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*.

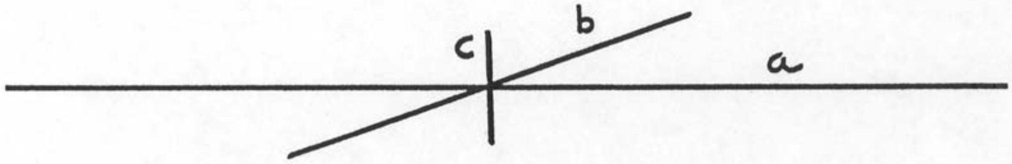

Figure 1-1 Describing a well rounded pebble by an ellipsoid with long axis (a), intermediate axis (b), and short axis (c)

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:

- **Let our choice of characteristic dimension be operationally defined**. That is, *how we measure size defines what we mean by size*. The standard way of measuring size in the coarser-grained rocks (coarser siltstones and upwards, above several tens of microns) is by ***sieving***. How does a sieve discriminate by size? It more or less *chooses the largest dimension in the least cross-sectional area*. This might not seem like a very satisfactory measure of size, but remember that most sediment grains coarser than clay size are equigranular, and so exactly how we discriminate is not too important as long as it is standardized.

- **Relate size to some other more easily measurable property.** The standard way of measuring fine-grained material is by ***settling***. (The size of sand-size material is often measured this way, too, in devices called settling tubes.) For small spheres there's an exact solution for settling rate. The difficulties of nonsphericity are even greater for clay-size sediment particles than for sand, but because there's no other way, we *assume* that they are spheres for the sake of definiteness.

1.1.5 All you need to know about fluid dynamics in order to use it for size measurement is this:

> **for a particle of a given shape exposed to a steady and uniform flow of fluid around it, the drag force exerted by the fluid depends on the velocity of flow, the size of the sphere, and the fluid properties density and viscosity.**

1.1.6 When you release a particle in a body of still water, the particle accelerates in response to the downward force of its weight, and the drag force builds up until it balances the weight; then the particle settles at its terminal settling velocity (Figure 1-2).

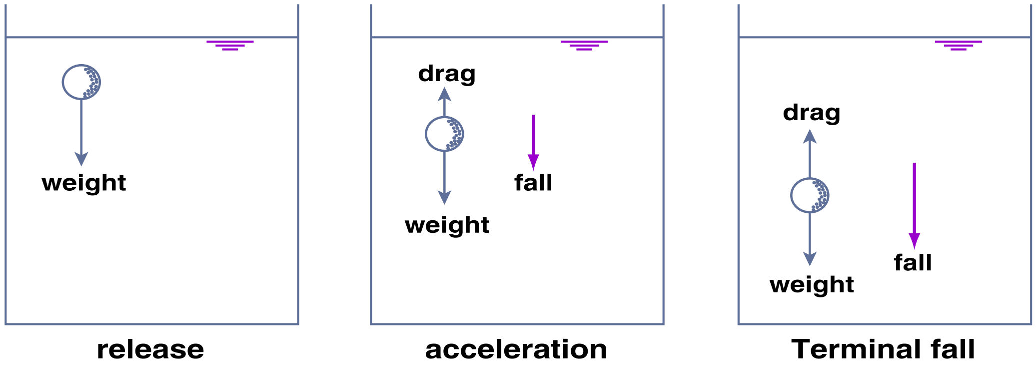

Figure by MIT OCW.

Figure 1-2: Balance between downward force of gravity and upward force of fluid drag on a sediment particle falling at its terminal fall velocity

1.1.7 For a given fluid, like water, this means that *there is a function that relates size to velocity*. Once the function is known, all you have to do is measure the settling velocity, go to a curve or table, and find the size. There are exact solutions for this function only for small particles with a few regular shapes, like spheres or triaxial ellipsoids, but *all you really need to know is the empirical function*. Figure 1-3 shows a cartoon of this function; it's only qualitatively correct. For a real version, see the next chapter.

Figure by MIT OCW.

Figure 1-3: Cartoon version of graph of terminal fall velocity against sphere size

1.1.8 The trouble with this approach is that particle shape varies all over the place. One way around this is to develop empirical functions for a wide range of shapes, but this isn't very workable because there are so many shapes. What's usually done is to work with the ***fall diameter*** or ***sedimentation diameter***: *the diameter of a quartz sphere with the same settling velocity as the given particle, in a standard fluid (water) at a standard temperature (20° C).*

1.1.9 So when it comes to measuring size, we have to live with an imperfect world. What's done in actual practice for the various size ranges is as follows:

gravel: usually sieving, or for special studies, direct measurement with calipers; the coarsest sizes usually aren't even measured.

sand: sieving is the standard method, but various kinds of settling tubes are also in common use.

mud: down to about 40–50 μm, sieving (wet sieving is better in the fine sizes), and for all mud sizes, settling techniques.

1.1.10 These techniques work only for unconsolidated sediment. *What do we do to measure grain size in rocks?* Typically, cut a thin section, measure sizes with a micrometer ocular, and then make a point count at the same time to get the distribution. The obvious difficulty with this is that the sectioning usually doesn't cut through the center of the particle. There have been studies, both theoretical and empirical, aimed at getting around that problem If the rock is not too indurated, gentle disaggregation and then sieving is often possible.

1.2 Grade Scales

1.2.1 It's natural to *split up the continuum of sizes into conventional divisions*, to facilitate both communication and thought. Such a split is called a ***grade scale***. Several grade scales are in use, depending upon field of study and where you're working. English-speaking sedimentologists use the Udden-Wentworth grade scale, built around powers of two. It's a geometric or logarithmic rather than an arithmetic scale. (This is desirable for largely obvious by partly inexplicable reasons.)

1.2.2 It's easiest to work directly with a linear or arithmetic measure of the log scale rather than the logs of the actual sizes. For this reason, the famous (infamous?) ***phi scale*** was introduced:

$$\phi = -\log_2(\text{diameter in mm}) \qquad (1.1)$$

Figure 1-4 gives a conversion table from millimeters to phi units.

millimeters	**4**	**2**	**1**	**$1/2$**	**$1/4$**	**$1/8$**	**$1/16$**	**$1/32$**
ϕ	**-2**	**-1**	**0**	**1**	**2**	**3**	**4**	**5**

Figure by MIT OCW.

Figure 1-4: A conversion table between millimeters and phi units

1.3 Statistical Analysis of Particle Size

1.3.1 Statistics is an important part of sedimentology. A great many variables sedimentologists measure and deal with are ***random variables***, in that *why they take a given value seems to be random, even though there may be underlying laws we can't sort out*. You have to deal with these variables statistically. Sediment size is only one example of such a random variable. What follows is a brief account of the statistics of random variables applied to sedimentology.

1.3.2 For a given sediment, why do we see the percentages we see in each of the class intervals or fractions? Usually we can't answer that, but we can at least develop a rational way of describing what we have.

1.3.3 The raw data, obtained from size measurement by sieving or settling analysis, is usually in the form of *weight percent of total sample in each of the various size classes defined by the grade scale*. These percentages add up to 100%. (If they don't quite add up to 100% when you make the measurements, just distribute the small error among all the fractions to make the total come out to 100%.)

1.3.4 Probably the most important size characteristic of a sediment is the *"average"* size. A natural way of expressing this is the *mean size*. (But there are other ways; see below.) The ***mean*** of a number of values is defined as

$$M = \frac{\Sigma x}{N} \tag{1.2}$$

where x is a value and N is the number of values. The mean of *grouped* data (the kind we get in sediment size analysis, by either sieving or settling) is

$$\mathrm{M} = \frac{\Sigma(fx')}{N} \tag{1.3}$$

where f is frequency in a class, x' is class midpoint, and N is number of values.

1.3.5 The most meaningful mean in our case is the ***logarithmic mean***. You get it by:

using the phi values of the midpoints of the class intervals,
multiplying these phi values by the frequencies in the size classes,
summing over all the values, and
dividing by the number of values.

Then you can get the mean size by converting from phi to mm.

1.3.6 One problem with the mean is that the *tails* of the size distribution, which are hard to measure accurately (fine stuff gets lost; big stuff is too "lumpy" statistically) have a strong effect on the computation.

1.3.7 You also need a measure of *how clustered or spread out the distribution is around the mean*. This idea is expressed by the word ***dispersion***. Dispersion expresses the ***sorting*** of the sediment: a sediment whose sizes tend to fall near the mean is said to be ***well sorted*** (or, in engineering terminology, poorly graded; engineering usage is opposite to sedimentological usage!), and a sediment whose sizes range widely around the mean is said to be ***poorly sorted***.

1.3.8 Here's the best way to find the true standard deviation σ:

$$\sigma = \sqrt{\frac{\sum (x - \bar{x})}{N}} \tag{1.4}$$

where x is the mean, or, for grouped data,

$$\sigma = \sqrt{\frac{\sum (x' - \bar{x})}{N}} \tag{1.5}$$

1.3.9 The ***standard deviation*** is *the square root of the* ***variance****, the sum of the squares of deviations of frequencies in each class interval from the mean, divided by the number of values*. Again this is most easily done using the phi scale. We have the same problem here as with the mean: the tails of the distribution are important.

1.3.10 In theory we could also compute the so-called *higher moments of the distribution*. (The mean and variance are the first two moments of the distribution.) Although I don't expect you to make use of it, here's the expression for pth moment of the distribution:

$$m_p = \frac{\sum (x - \bar{x})^p}{N} \tag{1.6}$$

1.3.11 What do these higher moments mean? It turns out that an expression involving the second and third moments describes *the degree of asymmetry of the distribution* (***skewness***), and an expression involving the second and fourth moments describes *the ratio between the spread in the middle part of the distribution and the spread in the tails* (***kurtosis***). Figure 1-5 shows them, just for your reference.

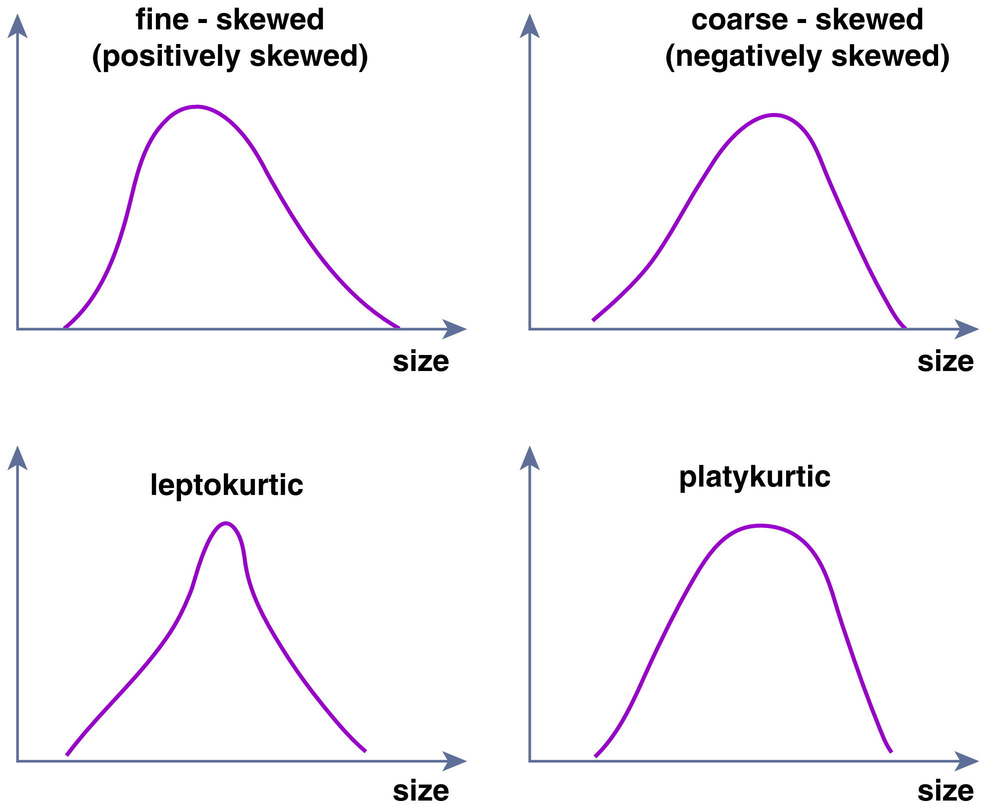

Figure by MIT OCW.

Figure 1-5: Terms to describe skewness and kurtosis in frequency distributions

1.3.12 Skewness seems to be sedimentologically significant, but the significance of kurtosis is unclear. With these higher-moment measures, errors in sampling and measurement get highly magnified. For meaningful and reproducible results, standardized and calibrated sieves at narrow size intervals have to be used.

1.4 Graphing Size Distributions

1.4.1 How do you express graphically the data of size analysis? The simplest way, but not the most meaningful, is to plot a ***histogram***, *a vertical bar graph of weight of sediment sample by size classes* (Figure 1-6). Obviously, the heights of the bars depends on the size of the sample you use.

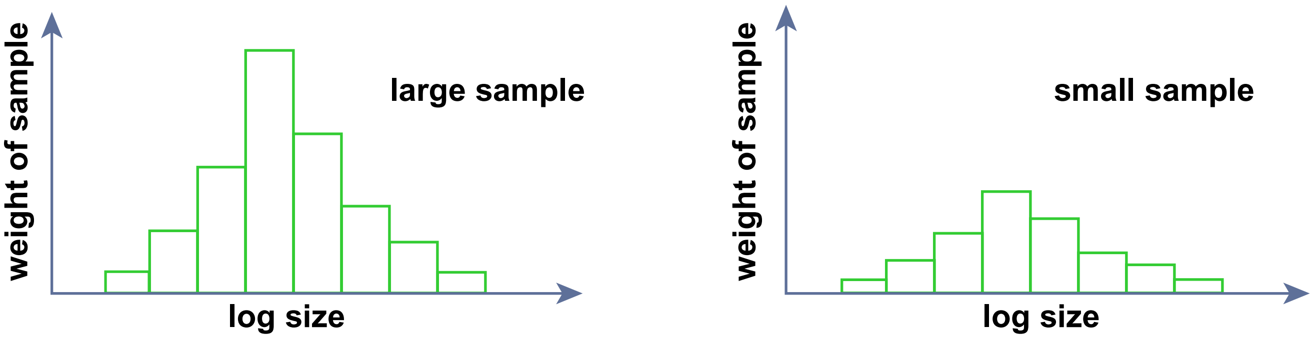

Figure by MIT OCW.

Figure 1-6: Histogram of weight of sample against particle size

1.4.2 Or, you could *normalize* the histogram by plotting *percent of total weight vs. size class interval* (Figure 1-7). Then, no matter how large or small the class interval, the total area under the histogram bars is unity (100%).

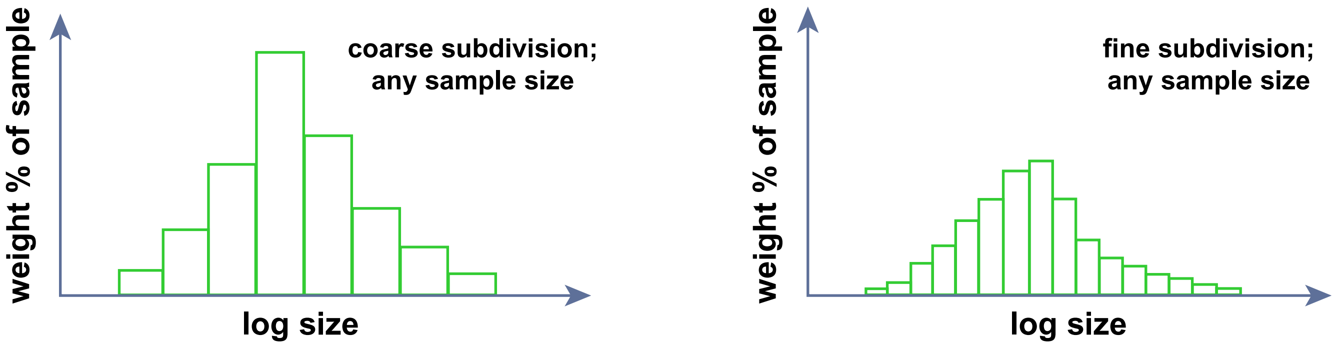

Figure by MIT OCW.

Figure 1-7: Histogram of weight percent of sample against particle size

1.4.3 But still the height of the bars varies with the fineness of division of the size scale. Better yet: *make the vertical axis be weight percent divided by size interval*. Then the overall course of the tops of the bars is the same regardless of the fineness of subdivision (Figure 1-8).

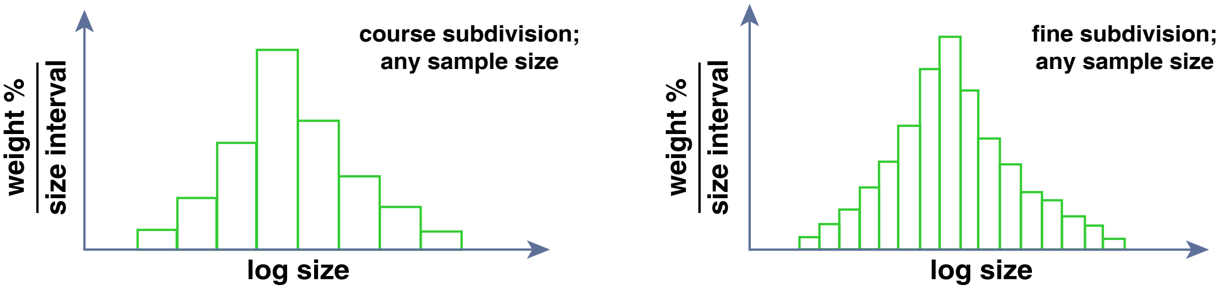

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Figure 1-8: Normalized histogram of weight percent of sample per unit size interval against particle size

1.4.4 As you progressively decrease the size class interval, the bars get thinner, but the overall shape and size of the histogram stays the same. This is best visualized in a vertical line graph, with data plotted as vertical lines from the midpoints of the class intervals (Figure 1-9). This kind of graph also lends itself better to thinking about a limiting process.

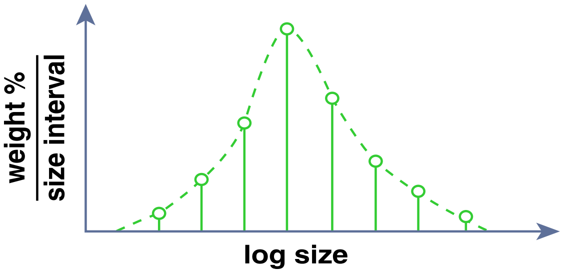


Figure by MIT OCW.

Figure 1-9: Vertical line graph of weight percent of sample per unit size interval against particle size (as in Figure 1-8) for better conceptualization of the limiting process that leads to a smooth frequency distribution function

1.4.5 This suggests the possibility of performing a limiting process to obtain a smooth curve from the histogram: *as the number of class intervals increases without limit, does the stepped curve connecting the tops of the vertical lines tend toward a smooth curve?* In principle, **NO**, because there are only a finite number of sand grains in a sample, and “most” sizes are missing. Past a certain point in the fineness of subdivision the histogram blows up and does funny things.

1.4.6 But this is being overly pedantic; you don’t need to worry about this operationally unless you have only a very small number of grains to work with. You have to assume that your dependent variable, weight of sample in a given size class, which is quasi-continuous on a large scale but definitely discrete on a small scale, is actually a continuous variable. If you performed the limiting process for a smoothly continuous variable, you would get *a limiting continuous curve toward* *which the successively finer histograms would tend.* This is called the ***frequency distribution curve*** or ***frequency distribution function*** (Figure 1-10).

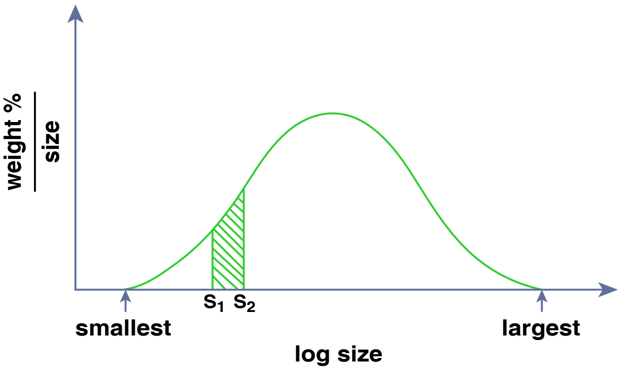


Figure by MIT OCW.

Figure 1-10: Frequency distribution function for sediment size, expressed as weight percent per unit size interval against sediment size

1.4.7 The physical meaning of the vertical coordinate is seen most clearly by considering two different sizes and saying that the percent of the total sample between these two sizes is gotten by integrating the frequency distribution function between the size limits chosen (i.e., find the area under the frequency curve between these two sizes).

$$\int_{s_1}^{s_2} f(\eta)d\eta = \text{weight \% of sediment between the two sizes } s_1 \text{ and } s_2 \qquad (1.7)$$

1.4.8 There's no easy way to find the frequency distribution directly. It might occur to you to plot the histogram and then fit a smooth curve through the tops of the histogram bars. But that provides only a crude estimate of the frequency distribution, unless you make the number of class intervals unworkably large. The way sedimentologists get around this problem is to work with a different kind of distribution, called the ***cumulative distribution***.

1.4.9 Both the concept and the practice of the cumulative distribution are fairly simple. Suppose that you have sieved a sample of sand and have weight percents for each size class. You can easily accumulate your data by computing a running cumulative total that gives you *weight percent finer than* (it could also be weight percent coarser than) for each size class boundary. Then plot a graph with weight percent finer than (arithmetically) on the vertical axis and size (logarithmically, or, what's the same, arithmetically in phi units) on the horizontal axis. You get a curve that starts at zero at the finest size in your sample and increases (monotonically!) to one hundred percent at the coarsest size in your sample (Figure 1-11). The curve is usually at least crudely S-shaped, as in the figure.

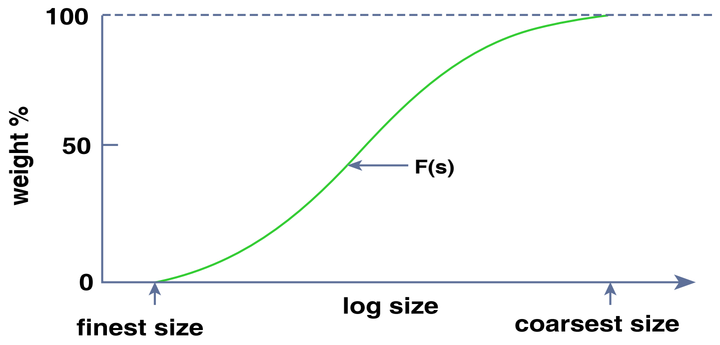


Figure by MIT OCW.

Figure 1-11: Cumulative distribution function showing weight percent of sample against particle size

1.4.10 The nice thing about the cumulative curve is that *it's easy to find just by plotting the cumulative data points and fitting a smooth curve*. By sieving your same sample using several different stacks of sieves, with different size class boundaries, you could demonstrate to yourself that the curve you get is almost insensitive to the particular choice of size intervals.

1.4.11 Following up on Equation 1.7, the cumulative distribution function can be written as the following definite integral:

$$F(s) = \int_{s_o}^{s} \mathrm{f}(\eta)\, d\eta \qquad (1.8)$$

The frequency curve and the cumulative curve are thus related as derivative and integral. Figure 1-12 shows the correspondence between the frequency curve and the cumulative curve. Both are useful and common in size analysis, as in many other applications of statistics.

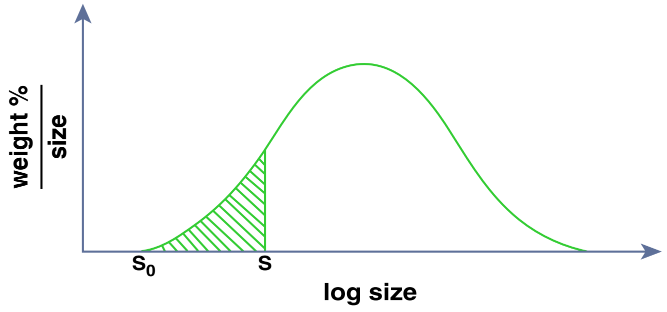


Figure by MIT OCW.

Figure 1-12: The frequency distribution curve for particle size, labeled to aid in understanding the correspondence between the frequency distribution and the cumulative distribution

1.4.11 The frequency curve is not as easy to obtain as the cumulative curve. You have to resort to things like graphical differentiation, or approximation of the cumulative curve by an analytical function (polynomial, power series, Fourier series, etc.) and then differentiation at various points. This used to be tedious before the advent of computers, but now it's easy. Various standard programs have been developed. This isn't something we'll pursue here.

1.5 More on the "Average"

1.5.1 In light of the frequency curve and cumulative curve, here's more detail the average (or the so-called "central tendency", in the parlance of statistics) of a size distribution. I've already defined the mean, but in addition to the mean, two other measures of the average properties of the distribution can easily be obtained from the cumulative and frequency curves.

Mean: On the frequency curve, the mean size is that for which the center of gravity of the area under the curve lies. On the cumulative curve, it's the point at which there are equal areas, above and below, out to the limits of the distribution.

Median: The size that cuts the distribution in half, 50% finer, 50% coarser. On the frequency curve, it's the point at which there are equal areas on each side; on the cumulative curve, it's the point at which the curve crosses the 50% figure.

Mode: The most abundant size in the distribution. On the frequency curve, this is the highest point. On the cumulative curve, it's the point at which the curve has the steepest slope.

1.5.2 It's hard to say which measure is the most significant. The median is the easiest to find but probably the least valuable.

1.6 Analytical Distributions

1.6.1 To what extent are actual size distributions approximated by analytical distributions? It is commonly assumed that sediment size distributions tend to be ***normally** (Gaussian) distributed*. Figure 1-13 shows the familiar bell curve of the normal distribution. And here's the equation for the normal distribution:

$$y = f(x) = \frac{1}{\sqrt{2\pi\sigma}} \exp\left[-\frac{1}{2}\left(\frac{x-\mu}{\sigma}\right)^2 \right] \tag{1.9}$$

where μ is the mean and σ is the standard deviation; they're called the *parameters* of the distribution.

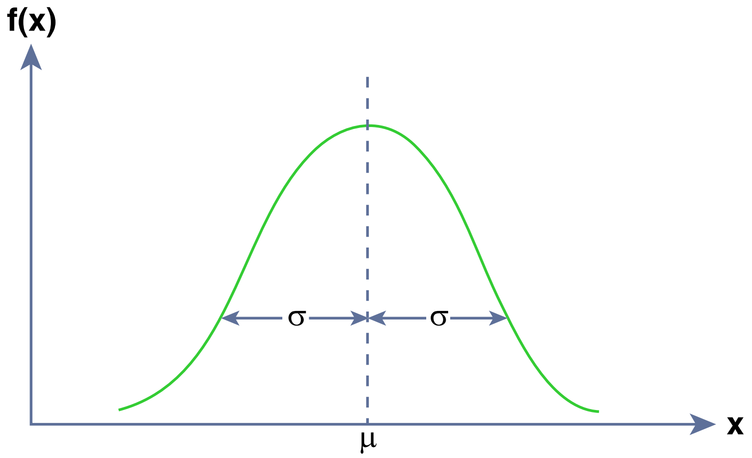


Figure by MIT OCW.

Figure 1-13: The Gaussian or normal frequency distribution

1.6.2 Properties of the normal distribution:

- Symmetrical about the mean;
- Distribution extends to +∞ and -∞, → 0 asymptotically;
- 68% of distribution lies between ± one standard deviation of the mean; this is at 16th and 84th percentiles. The inflection points in the frequency distribution are at these points.

1.6.3 A great many sedimentological variables tend to be normally distributed. (Or, more precisely, lognormally.) There's usually no evident reason why, and it's rarely exactly so. It seems to turn out that many if not most natural sands have a long straight-line segment on a graph like this, with tails that curve away.

1.6.4 Size distributions are commonly plotted on graph paper (called ***probability paper***) with the frequency axis rubber-sheeted so that a normal distribution plots as a straight line (Figure 1-14). This has become a reference standard.

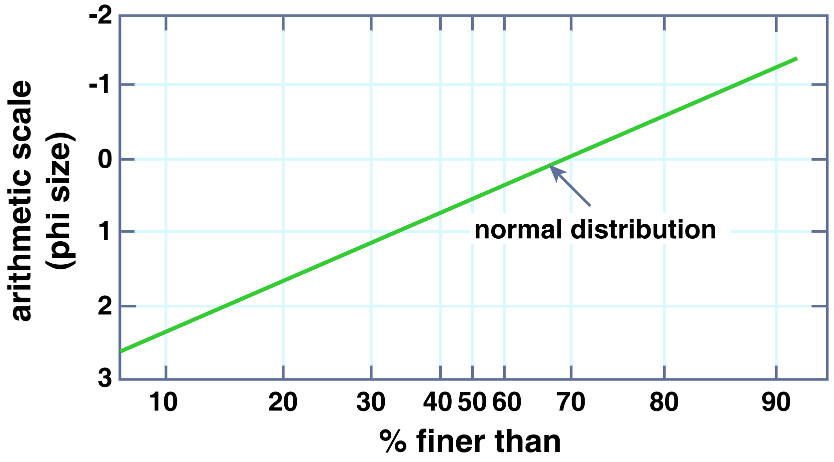


Figure by MIT OCW.

Figure 1-14: A normal frequency distribution plots as a straight line on probability graph paper

1.6.5 Some sediments have *two peaks on the size-frequency curve*; these are called ***bimodal distributions***, in contrast to the more common ***unimodal distributions*** (Figure 1-15). They are fairly common. Usually one of the maxima is dominant over the other. In rare cases there are more than two modes (polymodal distributions)

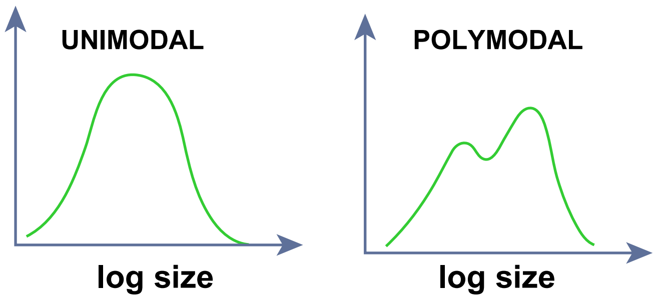


Figure by MIT OCW.

Figure 1-15: Unimodal and bimodal frequency distributions

1.6.6 It turns out that most skewed distributions show the relation between mean, the median, and the mode shown in Figure 1-16.

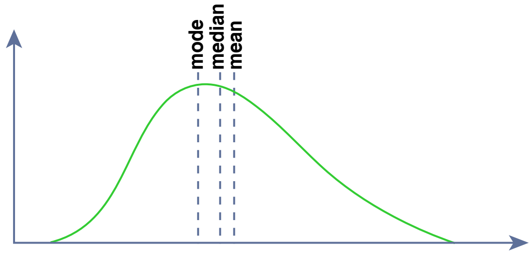


Figure by MIT OCW.

Figure 1-16: Cartoon of the relationships among the mean, the median, and the mode in a skewed frequency distribution

1.7 Things About Sediment Size

1.7.1 The size of material in a sediment is governed by two things, basically:

- the size of material available, and
- what has happened during transportation and deposition (complicated).

If only fine material is available, even an energetic transporting medium won't have sand or gravel to transport. One of the firmest interpretations we can make in sedimentology is that the presence of a sand deposit indicates a relatively energetic transporting medium, and a gravel deposit, even more so. (Assuming that the material really was deposited, and is not weathered in place.)

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

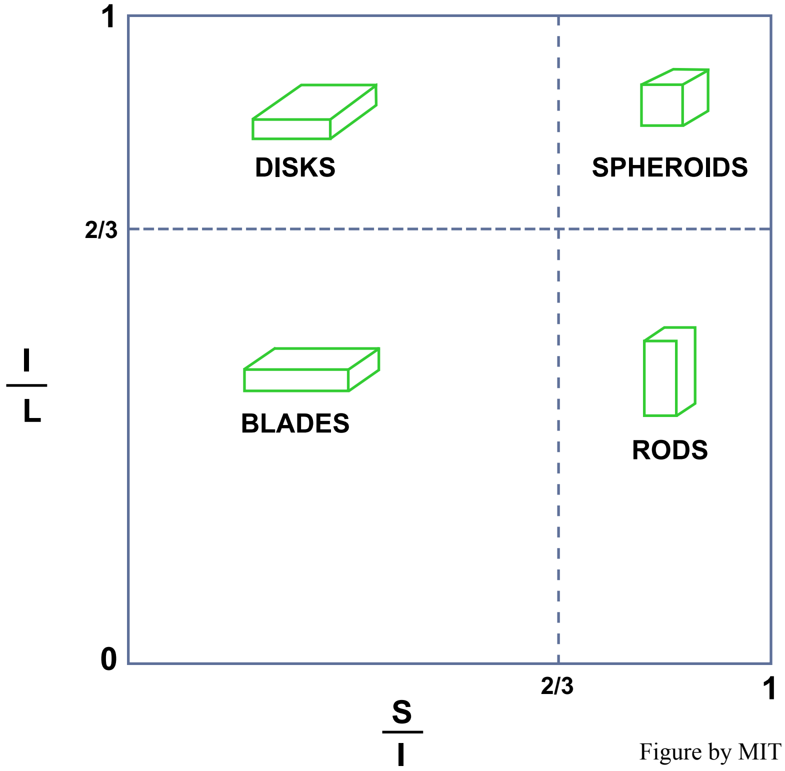


Figure by MIT OCW.

Figure 1-17: The Zingg diagram, to describe the shape of sediment particles that can be approximated by triaxial ellipsoids

Figure 1-18 is a chart for limiting values and terminology.

I/L	S/I	Shape
> 2/3	< 2/3	oblate (discoidal)
> 2/3	> 2/3	equiaxial (spheroidal)
< 2/3	< 2/3	triaxial (bladed)
< 2/3	> 2/3	prolate (rod-shaped)

Figure by MIT OCW.

Figure 1-18: Chart for limiting values and terminology of particle-shape classes in the Zingg diagram

Various more detailed diagrams of the same kind, with more shape categories within the same graph, have been proposed since. You can even contour the diagram with sphericity values (Figure 1-19).

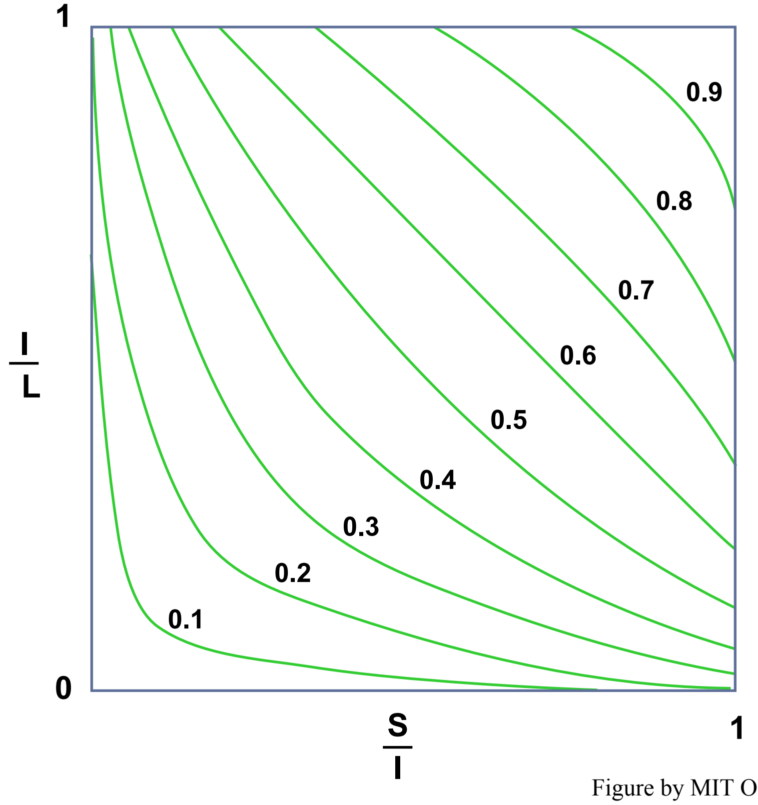


Figure by MIT OCW.

Figure 1-19: Contouring the Zingg diagram by sphericity

2.2.6 When you try measuring the long, intermediate, and short axes of real sediment particles for yourself, you'll find some ambiguity in defining these axes for irregular particles. What's usually done is to define L as the longest dimension through the particle, I as the longest dimension through the particle and perpendicular to L (but not necessarily intersecting L), and S as the longest dimension perpendicular to both L and I (but not necessarily passing through either L or I). In practice this works fairly well except for really weird shapes.

2.2.7 The Zingg diagram (or its more recent refinements) is about the best that can be done for description, but it still doesn't deal adequately with the hydrodynamic aspects of shape. It would be nice to have a measure of shape that correlates well with the settling behavior of the particles.

2.2.8 A sediment particle has a strong tendency to fall with its largest cross section approximately horizontal, for reasons having to do with the distribution of fluid pressure over the surface of the particle. The resistance to fall is governed by the difference in fluid pressure between the front (downward-facing) and back (upward-facing) surfaces and therefore by the largest cross-sectional area of the particle. On the other hand, the weight of the particle, which is the driving force for the settling in the first place, is proportional to the volume of the particle.

2.2.9 With this physics as a basis, Sneed and Folk long ago proposed something called the ***Maximum Projection Sphericity*** (MPS). It's defined as *the ratio of the maximum cross-sectional area of the volume-equivalent sphere to the maximum cross-sectional area of the particle itself.* For a triaxial ellipsoid, here's the mathematics:

Maximum cross-sectional area of particle:

$$\frac{\pi}{4}\ \mathrm{LI}$$

Volume of particle:

$$\frac{\pi}{6}\ \mathrm{LIS}$$

Diameter of volume-equivalent sphere:

$$\sqrt[3]{\mathrm{LIS}}$$

Maximum cross-sectional area of volume-equivalent sphere:

$$\frac{\pi}{4}\left(\sqrt[3]{\mathrm{LIS}}\right)^{2}$$

Maximum Projection Sphericity:

$$\frac{\pi}{4}\left(\sqrt[3]{\mathrm{LIS}}\right)^{2} \Big/ \frac{\pi}{4}\ \mathrm{LI} = \sqrt[3]{\frac{\mathrm{S}^{2}}{\mathrm{LI}}}$$

(1.10)

2.2.10 For a wide variety of regular shapes, there's a far better correlation between the MPS and settling velocity than between the sphericity defined above and settling velocity.

2.3 Roundness

2.3.1 ***Roundness*** has to do with *the sharpness of edges and corners* of a rock or mineral particle, and is almost entirely independent of shape defined above.

2.3.2 The official definition of ***roundness*** is this: *the ratio of average radius of curvature of edges and corners to the radius of the largest inscribed sphere*. A much more convenient alternative is to use the average radius of curvature of the corners of the projected grain image, and an even easier alternative is to use the radius of curvature of the single sharpest corner around the projected grain image.

2.3.3 But as you can easily imagine, even with simplifications this definition of roundness presents a formidable challenge to the measurer. When it's actually done, it's done by projecting or enlarging a grain image, fitting radii somehow, measuring, and computing. With the advent of high technology, one can use all sorts of automated border-tracing programs to speed things up enormously, but measurement of roundness still is not (and may never be) a standard sedimentological procedure.

2.3.4 Traditionally a series of five qualitative roundness grades has been used: angular, subangular, subrounded, rounded, and well rounded. Here are word descriptions of these grades:

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.

3. OTHER THINGS ABOUT THE TEXTURE OF DETRITAL SEDIMENTS

3.1 As I hope you know already, the word ***texture*** is used in a very specific way in sedimentology: it refers to *geometry on the scale of the constituent particles*. Two of the most important aspects of texture, *size* and *shape*, we've dealt with already. The other important aspect of texture comes under the term ***packing*** or ***fabric***: *the mutual arrangement of grains in the deposit*. This brief but important section outlines some things about what might be termed the *granular organization* of detrital sediments and sedimentary rocks.

3.2 The most natural assumption to make is that the detrital particles are originally deposited in a gravitationally stable framework; that is, each grain comes to rest on others below and is held there by its own weight. In aggregate, the grains are thus all in mutual contact.

3.3 Figure 1-20 is a cross-sectional view of a granular deposit. It's what a thin section of the deposit would look like if you impregnated the deposit with something to make it rigid and then cut a thin section. In such a section, you won't see many actual grain contacts.

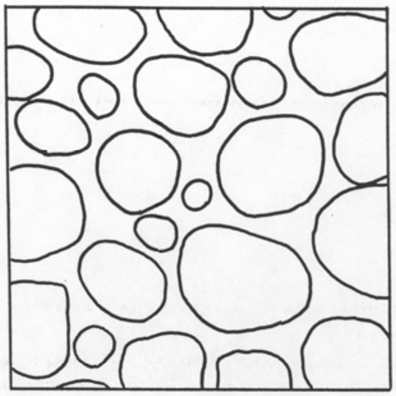


Figure 1-20 Cross-section view of a typical granular deposit

3.4 A simple geometrical property of such a packing arrangement is the ***porosity***, defined as *the volume of voids divided by the bulk volume for some volume of the sediment large enough to be representative* (i.e., containing a large number of grains). A somewhat more complicated but no less important property is the ***permeability***, which describes *the ease with which fluids can be forced through the porous sediment or rock under the influence of a gradient in fluid pressure*. There will be more detail on this important property of sediments and sedimentary rocks later in the course.

3.5 Two complications connected with this idea of a gravitationally stable framework are worth mentioning at this point:

(1) *It's strictly true only for large particles*, larger than something like a few tens of microns. Very fine particles weigh so little that forces of electrochemical origin are of equal importance. These forces have two important effects:

- *A thin layer or cushion of water molecules may be present around the particles, so the solid material of the particles may not really be in contact.* There's been considerable debate about the importance of such a water layer, but the possibility of its existence should be borne in mind.
- If they have any surface electric charge at all (and they typically do), *very fine particles become surrounded by a layer of ions called the **electric double layer***, which if the salinity is low effectively cushions the particles from one another. The existence and importance of this electric double layer is undoubted. I'll say a little more about it in the chapter on muds and shales.

(2) *Just because the particles are packed in mutual contact* doesn't *necessarily mean that the packing arrangement is a stable one*. The initial packing upon deposition is usually somewhat more open than the closest possible, and a little jiggling or spontaneous movement can cause the sediment to *rearrange itself into closer packing*. This might sound innocuous, but consider that while the particles are in the process of rearranging themselves they are *partly unsupported*, so *the rigidity of the framework disappears and the sediment can flow like a fluid*. This effect is called ***liquefaction***. Augmenting this effect is the displacement of that part of the pore fluid which all of a sudden finds itself in excess over what is needed in the closer packing. These effects are what give rise to the very common and diverse kinds of *soft-sediment deformation*. There will be a little more on such deformation in the chapter on structures.

3.6 What happens, in terms of geometry, when the sediment becomes lithified, by fairly deep burial? (Again I'm getting ahead of myself, because there will be more detail in this on the chapter on diagenesis.)

- *Deposition of new (authigenic) mineral material in the pore spaces*. This authigenic material, finely to coarsely crystalline, is called ***cement***. It's precipitated from solution, and it is unrelated to what was in the deposit before burial.
- *Reorganization of the detrital framework itself*, mainly by three processes:

deformation (change in shape) of the grains by intergranular forces. You can imagine the enormous magnitude of the forces per unit area that are brought to bear at grain contacts, in view of the great weight of overlying sediment that such grain contacts have to support. Relatively soft grains, like sedimentary rock fragments, can be squashed into unrecognizability (Figure 1-21).

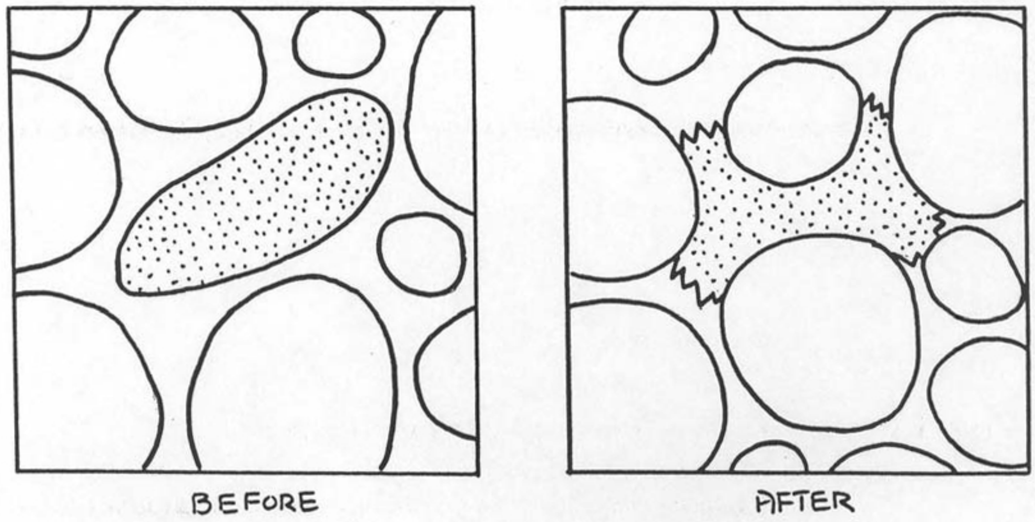


Figure 1-21 Squashing of relatively soft grains into unrecognizability during burial

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.

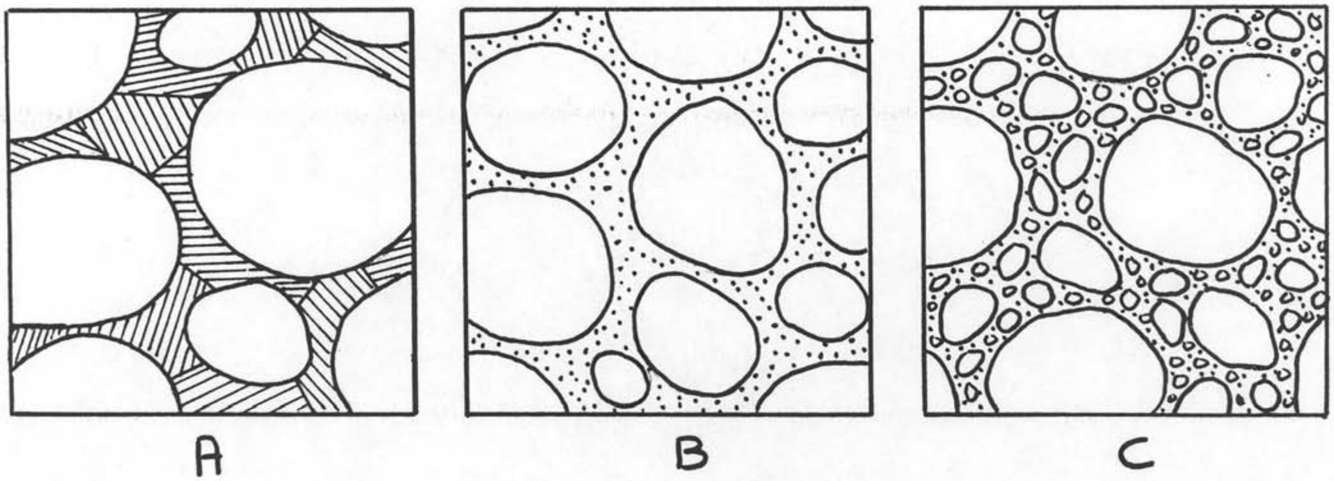


Figure 1-22 The granular organization of a clastic sediment. A: Framework grains are surrounded by precipitated mineral cement. B: Framework grains are set in a matrix of fine detrital sediment. C: In poorly sorted sediment, the distinction between framework and matrix is less natural.

- *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

because of low-temperature metastability. (But this is not to say there are not solution reactions that can eliminate quartz from sedimentary rocks; there are.)

- *The solubility of quartz in pure water is extremely small*, something like 10-15 ppm. This is measured indirectly: if we put a chunk of quartz in pure water, we would have to wait a prohibitively long time to see this equilibrium concentration of SiO_2 in solution. Most silica probably gets into solution by *weathering of silicate minerals*, or solution of amorphous silica in preexisting sediments (less important), not by solution of quartz.

- As noted in the chapter on shape and roundness, *abrasion of quartz during transportation is small but nonnegligible*. Transportation by wind reduces quartz grains much faster than does transportation by water. We still don't know whether well rounded quartz grains are mainly the result of (1) extreme multicycle subaqueous abrasion (traditional view); (2) desert wind action; (3) slight preferential solution of silica at edges and corners while temporarily stored in transport (untested possibility). Or perhaps some mix of these processes.

4.2.5 Because of the very low solubility of quartz, a good general rule in sedimentary petrology is that *as much quartz should reach sedimentary basins as leaves the source area* (although not necessarily in the same shape and size).

4.2.6 What's the *provenance* of quartz? (The word ***provenance*** is used to describe *the place and/or the materials from which a sediment is derived*.) The provenance of quartz grains is varied: the ultimate source of detrital quartz is igneous and metamorphic rocks, but obviously preexisting quartz-bearing sedimentary rocks provide detrital quartz too. A certain percentage, very high in some areas, is recycled from such preexisting sedimentary rocks.

4.2.7 Here's a brief rundown on *the particulate form of quartz as a function of grain size*:

- Most *gravel-size* quartz clasts are derived from *vein quartz*; they're usually polycrystalline and milky white.
- *Sand-size* quartz consists mainly of single quartz crystals, but also polycrystalline grains, either quartzite grains, recrystallized chert, or primary crystalline aggregates from crystalline source rocks.
- *Silt-siz*e quartz grains are almost exclusively single-crystal grains (derivation: fine-grained metamorphic rocks; preexisting fine-grained sediments; chipping and spalling of larger grains, perhaps mainly in deserts).
- There's not much *clay-size* quartz around to worry about.

4.2.8 *What's the size of quartz grains in sedimentary rocks?* It ranges from middling gravel sizes down into the very coarsest clay sizes. There's almost no quartz $< 2\ \mu m$ (and unfortunately, the clay–silt boundary is at $4\ \mu m$). I mentioned earlier that there seems to be a shortage of terrigenous sedimentary material 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:

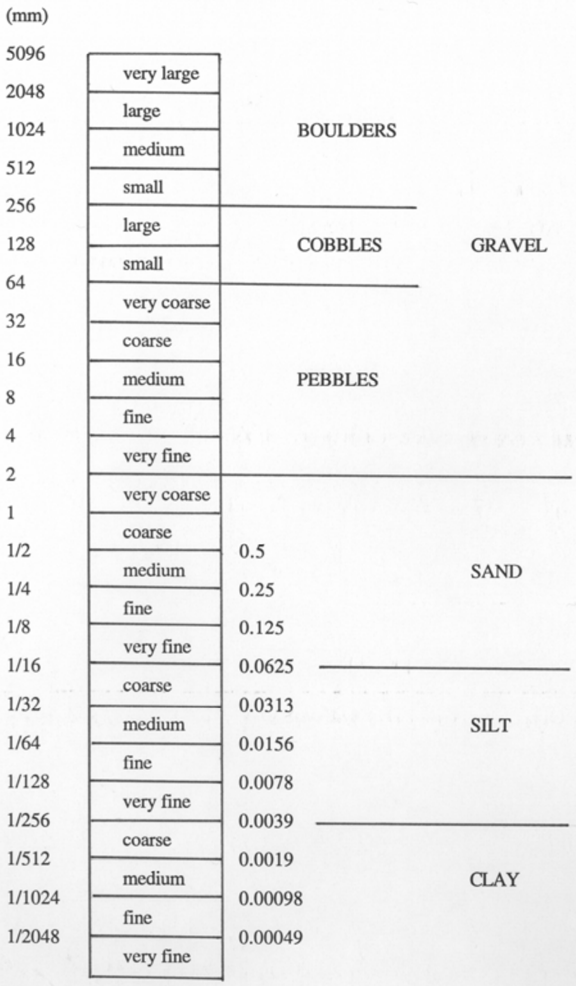
- *The heavies tend to be in the finest fractions of the detritus*. Their greater density means that their settling velocity is greater for a given mass, so finer heavies tend to be deposited with coarser lights. (This is a considerable oversimplification of what's called *hydraulic equivalence* of light 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?

FIGURE 1-24 The Udden-Wentworth grade scale



(mm)				
5096	very large			
2048	large			
1024	medium		BOULDERS	
512	small			
256	large			
128	small		COBBLES	GRAVEL
64	very coarse			
32	coarse			
16	medium		PEBBLES	
8	fine			
4	very fine			
2	very coarse			
1	coarse			
1/2	medium	0.5		SAND
1/4	fine	0.25		
1/8	very fine	0.125		
1/16	coarse	0.0625		
1/32	medium	0.0313		SILT
1/64	fine	0.0156		
1/128	very fine	0.0078		
1/256	coarse	0.0039		
1/512	medium	0.0019		CLAY
1/1024	fine	0.00098		
1/2048	very fine	0.00049		

5.3 Various classifications of mixed-size sediments have been proposed. Two such classifications are shown in Figure 1-25. The one on the left is very logical and symmetrical but largely ignores the real world of naming mixed sediments, and the one on the right, while asymmetrical, takes greater account of actual sedimentological practice. I don’t want to denigrate any aspect of sedimentary geology in this course, but I think it’s reasonable to point out here that not many sedimentologists worry much about the precise placement of boundaries in classifications like this: they just use their own judgment or predilections or prejudices in giving a name to a mixed sediment.

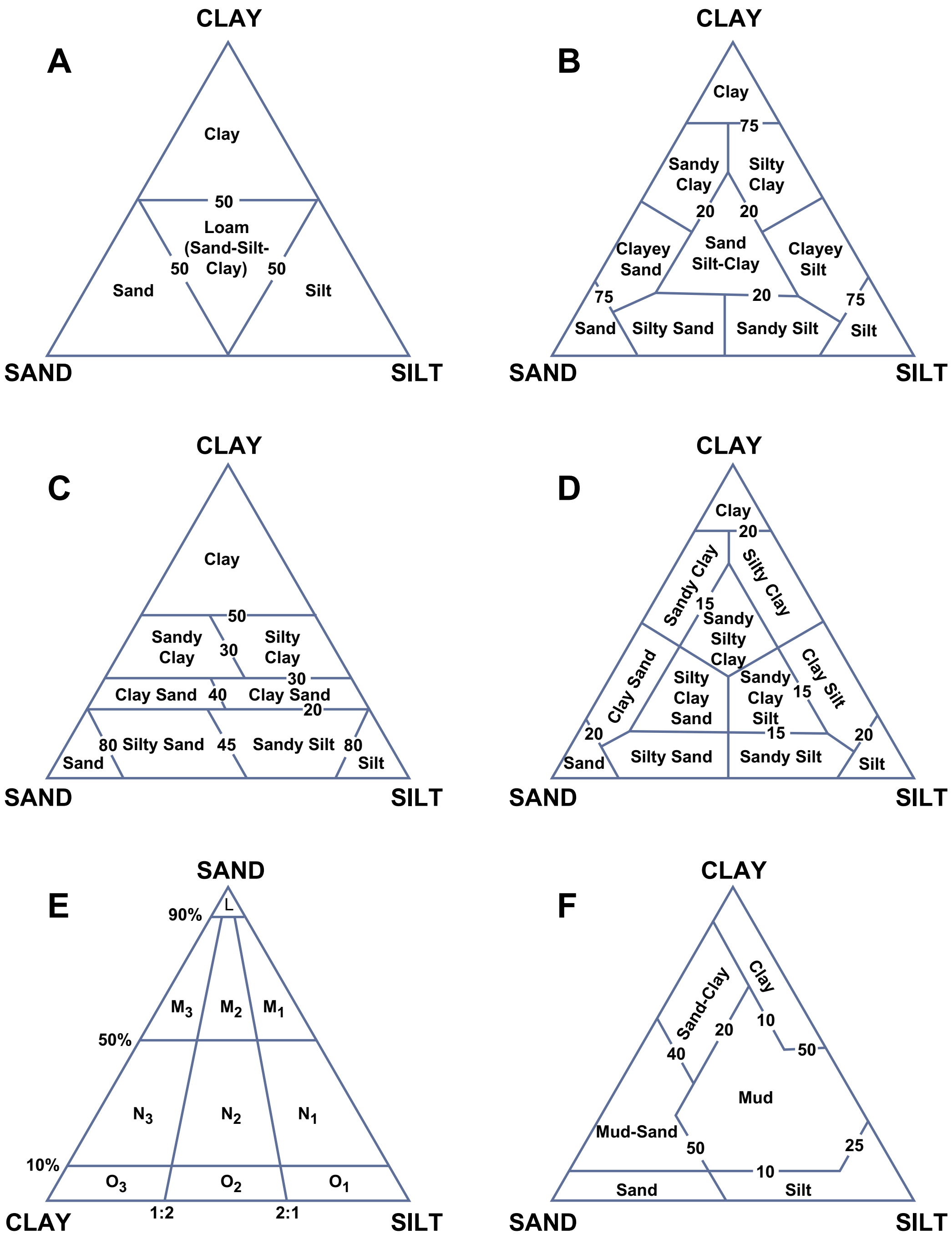


Figure by MIT OCW.

Figure 1-25: Several different classifications of sediments that contain mixtures of the various named size classes