Chapter 2
Gravity methods

2.1 INTRODUCTION

Gravity surveying measures variations in the Earth's gravitational field caused by differences in the density of sub-surface rocks. Although known colloquially as the 'gravity' method, it is in fact the variation of the *acceleration* due to gravity that is measured. Gravity methods have been used most extensively in the search for oil and gas, particularly in the early twentieth century. While such methods are still employed very widely in hydrocarbon exploration, many other applications have been found (Table 2.1), some examples of which are described in more detail in Section 2.7.

Micro-gravity surveys are those conducted on a very small scale – of the order of hundreds of square metres – and which are capable of detecting cavities, for example, as small as 1 m in diameter within 5 m of the surface.

Perhaps the most dramatic change in gravity exploration in the 1980s has been the development of instrumentation which now permits *airborne* gravity surveys to be undertaken routinely and with a high degree of accuracy (see Section 2.5.7). This has allowed aircraft-borne gravimeters to be used over otherwise inaccessible terrain and has led to the discovery of several small but significant areas with economic hydrocarbon potentials.

Table 2.1 Applications of gravity surveying

Hydrocarbon exploration
Regional geological studies
Isostatic compensation determination
Exploration for, and mass estimation of, mineral deposits
Detection of sub-surface cavities (micro-gravity)
Location of buried rock-valleys
Determination of glacier thickness
Tidal oscillations
Archaeogeophysics (micro-gravity); e.g. location of tombs
Shape of the earth (geodesy)
Military (especially for missile trajectories)
Monitoring volcanoes

2.2 PHYSICAL BASIS

2.2.1 Theory

The basis on which the gravity method depends is encapsulated in two laws derived by Sir Isaac Newton, which he described in *Principia Mathematica* (1687) – namely his Universal Law of Gravitation, and his Second Law of Motion.

The first of these two laws states that the force of attraction between two bodies of known mass is directly proportional to the product of the two masses and inversely proportional to the square of the distance between their centres of mass (Box 2.1). Consequently, the greater the distance separating the centres of mass, the smaller is the force of attraction between them.

Box 2.1 Newton's Universal Law of Gravitation

$$\text{Force} = \text{gravitational constant} \times \frac{\text{mass of Earth } (M) \times \text{mass } (m)}{(\text{distance between masses})^2}$$

$$F = \frac{G \times M \times m}{R^2} \qquad \text{(equation (1))}$$

where the gravitational constant $(G) = 6.67 \times 10^{-11}\,\mathrm{N\,m^2\,kg^{-2}}$

Newton's law of motion states that a force (F) is equal to mass (m) times acceleration (Box 2.2). If the acceleration is in a vertical direction, it is then due to gravity (g).

Box 2.2 Newton's Second Law of Motion

$$\text{Force} = \text{mass } (m) \times \text{acceleration } (g)$$

$$F = m \times g \qquad \text{(equation (2))}$$

Equations (1) and (2) can be combined to obtain another simple relationship:

$$F = \frac{G \times M \times m}{R^2} = m \times g; \quad \text{thus } g = \frac{G \times M}{R^2} \quad \text{(equation (3))}.$$

This shows that the magnitude of the acceleration due to gravity on Earth (g) is directly proportional to the mass (M) of the Earth and inversely proportional to the square of the Earth's radius (R). Theoretically, acceleration due to gravity should be constant over the Earth. In reality, gravity varies from place to place because the Earth has the shape of a flattened sphere (like an orange or an inverted pear),

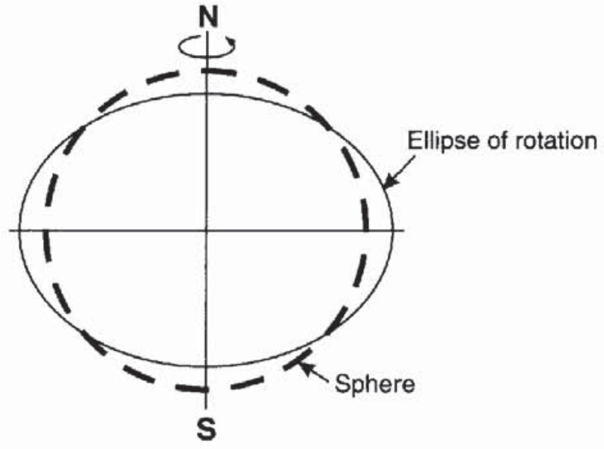

Figure 2.1 Exaggerated difference between a sphere and an ellipse of rotation (spheroid)

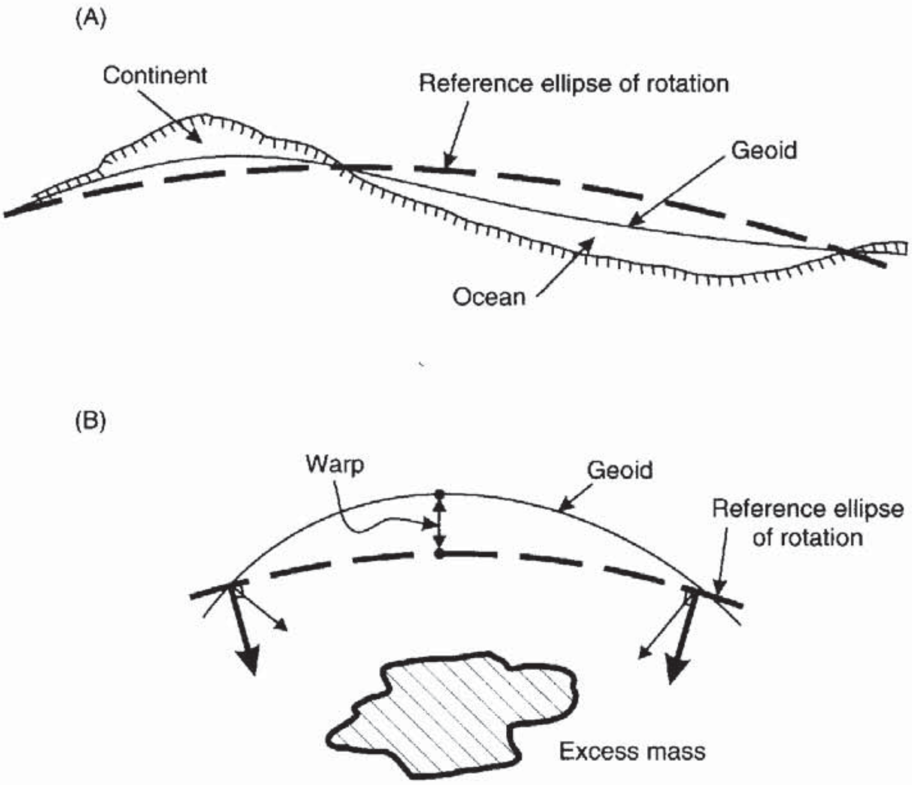

Figure 2.2 Warping of the geoid: (A) continental-scale effects, and (B) localised effects due to a sub-surface excess mass

rotates, and has an irregular surface topography and variable mass distribution (especially near the surface).

The shape of the Earth is a consequence of the balance between gravitational and centrifugal accelerations causing a slight flattening to form an oblate spheroid. Mathematically it is convenient to refer to the Earth's shape as being an *ellipse of rotation* (Figure 2.1).

The sea-level surface, if undisturbed by winds or tides, is known as the *geoid* and is particularly important in gravity surveying as it is horizontal and at right angles to the direction of the acceleration due to gravity everywhere. The geoid represents a surface over which the gravitational field has equal value and is called an *equipotential surface*. The irregular distribution of mass, especially near the Earth's surface, warps the geoid so that it is not identical to the ellipse of rotation (Figure 2.2). Long-wavelength anomalies, which can be mapped using data from satellites (Wagner *et al.* 1977), relate to very deep-seated masses in the mantle (Figure 2.2A), whereas density features at shallow depths cause shorter-wavelength warps in the geoid (Figure 2.2B). Consequently, anomalies within the gravitational field can be used to determine how mass is distributed. The particular study of the gravitational field and of the form of the Earth is called *geodesy* and is used to determine exact geographical locations and to measure precise distances over the Earth's surface (*geodetic surveying*).

2.2.2 Gravity units

The first measurement of the acceleration due to gravity was made by Galileo in a famous experiment in which he dropped objects from the top of the leaning tower of Pisa. The normal value of g at the Earth's surface is $980\,\mathrm{cm/s^2}$. In honour of Galileo, the c.g.s. unit of acceleration due to gravity ($1\,\mathrm{cm/s^2}$) is the *Gal*. Modern gravity meters (gravimeters) can measure extremely small variations in acceleration due to gravity, typically 1 part in 10^9 (equivalent to measuring the distance from the Earth to the Moon to within a metre). The sensitivity of modern instruments is about ten parts per million. Such small numbers have resulted in sub-units being used such as the milliGal ($1\,\mathrm{mGal} = 10^{-3}$ Gals) and the microGal ($1\,\mu\mathrm{Gal} = 10^{-6}$ Gals). Since the introduction of SI units, acceleration due to gravity is measured in $\mu\mathrm{m/s^2}$, which is rather cumbersome and so is referred to as the *gravity unit* (g.u.); 1 g.u. is equal to 0.1 mGal [10 g.u. = 1 mGal]. However, the gravity unit has not been universally accepted and 'mGal' and 'μGal' are still widely used.

2.2.3 Variation of gravity with latitude

The value of acceleration due to gravity varies over the surface of the Earth for a number of reasons, one of which is the Earth's shape. As the polar radius (6357 km) is 21 km shorter than the equatorial radius (6378 km) the points at the poles are closer to the Earth's centre of mass (so smaller value of R) and, therefore, the value of gravity at the poles is greater (by about 0.7%) than that at the equator (Figure 2.3) (see equation (3) under Box 2.2). Furthermore, as the Earth rotates once per sidereal day around its north–south axis, there is a centrifugal acceleration acting which is greatest where the rotational velocity is largest, namely at the equator (1674 km/h; 1047 miles/h) and decreases to zero at the poles (Figure 2.3). The centrifugal

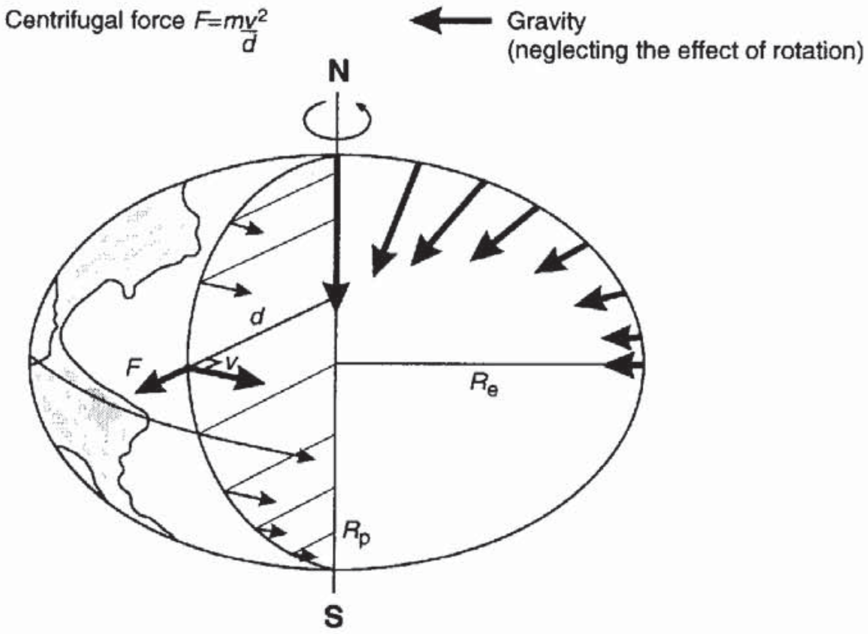

Figure 2.3 Centrifugal acceleration and the variation of gravity with latitude ϕ (not to scale)

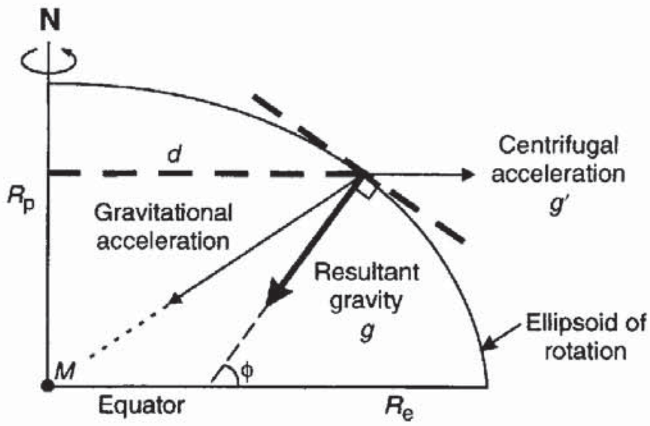

Figure 2.4 Resultant of centrifugal acceleration (g') and the acceleration due to gravity (g) (not to scale); the geographic (geodetic) latitude is given by ϕ. After Robinson and Coruh (1988)

acceleration, which is equal to the rotational velocity (ω) squared times the distance to the rotational axis (d), serves to decrease the value of the gravitational acceleration. It is exactly the same mechanism as that which keeps water in a bucket when it is being whirled in a vertical plane.

The value of gravity measured is the resultant of that acting in a line with the Earth's centre of mass with the centrifugal acceleration (Figure 2.4). The resultant acts at right-angles to the ellipsoid of rotation so that a plumb line, for example, hangs vertically at all locations at sea level. The angle ϕ in Figure 2.4 defines the geodetic (ordinary or geographic) latitude. The resultant gravity at the poles is 5186 mGal (51 860 g.u.) greater than at the equator and varies systematically with latitude in between, as deduced by Clairaut in 1743.

Subsequent calculations in the early twentieth century, based on Clairaut's theory, led to the development of a formula from which it was possible to calculate the theoretical acceleration due to gravity (g_ϕ) at a given geographic latitude (ϕ) relative to that at sea level (g_0). Parameters α and β are constants which depend on the amount of flattening of the spheroid and on the speed of rotation of the Earth.

Box 2.3 General form of the International Gravity Formula

$$g_\phi = g_0(1 + \alpha \sin^2 \phi - \beta \sin^2 2\phi)$$

In 1930 the International Union of Geodesy and Geophysics adopted the form of the *International Gravity Formula* (Nettleton 1971; p. 20) shown in Box 2.3. This became the standard for gravity work. However, refined calculations using more powerful computers and better values for Earth parameters resulted in a new formula – known as the *Geodetic Reference System 1967* (GRS67) – becoming the standard (Woollard 1975) (Box 2.4). If gravity surveys using the 1930 gravity formula are to be compared with those using the 1967 formula, then the third formula (Kearey and Brooks 1991) in Box 2.4 should be used to compensate for the differences between them. Otherwise, discrepancies due to the differences in the equations may be interpreted wrongly as being due to geological causes.

Box 2.4 Standard formulae for the theoretical value of g at a given latitude ϕ

$$g_\phi(1930) = 9.78049\,(1 + 0.0052884 \sin^2 \phi - 0.0000059 \sin^2 2\phi)\ \mathrm{m/s^2}$$

$$\boldsymbol{g_\phi(1967) = 9.78031846\,(1 + 0.005278895 \sin^2 \phi + 0.000023462 \sin^4 \phi)\ \mathrm{m/s^2}}$$

$$g_\phi(1967) - g_\phi(1930) = (-172 + 136 \sin^2\phi)\ \mu\mathrm{m/s^2}\ (\mathrm{g.u.})$$

2.2.4 Geological factors affecting density

Gravity surveying is sensitive to variations in rock density, so an appreciation of the factors that affect density will aid the interpretation of gravity data. Ranges of bulk densities for a selection of different material types are listed in Table 2.2 and shown graphically in Figure 2.5.

It should be emphasised that in gravity surveys, the determination of densities is based on rocks that are accessible either at the surface, where they may be weathered and/or dehydrated, or from boreholes, where they may have suffered from stress relaxation and be far more

Table 2.2 Densities of common geologic materials (data from Telford *et al.* 1990)

Material type	Density range (Mg/m^3)	Approximate average density (Mg/m^3)
Sedimentary rocks		
Alluvium	1.96–2.00	1.98
Clay	1.63–2.60	2.21
Gravel	1.70–2.40	2.00
Loess	1.40–1.93	1.64
Silt	1.80–2.20	1.93
Soil	1.20–2.40	1.92
Sand	1.70–2.30	2.00
Sandstone	1.61–2.76	2.35
Shale	1.77–3.20	2.40
Limestone	1.93–2.90	2.55
Dolomite	2.28–2.90	2.70
Chalk	1.53–2.60	2.01
Halite	2.10–2.60	2.22
Glacier ice	0.88–0.92	0.90
Igneous rocks		
Rhyolite	2.35–2.70	2.52
Granite	2.50–2.81	2.64
Andesite	2.40–2.80	2.61
Syenite	2.60–2.95	2.77
Basalt	2.70–3.30	2.99
Gabbro	2.70–3.50	3.03
Metamorphic rocks		
Schist	2.39–2.90	2.64
Gneiss	2.59–3.00	2.80
Phylite	2.68–2.80	2.74
Slate	2.70–2.90	2.79
Granulite	2.52–2.73	2.65
Amphibolite	2.90–3.04	2.96
Eclogite	3.20–3.54	3.37

cracked than when *in situ*. Consequently, errors in the determination of densities are among the most significant in gravity surveying. This should be borne in mind when interpreting gravity anomalies so as not to over-interpret the data and go beyond what is geologically reasonable.

There are several crude 'rules of thumb' which can be used as general guides (Dampney 1977; Telford *et al.* 1990; Nettleton 1971, 1976). Sedimentary rocks tend to be the least dense (average density about 2.1 ± 0.3 Mg/m^3). Within the three fundamental rock classifications there are crude trends and associations which are outlined in the next section. Commonly, units are quoted in terms of grams per cubic centimetre (g/cm^3) but are herein referred to in the SI derived units of Mg/m^3 which are numerically equivalent.

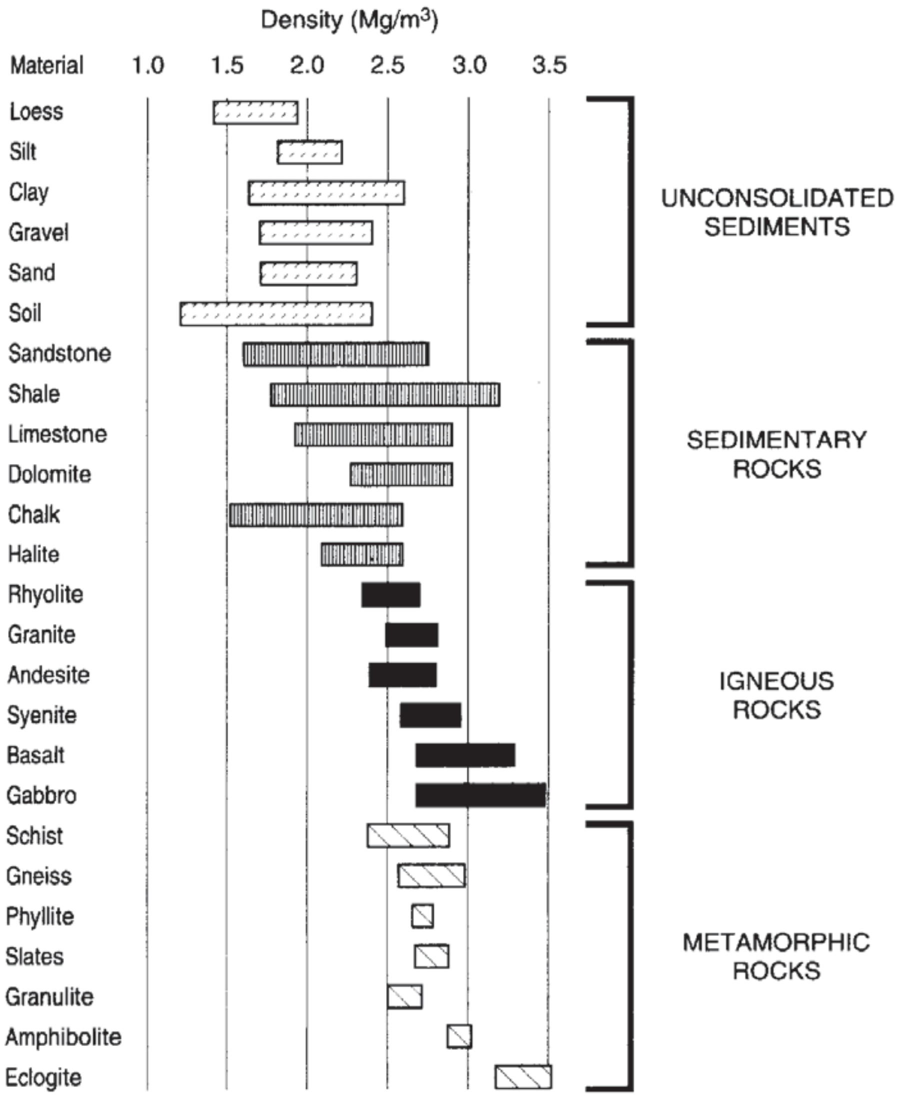

Figure 2.5 Variations in rock density for diffeent rock types. Data from Telford *et al.* (1990)

2.2.4.1 *Sedimentary rocks*

At least seven factors affect the density of sedimentary materials: composition, cementation, age and depth of burial, tectonic processes, porosity and pore-fluid type. Any or all of these may apply for a given rock mass. The degree to which each of these factors affects rock density is given in Table 2.3; but experience shows that, under normal circumstances, the density contrast between adjacent sedimentary strata is seldom greater than 0.25 Mg/m^3.

Density varies depending on the material of which the rock is made, and the degree of consolidation. Four groups of materials are listed in order of increasing density in Table 2.4. Sediments that remain buried for a long time consolidate and lithify, resulting in reduced porosity and consequently an increased density.

Table 2.3 The effect of different physical factors on density

Factor	Approximate percentage change in density
Composition	35
Cementation	10
Age and depth of burial	25
Tectonic processes	10
Porosity and pore fluids	10

Table 2.4 Approximate average densities of sedimentary rocks

Material type	Approximate average Density (Mg/m^3)
Soils and alluvium	2.0
Shales and clays	2.3
Sandstones and conglomerates	2.4
Limestone and dolomite	2.6

In sandstones and limestones, densification is achieved not by volume change but by pore spaces becoming infilled by natural cement. In shales and clays, the dominant process is that of compaction and, ultimately, recrystallisation into minerals with greater densities.

2.2.4.2 *Igneous rocks*

Igneous rocks tend to be denser than sedimentary rocks although there is overlap. Density increases with decreasing silica content, so basic igneous rocks are denser than acid ones. Similarly, plutonic rocks tend to be denser than their volcanic equivalents (see Table 2.5).

Table 2.5 Variation of density with silica content and crystal size for selected igneous rocks; density ranges and, in parentheses, average densities are given in Mg/m^3. Data from Telford *et al.* (1990)

Crystal size	Silica content		
	Acid	Intermediate	Basic
Fine-grained (volcanic)	Rhyolite 2.35–2.70 (2.52)	Andesite 2.4–2.8 (2.61)	Basalt 2.70–3.30 (2.99)
Coarse-grained (plutonic)	Granite 2.50–2.81 (2.64)	Syenite 2.60–2.95 (2.77)	Gabbro 2.70–3.50 (3.03)

2.2.4.3 *Metamorphic rocks*

The density of metamorphic rocks tends to increase with decreasing acidity and with increasing grade of metamorphism. For example, schists may have lower densities than their gneissose equivalents. However, variations in density within metamorphic rocks tend to be far more erratic than in either sedimentary or igneous rocks and can vary considerably over very short distances.

2.2.4.4 *Minerals and miscellaneous materials*

As the gravity survey method is dependent upon contrast in densities, it is appropriate to highlight some materials with some commercial

Table 2.6 Densities of a selection of metallic and non-metallic minerals and some miscellaneous materials. Data from Telford *et al.* (1990)

Material type	Density range (Mg/m^3)	Approximate average density (Mg/m^3)
Metallic minerals		
Oxides, carbonate		
Manganite	4.2–4.4	4.32
Chromite	4.2–4.6	4.36
Magnetite	4.9–5.2	5.12
Haematite	4.9–5.3	5.18
Cuprite	5.7–6.15	5.92
Cassiterite	6.8–7.1	6.92
Wolframite	7.1–7.5	7.32
Uraninite	8.0–9.97	9.17
Copper	n.d.	8.7
Silver	n.d.	10.5
Gold	15.6–19.4	17.0
Sulphides		
Malachite	3.9–4.03	4.0
Stannite	4.3–4.52	4.4
Pyrrhotite	4.5–4.8	4.65
Molybdenite	4.4–4.8	4.7
Pyrite	4.9–5.2	5.0
Cobaltite	5.8–6.3	6.1
Galena	7.4–7.6	7.5
Cinnabar	8.0–8.2	8.1
Non-metallic minerals		
Gypsum	2.2–2.6	2.35
Bauxite	2.3–2.55	2.45
Kaolinite	2.2–2.63	2.53
Baryte	4.3–4.7	4.47
Miscellaneous materials		
Snow	0.05–0.88	n.d.
Petroleum	0.6–0.9	n.d.
Lignite	1.1–1.25	1.19
Anthracite	1.34–1.8	1.50