

2. POROSITY

2.1 Theory

The *porosity* of a rock is the fraction of the volume of space between the solid particles of the rock to the total rock volume. The space includes all pores, cracks, vugs, inter- and intra-crystalline spaces. The porosity is conventionally given the symbol f , and is expressed either as a fraction varying between 0 and 1, or a percentage varying between 0% and 100%. Sometimes porosity is expressed in ‘porosity units’, which are the same as percent (i.e., 100 porosity units (pu) = 100%).

However, the fractional form is **ALWAYS** used in calculations.

Porosity is calculated using the relationship

$$f = \frac{V_{pore}}{V_{bulk}} = \frac{V_{bulk} - V_{matrix}}{V_{bulk}} = \frac{V_{bulk} - (W_{dry} / \rho_{matrix})}{V_{bulk}}, \quad (2.1)$$

where: V_{pore} = pore volume

V_{bulk} = bulk rock volume

V_{matrix} = volume of solid particles composing the rock matrix

W_{dry} = total dry weight of the rock

ρ_{matrix} = mean density of the matrix minerals.

It should be noted that the porosity does not give any information concerning pore sizes, their distribution, and their degree of connectivity. Thus, rocks of the same porosity can have widely different physical properties. An example of this might be a carbonate rock and a sandstone. Each could have a porosity of 0.2, but carbonate pores are often very unconnected resulting in its permeability being much lower than that of the sandstone. A range of differently defined porosities are recognized and used within the hydrocarbon industry. For rocks these are:

- | | | |
|--------|------------------------|---|
| (i) | Total porosity | Defined above. |
| (ii) | Connected porosity | The ratio of the connected pore volume to the total volume. |
| (iii) | Effective porosity | The same as the connected porosity. |
| (iv) | Primary porosity | The porosity of the rock resulting from its original deposition. |
| (v) | Secondary porosity | The porosity resulting from diagenesis. |
| (vi) | Microporosity | The porosity resident in small pores (< 2 μm) commonly associated with detrital and authigenic clays. |
| (vii) | Intergranular porosity | The porosity due to pore volume between the rock grains. |
| (viii) | Intragranular porosity | The porosity due to voids within the rock grains. |
| (ix) | Dissolution porosity | The porosity resulting from dissolution of rock grains. |
| (x) | Fracture porosity | The porosity resulting from fractures in the rock at all scales. |
| (xi) | Intercrystal porosity | Microporosity existing along intercrystalline boundaries usually in carbonate rocks. |
| (xii) | Moldic porosity | A type of dissolution porosity in carbonate rocks resulting in molds of original grains or fossil remains. |
| (xiii) | Fenestral porosity | A holey (‘bird’s-eye’) porosity in carbonate rocks usually associated with algal mats. |
| (xiv) | Vug porosity | Porosity associated with vugs, commonly in carbonate rocks. |

2.2 Controls on Porosity

The initial (pre-diagenesis) porosity is affected by three major microstructural parameters. These are grain size, grain packing, particle shape, and the distribution of grain sizes. However, the initial porosity is rarely that found in real rocks, as these have subsequently been affected by secondary controls on porosity such as compaction and geochemical diagenetic processes. This section briefly reviews these controls.

2.2.1 Grain Packing

The theoretical porosities for various grain packing arrangements can be calculated. The theoretical *maximum* porosity for a cubic packed rock made of spherical grains of a uniform size is 0.4764, and is independent of grain size. The maximum porosity of other packing arrangements is shown in Table 2.1 and Figure 2.1.

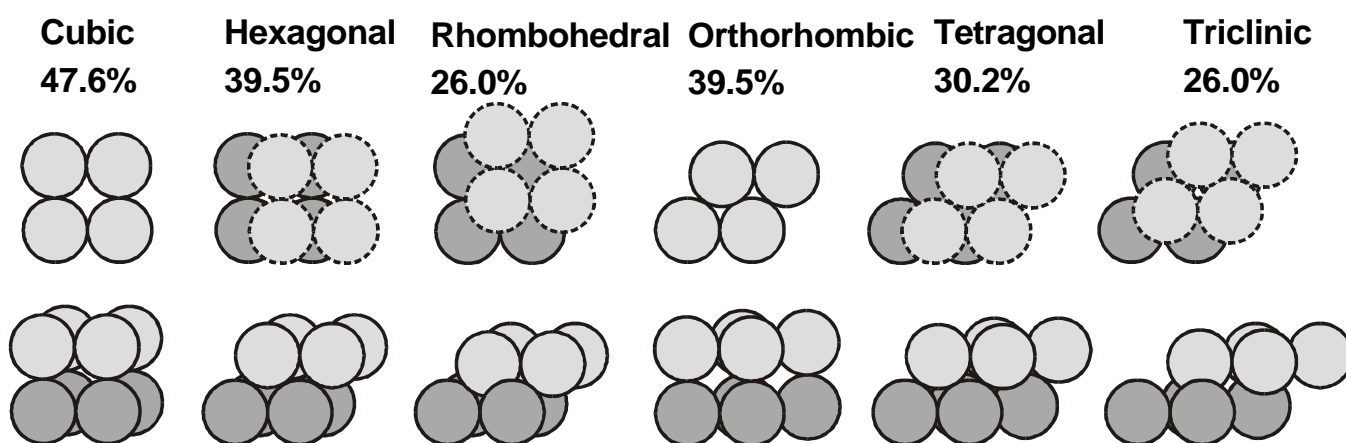


Figure 2.1 Ordered packing arrangements.

Table 2.1 Maximum porosity for different packing arrangements

Packing	Maximum Porosity (fractional)
Random	≥ 0.399 (dependent on grain size)
Cubic	0.476
Hexagonal	0.395
Orthorhombic	0.395
Rhombohedral	0.260
Tetragonal	0.302
Triclinic	0.260

The calculations of these ideal porosities is relatively simple. For example, taking the cubic arrangement of identical spheres of radius r occupying a cubic unit cell of length L , as shown in Fig. 2.2, the following calculation is possible.

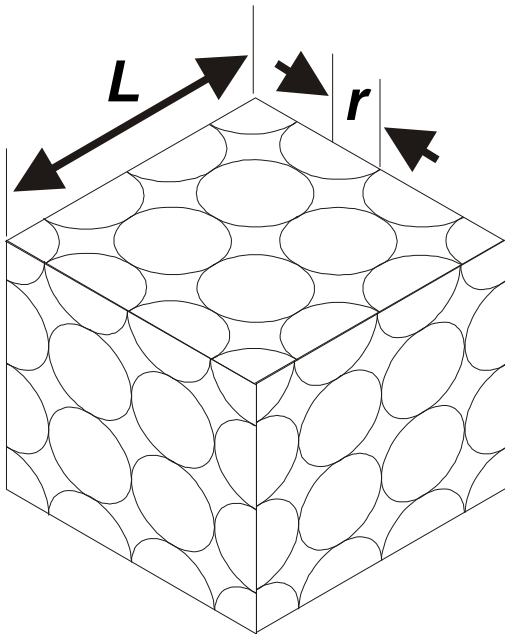


Figure 2.2 Cubic packing of identical spheres.

The bulk volume of the cell $V_{bulk} = L^3$, and the number of spheres in the cell $n = (L/2r)^3$. Hence the volume of the matrix $V_{matrix} = (4 n \pi r^3)/3 = (L/2r)^3 (4 \pi r^3)/3 = (\pi L^3)/6$. The porosity can now be calculated from Eq. (2.1) as

$$f = \frac{L^3 - (\pi L^3 / 6)}{L^3} = (1 - \pi / 6) = 0.4764 \tag{2.2}$$

which is independent of the sphere size. Most of the other values in Table 2.1 can be calculated similarly, although with a little more difficulty as a result of their different packing geometries. There are 6 different ways that identical spheres can be packed, and these are shown in Fig. 2.1.

2.2.2 Grain Size

It was noted above that the ordered cubic packing of identical sphere leads to a porosity that is grain size independent. This is also true for the other ordered packing lattices, but not true for the random arrangement of spheres. In real depositional environments, ordered packings are not formed because they are energetically unstable, and the grains become randomly distributed.

The equilibrium porosity of a porous material composed of a random packing of spherical grains is dependent upon the stability given to the rock by frictional and cohesive forces operating between individual grains. These forces are proportional to the exposed surface area of the grains. The specific surface area (exposed grain surface area per unit solid volume) is inversely proportional to grain size. This indicates that, when all other factors are equal, a given weight of coarse grains will be stabilized

at a lower porosity than the same weight of finer grains. For a sedimentary rock composed of a given single grain size this general rule is borne out in Figure 2.3. It can be seen that the increase in porosity only becomes significant at grain sizes lower than 100 μm , and for some recent sediments porosities up to 0.8 have been measured. As grain size increases past 100 μm , the frictional forces decrease and the porosity decreases until a limit is reached that represents random frictionless packing, which occurs at 0.399 porosity, and is independent of grain size. No further loss of porosity is possible for randomly packed spheres, unless the grains undergo irreversible deformation due to dissolution-recrystallisation, fracture, or plastic flow, and all such decreases in porosity are termed *compaction*.

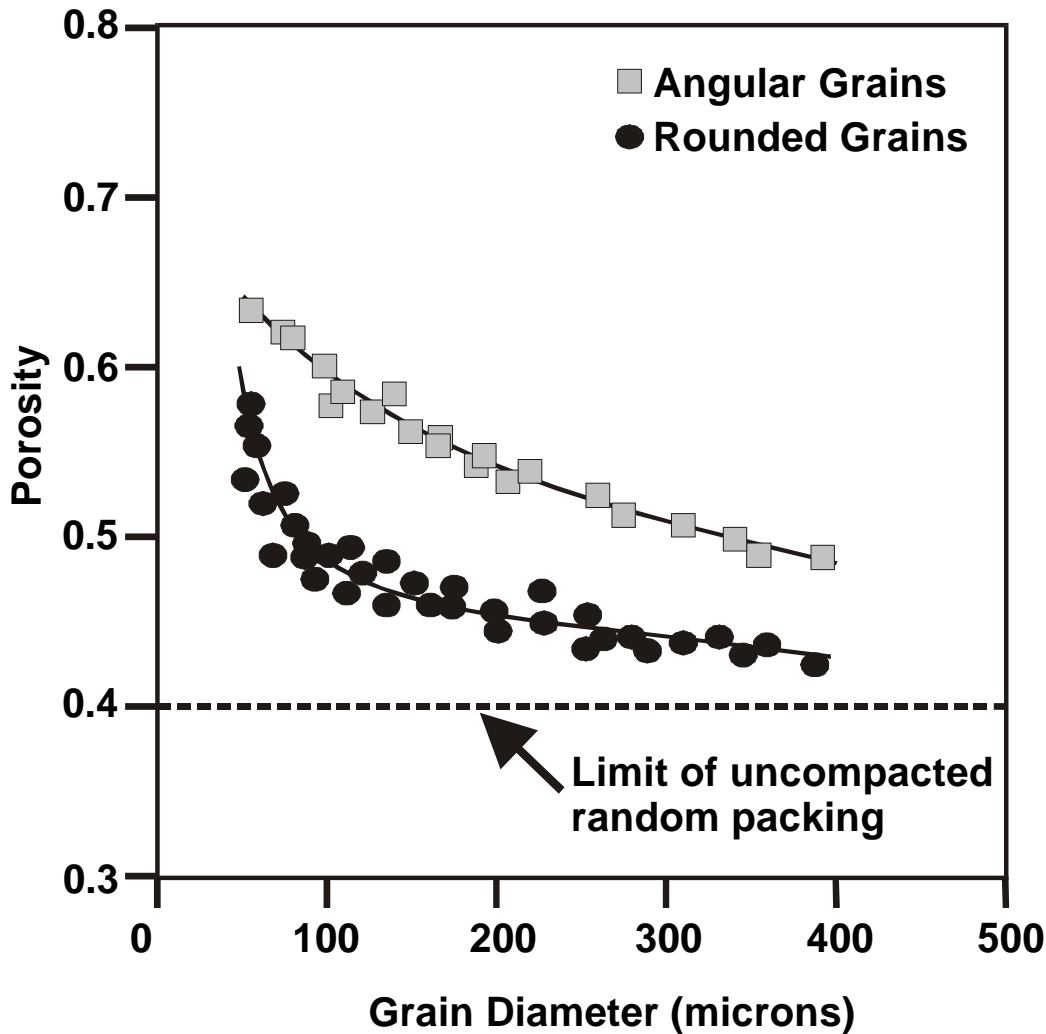


Figure 2.3 Relation between porosity, grain size and grain shape

Rarely, in borehole petrophysics do we need to look at accurate grain size determinations as many of the tools that we use have minimum vertical resolutions of the order of tens of centimetres. However, as well logs are correlated to core logging it is well to bear in mind the agreed semi-quantitative classifications for grain size in siliclastic and carbonate rocks (Tables 2.2 and 2.3).

Table 2.2 Siliclastic grain size definitions

Category	Median Grain Size (mm)
Gravel	2000
Very Coarse	1000
Coarse	500
Medium	250
Fine	125
Very Fine	62
Silt	

Table 2.3 Carbonate grain size definitions

Category	Median Grain Size (mm)
L	400
M	200
F	100
VF	50
EF	
Symbol	Pore Size (mm)
B	<100
C	100-200
D	>200
Matrix Code	Texture
I	Compact
II	Chalky
III	Sucrose

2.2.3 Grain Shape

This parameter is not widely understood. Several studies have been carried out on random packings of non-spherical grains, and in all cases the resulting porosities are larger than those for spheres. Table 2.4 shows data for various shapes, where the porosity is for the frictionless limit. Figure 2.3 shows data comparing rounded and angular grains, again showing that the porosity for more angular grains is larger than those that are sub-spherical.

Table 2.4 The effect of grain shape on porosity

Grain Shape	Maximum Porosity (fractional)
Sphere	≥ 0.399 (dependent on grain size)
Cube	0.425
Cylinder	0.429
Disk	0.453

2.2.4 Grain Size Distribution

Real rocks contain a distribution of grain sizes, and often the grain size distribution is multi-modal. The best way of understanding the effect is to consider the variable admixture of grains of two sizes (Figure 2.4). The porosity of the mixture of grain sizes is reduced below that for 100% of each size. There are two mechanisms at work here. First imagine a rock with two grain sizes, one of which has 1/100th the diameter of the other. The first mechanism applies when there are sufficient of the larger grains to make up the broad skeleton of the rock matrix. Here, the addition of the smaller particles reduces the porosity of the rock because they can fit into the interstices between the larger particles. The second mechanism is valid when the broad skeleton of the rock matrix is composed of the smaller grains. There small grains will have a pore space between them. Clearly, if some volume of these

grains are removed and replaced with a single solid larger grain, the porosity will be reduced because both the small grains and their associated porosity have been replaced with solid material. The solid lines GR and RF or RM in Figure 2.4 represent the theoretical curves for both processes. Note that as the disparity between the grain sizes increases from 6:3 to 50:5 the actual porosity approaches the theoretical lines. Note also that the position of the minimum porosity is not sensitive to the grain diameter ratio. This minimum occurs at approximately 20 to 30% of the smaller particle diameter. In real rocks we have a continuous spectrum of grain sizes, and these can give rise to a complex scenario, where fractal concepts become useful.

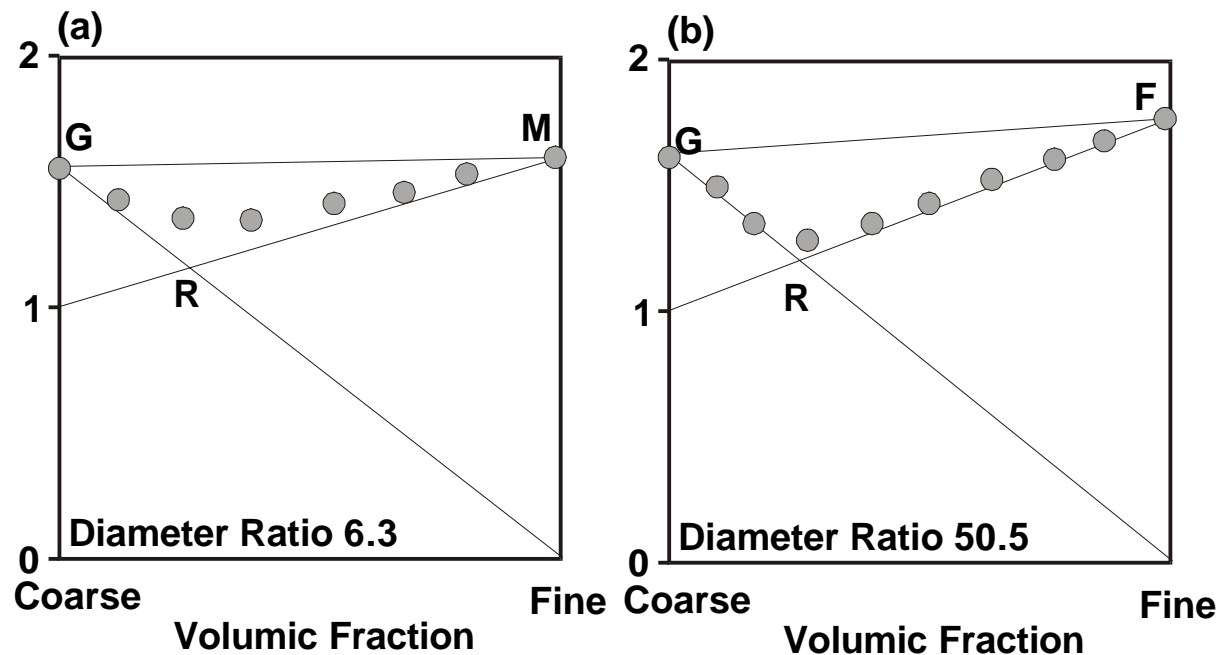


Figure 2.4 Apparent volume for assemblages of spheres of different diameter ratios.

In practice such detail is rarely entered into. Grain size distributions can be inferred from mercury porisimetry measurements that will be studied later in the MSc course, or by direct sieving of a disaggregated sample. In the sieving technique, the dried, weighed and disaggregated sample is placed in the topmost of a stack of sieves with decreasing mesh size D . A motor vibrates the stack such that given particle sizes remain in each of the sieves. The cumulative percentage weight of these remains are plotted against the mesh size of the sieves D , resulting in a distribution curve similar to that in Fig. 2.5. The degree of sorting (grain size distribution) is expressed by the *Trask Coefficient* s_o , which is expressed as

$$s_o = \sqrt{\frac{D_{25}}{D_{75}}}, \tag{2.3}$$

where: D_{25} = mesh size (grain diameter) at 25% of sample weight (larger grain fraction).
 D_{75} = mesh size (grain diameter) at 75% of sample weight (all but smallest grain fraction).

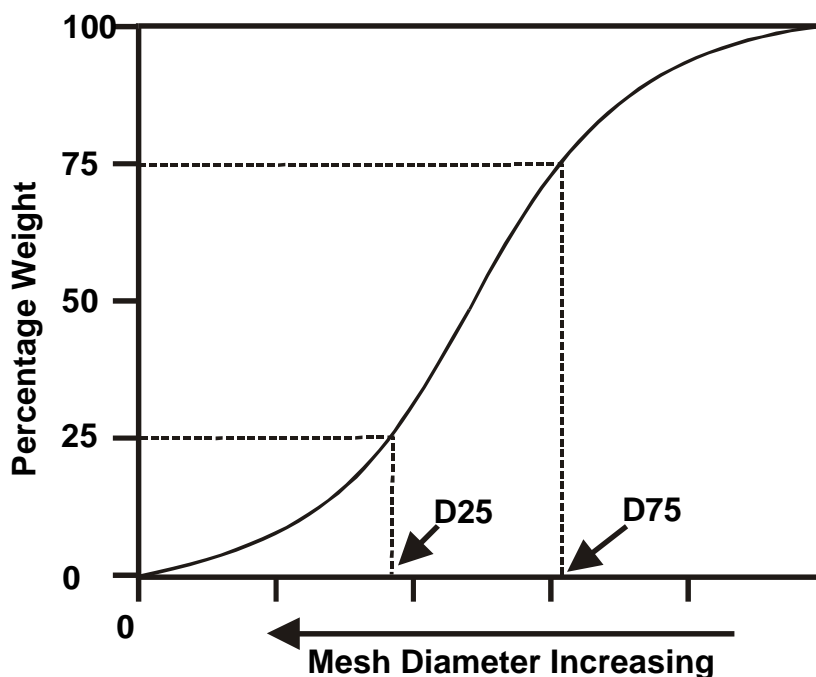


Figure 2.5 Cumulative grain size distributions by sieved weight.

In general the porosity of a sandstone decreases as the degree of sorting decreases, where the degree of sorting is expressed as a Trask coefficient or classified according to Table 2.5.

Table 2.5 The Trask sorting classification

Sorting Classification	Trask Coefficient
Extremely well	1.00
Very well	1.10
Well	1.20
Moderately	1.40
Poorly	2.00
Very poorly	2.70
	5.70

2.2.5 Secondary Controls on Porosity

Porosity is also controlled by a huge range of secondary processes that result in compaction and dilatation. These can be categorized into (i) mechanical processes, such as stress compaction, plastic deformation, brittle deformation, fracture evolution etc., and (ii) geochemical processes, such as dissolution, reprecipitation, volume reductions concomitant upon mineralogical changes etc. The description of these process will be examined in the sedimentary techniques courses.

2.2.6 The Range of Porosity Values in Nature

In recently deposited, unconsolidated sediments, such as those that you might find on the floor of a lake, porosity may be very high (values up to 80% have been recorded). However, more common materials, such as loose sands, can have porosities as high as 45% that are either extremely unstable or stabilized by cements. High porosities can also occur when the porosity is due to dissolution (secondary porosity), particularly in carbonates. In the case of carbonates the total porosity may be very high, but their permeability can be very low as the pores and vugs that make up the pore structure are unconnected. Similarly, porosities can be very low. In massive fractured carbonates it can commonly be as low as 1%, and igneous and metamorphic rocks almost always have porosities less than 1%. Sandstones, generally, lie in the range 5% to 20%. Table 2.6 gives approximate ranges of porosities for some common lithologies.

Table 2.6 The range of porosity values for rocks.

Lithology	Porosity Range (%)
Unconsolidated sands	35-45
'Reservoir' Sandstones	15-35
Compact Sandstones	5-15
Shales	0-45
Clays	0-45
Massive Limestones	5-10
Vuggy Limestones	10-40
Dolomite	10-30
Chalk	5-40
Granite	<1
Basalt	<0.5
Gneiss	<2
Conglomerate	1-15

2.3 Porosity Determination

The best way of determining porosity is to carry out experiments on core extracted from the well. These techniques will be examined in detail in the Formation Evaluation course later in the MSc. However, the basic techniques will be described here. It should be noted that core determined porosities have a much higher degree of accuracy than porosities determined from down-hole tools, but suffer from sampling problems. Taken together core and borehole determined porosities optimize accuracy and high resolution sampling.

There are at least 4 common methods of measuring the porosity of a core. These are:

- **Buoyancy**
- **Helium porosimetry**
- **Fluid saturation**
- **Mercury porosimetry**

The first three will be described here, and the last will be described in Chapter 4.

2.3.1 Saturation Method

This method only measures the effective (connected) porosity of the rock as it depends on the saturation of the rock with a fluid.

- Clean and dry the rock.
- Weigh the rock in its dry state to give the dry weight, W_{dry} .
- Fully saturate the rock in a wetting fluid. Conventionally toluene or dichloroethene were used, but now it is more common to saturate the rock with a brine that has been made to mimic that in the reservoir, i.e., contain the same concentrations of major dissolved salts (a *synthetic brine*).
- Weigh the saturated sample after drying any excess fluid from its surface to give its saturated weight, W_{sat} .
- Determine the bulk volume of the rock, V_{bulk} . This can be done using measuring calipers and assuming the core to be cylindrical, but is prone to large errors even if the core is almost cylindrical. Another method is to immerse the sample in mercury, which is a non-wetting fluid and will not enter the pores of a rock at atmospheric pressure, and measure the volume of the mercury displaced. This method used to be common, but the sample can then no longer be used for experiments and must be safely disposed of. Archimedes method is preferred, which will be described in the Formation Evaluation course.
- Determine the density ρ_{fluid} of the saturating fluid by weighing a known volume of it.

The porosity is then

$$f = \frac{V_{bulk} - V_{matrix}}{V_{bulk}} = \frac{(W_{sat} - W_{dry}) / \rho_{fluid}}{V_{bulk}} \quad (2.4)$$

Note that the mean grain density of the sample can also be calculated from this data using the relationship

$$\rho_{ma} = \frac{W_{dry}}{V_{matrix}} = \frac{W_{dry}}{V_{bulk} - ((W_{sat} - W_{dry}) / \rho_{fluid})} \quad (2.5)$$

2.3.2 Buoyancy Method

This method is somewhat similar to the saturation method, and also measures the effective (connected) porosity.

- Clean and dry the rock.
- Weigh the rock in its dry state to give the dry weight, W_{dry} .
- Fully saturate the rock in a wetting fluid as before.
- Weigh the saturated sample suspended in a bath of the same fluid with which it was saturated to give its suspended weight, W_{sus} . This is shown in Fig. 2.6. Note that a cradle is needed to suspend the sample in the fluid, and this has a weight W_{cradle} when in the fluid. So the actual measured weight when the sample and cradle are suspended in the fluid is $(W_{sus} + W_{cradle})$. The effect of the cradle must be taken away, hence the weight of the cradle when suspended in the fluid (up to the same level), W_{cradle} , must be taken account of in the final calculation
- Determine the bulk volume of the rock, V_{bulk} as before.

- Determine the density ρ_{fluid} of the saturating fluid by weighing a known volume of it.

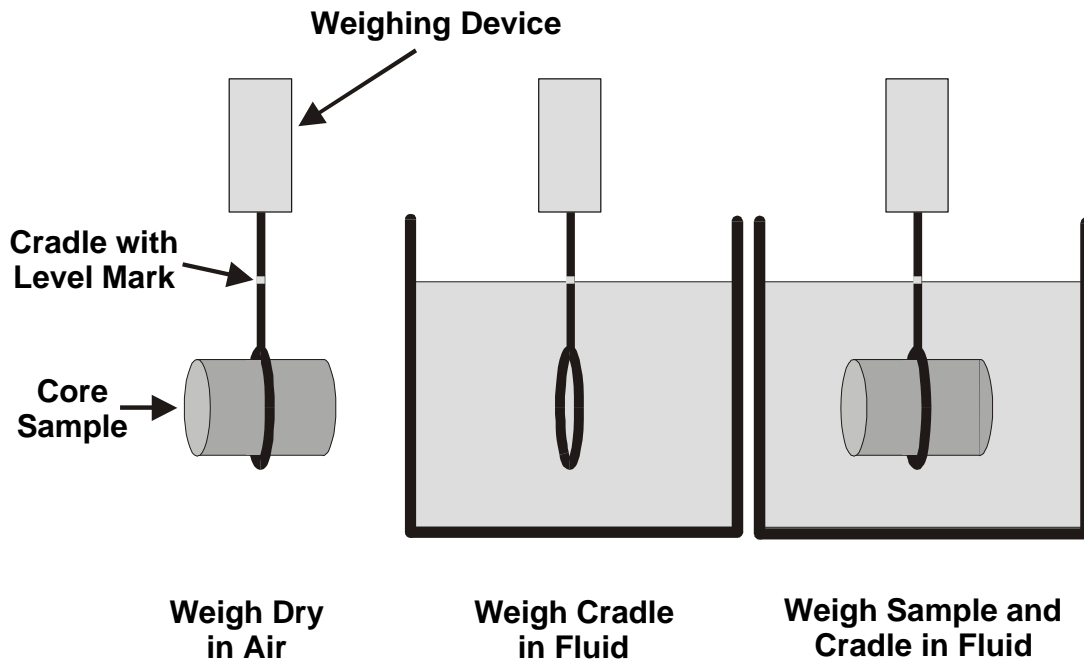


Figure 2.6 The buoyancy method for porosity determination.

The porosity is given by the relationship

$$f = \frac{V_{bulk} - V_{matrix}}{V_{bulk}} = \frac{V_{bulk} - (W_{dry} - (W_{sus} + W_{cradle}) + W_{cradle}) / \rho_{fluid}}{V_{bulk}} \quad (2.6)$$

Again, the mean grain density can also be calculated as

$$\rho_{ma} = \frac{W_{dry}}{V_{matrix}} = \frac{W_{dry}}{(W_{dry} - (W_{sus} + W_{cradle}) + W_{cradle}) / \rho_{fluid}} \quad (2.7)$$

2.3.3 Helium Porisimetry

This method relies on the expansion of helium gas, and also measures the effective porosity of the rock. However, since helium is a slippery little molecule, it can penetrate pores which are much smaller than those that the fluid used in the two previous methods can. Hence, this method provides slightly higher porosity measurements on any given rock sample.

This method uses the apparatus shown in Fig. 2.7, together with the following procedure.

- Clean and dry the core sample.
- Obtain the bulk volume of the sample as in the previous methods.

- Ensure Valve 2 is closed.
- Gas is introduced into Chamber A of the apparatus using Valve 1 until a pressure P_i is reached, when Valve 1 is closed. This pressure is arbitrary, and commonly about 100 psi.
- Insert the sample into Chamber B, which is at atmospheric pressure, P_{atm} , and ensure Valve 3 is closed.
- Slowly open Valve 2 to let the gas equilibrate through the entirety of both chambers. The gas will penetrate into the pores of the rock sample. During this process the pressure will decrease to a new stable level, P_f .

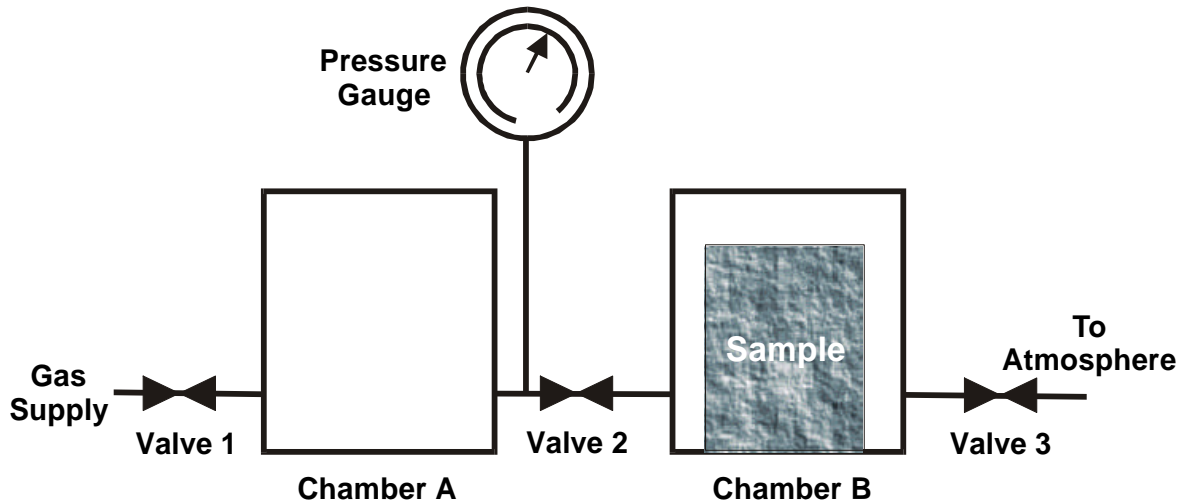


Figure 2.7 Helium expansion porisimeter.

The drop of pressure depends upon how much space there is in Chamber B, and that depends on how much of Chamber B is occupied by solid rock particles. The measured gas pressures P_i , P_{atm} and P_f alone are not sufficient to obtain the volume of the sample rock matrix. However, we can calibrate the system by putting a range of metal cylinders of accurately known volume into Chamber B and repeating the experiment. When this is done, the calibration constant, and the pressures allow the volume of the rock matrix V_{matrix} to be obtained. The porosity is then calculated using

$$f = \frac{V_{bulk} - V_{matrix}}{V_{bulk}} \quad . \quad (2.8)$$