantities). Feldspar tends to come from deep-seated rocks, rock fragments from shallow-seated rocks. So *the ratio of feldspar to rock fragments is usually considered to be a good first-order guide to source-rock composition*. This measure is far from perfect, however, because

- the proportions of feldspar and fine-grained rocks vary in the source areas, and
- the weatherabilities of feldspars and rock fragments are not identical.

4.4.5 Also, *the ratio of quartz to the aggregate of feldspar and rock fragments is a good guide to what could be called the maturity of a sedimentary rock*. The ***maturity*** of a detrital sedimentary rock is *the extent to which the sediment approaches the end product to which it is driven by the sedimentary processes* (like weathering and transport) that operate on it. The ultimate end product is an aggregate of well-sorted and well-rounded quartz grains plus a small percentage of highly stable heavy minerals like zircon and tourmaline. This measure isn't perfect either, however, because the initial proportions of quartz and other constituents vary greatly in the source rocks.

4.5 Micas

4.5.1 With regard to mica particles in detrital sediments, it's nice to think in terms of

- large detrital mica flakes
- fine interstitial clay-mineral particles

But often there's no break in the size distribution of the micas, or even any bimodality in the distribution, so we can't make this distinction. Also, there's lots of authigenic clay-mineral material in many sedimentary rocks, and *you may not be able to distinguish detrital from authigenic*. (Some of the authigenic is derived from the detrital, and some is derived from alteration of feldspars.) I'll concentrate here on the coarse detrital micas, and deal with the fine stuff in the chapter on muds and shales.

4.5.2 Muscovite, biotite, and chlorite are all common, but muscovite is far more abundant that biotite or detrital chlorite.

muscovite: the most stable, both chemically and mechanically.

biotite: undergoes chemical weathering fairly readily, largely to fine-grained chlorite.

chlorite: tends to break down into fine particles and be present as a "clay mineral" constituent. It's also very common as an incipient metamorphic mineral in shales and slates.

4.5.3 In most sandstones, muscovite is a minor constituent, a few percent, but it is *ubiquitous*. It commonly gets concentrated on bedding planes, by settling horizontally onto the sediment surface at times of weaker currents. This tends to facilitate parting at that bedding plane, because the excellent cleavage in the mica makes that bedding plane a little weaker when the sediment is lithified. Much of the parting (that is, splitting along bedding surfraces) seen in detrital sedimentary rocks is caused just by greater concentrations of muscovite.

4.5.4 Muscovite characteristically has a much lower settling velocity than chunky detrital grains of the same mass. So *you tend to find muscovite together with much finer detritus*. The hydrodynamic reason, in a nutshell, is this: flaky particles tend to settle with their planes horizontal, and the fluid drag force is about proportional to the "frontal area" of the settling particle. Figure 1-23 shows two particles of the same volume. The flake has the larger frontal area, and so the necessary balance between the weight and the drag force is struck at a smaller settling velocity.

4.6 Heavy Minerals

4.6.1 The ***heavy minerals*** (often just called heavies) of detrital sedimentary rocks are *a large and nongenetic mineral group*. The minerals are not necessarily related to each other in any way: it's the operational procedure in separating them from the sediment that defines them as group. By definition, the density of heavy minerals is greater than that of some heavy organic liquid, about 2.9 g/cm^3 (usually bromoform or tetrabromoethane). It's a mineralogical fact of nature that the major sedimentary minerals (quartz, feldspars, carbonates) are around 2.65 g/cm^3 in density, and the minor minerals are mostly much heavier.

4.6.2 Operationally, the heavies are usually treated in two different groups: *opaque* ("ore") minerals and *nonopaque* minerals. These two groups have to be studied by different microscopic techniques. The nonopaque minerals can be studied by transmitted light, just like the light minerals, whereas the opaque minerals are studied like ores, in reflected light. The opaque minerals therefore tend to be ignored by sedimentary petrologists.

4.6.3 Unless you are dealing with a placer deposit (a ***placer*** is a *sedimentary accumulation of heavy-mineral grains rich enough to be economically workable*) heavies almost always form less than 1% of the sediment, and usually less than 0.1%. For this reason, you don't often even see heavy-mineral grains in thin section! But an important point to remember is that *the heavy-mineral fraction of*

a detrital rock contains almost all the minerals present in that rock. If only for this reason alone, they are of more than casual interest.

4.6.4 The heavy minerals, opaque and nonopaque together, fall into two broad categories:

• *Very stable minor accessory minerals of the source rocks.* These aren't abundant in the source rock, but they are so resistant to weathering and abrasion that they get into the sediment anyway. As with quartz, it's safe to assume for these minerals that as much gets into the sediment as leaves the source area. Here's a list of some of the common stable heavy minerals:

zircon
tourmaline
rutile
anatase
brookite
garnet
magnetite
ilmenite
apatite
titanite

• *Surviving remnants of abundant but unstable femags in the source rocks.* To the extent that these escape weathering, they get into the sediment, so depending on the particular conditions they can either be very abundant or (more likely) rare or entirely absent. Examples: olivine, pyroxenes, amphiboles, staurolite, kyanite, sillimanite.

Some minerals, like garnet, fall in between these two categories.

4.6.5 Here are two interpretive uses of heavy minerals:

• *delineation of petrographic provinces*: in coastal marine geology, it's common to study the heavy-mineral content of detrital sediments to delineate the different continental source areas of the sediments. Theoretically this could be done for ancient rocks as well, but the difficulties are increased because of loss of three-dimensional control.

• *identification of type of crystalline source rocks*: It would be nice to use heavy minerals as a guide to the provenance of a sedimentary rock: what were the source rocks? To some extent we can do a fairly good job in using the heavy minerals to tell what the source rocks were, at least ultimately. Minerals like zircon, sphene, monazite, and pink tourmaline are characteristic of granitic rocks; garnet, kyanite, sillimanite, staurolite, and epidote, of medium to high grade metamorphics. But obviously the great variety of source rocks and mixing from a number of different sources often makes this kind of approach difficult.

4.6.6 Here are two significant things about heavy minerals:

• *The heavies tend to be in the finest fractions of the detritus*. Their greater density means that their settling velocity is greater for a given mass, so finer heavies tend to be deposited with coarser lights. (This is a considerable oversimplification of what's called *hydraulic equivalence* of light and heavy minerals; you have to worry about traction as well as settling. I can't pursue this here.)

• In a succession of interbedded sandstones and shales, *the number of heavies is much greater in the shales than in the sandstones*; the greater permeability of sandstones has allowed dissolution or transformation of the heavies, whereas the impermeable shales have retained their original heavy-mineral content. So *if you are looking for heavies, look in fine-grained rocks*.

5. CLASSIFICATION OF DETRITAL SEDIMENTS

5.1 *The fundamental classification of detrital sediments is by particle size*. That said, however, I think I should point out that such a classification tends to be treated rather informally by most sedimentologists. The major size categories are ***gravel***, ***sand***, and ***mud***, the first being subdivided into ***pebbles***, ***cobbles***, and ***boulders***, and the last being subdivided into ***silt*** and ***clay***. The boundaries are as given by the Udden-Wentworth grade scale, which is now conventional in the Earth sciences, at least in English-speaking countries. The Udden-Wentworth grade scale is based on powers of two; each successive power of two coarser than the preceding by one phi unit (see Section 1 above). Figure 1-24 shows this grade scale, and its conventional terminology. Especially in the sand sizes, the adjectives, ranging from very fine to very coarse, are heavily used in practice. And the finer subdivisions of sand are often split even further—for example into upper fine or lower medium—even when working with a hand lens and a grain-size comparator in the field.

5.2 The classification shown in Figure 1-24 would be very nice if sediments were always very well sorted. In real life, however, there is always a spread of sizes around the mean size. What does one do about naming a sediment when one or both of its size-distribution tails laps over into another size category?