I

INTRODUCTION

There have been many definitions of stratigraphy, differing slightly in scope and emphasis, but its core is the study of the earth's history, in so far as it can be interpreted from the outermost layers of the crust and particularly from the succession of stratified rocks. As a formal term it is a latecomer (appearing in 1865, according to the *Oxford English Dictionary*); but the description of stratified rocks had been placed on a scientific basis long before, in the late eighteenth and early nineteenth centuries – a phase of geological history that in this country will always be associated with William Smith. Since that time, when investigations began on the more accessible fossiliferous strata, with emphasis on the essential geological map, there have been far-reaching extensions in methods and scope. Even the last few decades have seen major developments.

Much more has been discovered about rocks at depth, both on land and beneath the sea, especially by geophysical methods and deep boreholes. Economic incentives are influential here and with problems of energy and material resources much to the fore these incentives are likely to continue. A range of structural and geochronological techniques has been applied to highly metamorphosed and deformed rocks, so that Precambrian history in particular is much better known; even the traces of Precambrian life are beginning to present a logical sequence. From a greater knowledge and analysis of modern sediments our interpretations of sedimentary rocks and their origins are on a more secure basis. With more international collaboration some standardization in stratigraphical nomenclature has been initiated. Finally the plate tectonic hypothesis has been an outstanding catalyst in promoting new syntheses, new collaborations between the various branches of earth sciences and the questioning of old assumptions. Not only have maps of the world's surface at various periods been radically altered but also the thinking of geologists in more subtle ways.

For the most part these developments are complementary to, and not at variance with, the older classic views of stratigraphy and earth history.

One major principle, however, has come into question, that of uniformitarianism. This is not, of course, a complete overthrow. The simplified version, that the present is key to the past, is still largely relevant to that part of earth history where it was first formulated – the Phanerozoic systems (i.e. Cambrian and later periods). It is the much greater understanding of Precambrian history, more than five-sixths of the total span, that has caused this revision. As outlined in the next chapter this enormous stretch of time saw fundamental changes in the composition of oceans and atmosphere, with concurrent changes in the chemistry of sediments and the slow evolution of the primitive forms of life. Even within Phanerozoic history there have been some unidirectional changes, most obvious in the biological sphere. For instance, the clothing of the land surface with an advanced rooted type of vegetation, which developed in late Silurian times, must have altered regimes of denudation, transport and deposition.

Climatic comparisons of past and present are less pronounced. In one sense the Quaternary glaciation is still with us, and the present time may well represent only the modest opening of an interglacial regime. In earlier periods when there were no polar ice-caps (possibly from the Trias to the Eocene) it is likely that the climatic belts were slightly different, which would influence the distribution of faunas and floras and their relations to latitude.

GEOCHRONOLOGY

Time, enormous stretches of time which are measured in millions of years, forms a background to most aspects of geology, particularly to historical geology and stratigraphy. In the last few decades the measurement of this basic dimension has become a major subject in its own right, geochronology, in which the age of a rock group or a geological event is measured in years.

These age determinations are based on the analysis of one or more pairs of isotopes (radioactive parent and daughter), the most commonly employed being given in Table 2. The first of each pair is transformed into the second at a known rate and from the relative amounts present an estimate can be made of the time since the mineral was formed, when it contained none of the daughter isotope. Analyses may be made on individual mineral grains collected from a rock outcrop or from whole rock samples. In modern work experimental errors are small, normally less than 5% and sometimes as low as 1%. However, there remain numerous uncertainties in the interpretation of isotopic ages, mainly because of the possibility that the daughter isotope may be lost through diffusion out of the mineral in which it was formed.

There are various aims in geochronology. One is to calibrate the Phanerozoic time-scale so that the boundaries of the systems and their major divisions can be expressed in years. Ideally the measurement would be made on a sedimentary rock that also contains reliable stratigraphical

Table 2. *Some isotopes used in age determination*

Method	Parent	Half-life	Daughter	Minerals
Uranium–lead	^{238}U	4 500	^{206}Pb (+ $8\,^4He$)	Zircon, uraninite and pitchblende
Uranium–lead	^{235}U	700	^{207}Pb (+ $7\,^4He$)	
Rubidium–strontium	^{87}Rb	50 000	^{87}Sr	Micas, potassium feldspars
Potassium–argon	^{40}K	1 300	^{40}Ar	Micas, sanidine, hornblende and glauconite

The half-life (approximately, in million years) is that in which the quantity of the parent isotope is reduced by half. The decay constant (a factor quoted in the Rb–Sr method) is the proportion of a certain number of atoms that decay in a certain time. The isochron Rb–Sr method depends on the ratios ^{87}Rb to ^{86}Sr and ^{87}Sr to ^{86}Sr in several rock or mineral samples and is valuable in deciphering the history of igneous and metamorphic complexes.

Individual age determinations quoted in later chapters have been recalculated where necessary in accordance with internationally agreed decay constants (Steiger and Jäger, 1977).

fossils but this is only rarely possible. There are few sedimentary minerals that are formed at or near the time of sedimentation and can be separated in a pure state for analysis; they include some clay minerals in shales but here the separation problems are formidable. Glauconite is more generally useful and is not uncommon in marine rocks, but loses argon easily, particularly if it has been deeply buried. Lavas and tuffs containing zircon, sanidine and micas may be very useful if interbedded with suitable sediments. In general, however, these direct methods have been most successfully applied to late Mesozoic and Tertiary rocks, for instance in North America, and not so much in this country. The use of intrusive rocks is more indirect. To be a reliable gauge of stratigraphic position the intrusion should cut fossiliferous strata of one age and be overlain by, or contribute recognizable detritus into, a later group with only a small time-gap between the two. In this country many Palaeozoic dates are based on intrusions, including Caledonian and Hercynian granites, with varying degrees of precision. The methods quoted so far apply primarily to regions little affected by later metamorphism or intrusion – that is, once the radioactive clock has been set the rock has remained a closed chemical system. In other situations, such as a major orogenic belt, the clock may be re-set by later tectonic or thermal events, sometimes millions of years later than the formation of the rock. There are two main causes for these late dates, or apparent ages. Deep-seated igneous or metamorphic complexes may cool very slowly before reaching the crucial 'blocking' temperature at which isotopic diffusion ceases and the system becomes closed. These late effects usually follow uplift and erosion, and have been called cooling ages. They show up as systematic variations in age related to structural level, or even altitude, with certain minerals always older than the others; for example, muscovite usually appears older than co-genetic biotite because its blocking temperature is higher. Then in regions of multiple orogenic phases late dates often reflect the last tectonic or thermal event which has extensively 'overprinted' the earlier intrusive or metamorphic phases. Both situations commonly necessitate the cautious ascription of minimum ages only, and both are well known from the history of the Caledonian orogeny in Britain. One of the virtues of whole rock methods is that they help to distinguish early effects from later overprinting.

A third, and probably the most influential, outcome of geochronology has been to establish a time-sequence and broad correlations within the great Precambrian shield areas of the world. Here isotopic data form the essential framework within which the tectonic and metamorphic sequence can be set; together with this sequence the data establish and define the major provinces or ancient fold belts of the shield. Although British Precambrian rocks are small in area they have been investigated by the same methods and year by year a greater understanding of their grouping and history has been gained.

A final isotopic method, on a very different scale, is based on the radioactive isotope of carbon, ^{14}C, and it is often called radiocarbon dating. Very small amounts of ^{14}C are produced in the upper atmosphere and as carbon dioxide it is incorporated in the tissues of plants and animals. While these are alive the proportion of ^{14}C is in equilibrium with that of the atmosphere, but after the organism dies no more is absorbed and the isotope is gradually lost, the half-life being 5730 years. Organic substances that originated as much as 35 000–40 000 years ago can be dated by radiocarbon analysis, and by special techniques the dating can sometimes be carried back to about 50 000 years. The method is thus valuable in late Quaternary stratigraphy and for dating archaeological objects.

SEDIMENTARY ROCKS, STRUCTURES AND ASSOCIATIONS

The raw materials of stratigraphy are rocks (sedimentary, volcanic, metamorphic), their structures and relationships and any fossils they may contain. Within the scope of this book the sedimentary rocks are the most voluminous, and some of their characters are reviewed here.

Sedimentary structures, cross-bedding and graded bedding

These two types of bedding have a double significance in stratigraphy; they are important as environmental indicators and may also be used as guides to the orientation of strata (Fig. 2). The foreset beds of cross-bedding normally dip in the direction of the current immediately responsible for them. In certain conditions, such as a meandering stream, these

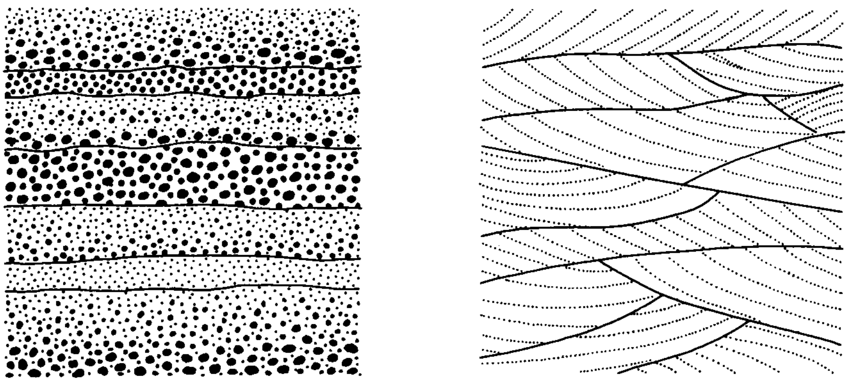

Fig. 2. Graded bedding and cross-bedding. On the left four beds are graded and show an upward decrease in grain size; on the right the majority of inclined foreset beds are concave upwards. (Adapted from several sources.)

directions may swing widely on either side of the valley trend as a whole, but in general when many readings are taken and analysed, in conjunction with larger structures such as channels, they constitute valuable data on the current directions of, say, a delta or a river system. In graded beds there is a gradual upward decrease in grain size, for instance from coarse sand at the base to silt or even mudstone at the top.

Graded bedding, from greywackes and quartzites in particular, has been used to determine the orientation (i.e. the original top and bottom) of formations in many folded belts of the world, including the Scottish Highlands, Southern Uplands and north-west Ireland. The use of cross-bedding depends on the truncation of the foreset beds. In certain types these tend to curve tangentially into the normal bedding planes at the base but to be truncated at the top, having been eroded by the next influx of sediment. Nevertheless if such structures are examined in a succession whose superposition is indisputable they commonly include some indeterminate examples. It is therefore imprudent to rely on isolated items of either type of bedding, but when repeated many times and combined with structural evidence they are valuable stratigraphical guides.

Cross-bedding may develop in many situations, given the necessary current speed and sediment supply, the latter being usually of sand grade. It tends to characterize the shallow waters of shores, deltas or rivers. Aeolian cross-bedding may occur in rather large units and sometimes with slightly steeper foresets. A variety of ripple marks is formed by currents in shallow water but there are also deep water variants and an aeolian type. Rootlet beds result from plant growth both above and below water-level and abundant large desiccation cracks are one of the clearest indications of temporary emergence. Intertidal and subtidal muds and silts are often reworked by many types of marine organism (bioturbation), leaving tracks and burrows or largely destroying the original bedding.

Thus a combination of several structures may suggest sedimentation in shallow water. 'Deep water' is a more problematic ascription. Turbidites, resulting from sediment-charged gravity flows, are known to accumulate at the foot of the continental slope while fine pelagic muds spread widely on the more distant parts of the sea floor. Nevertheless some turbidites have been found in lakes or at relatively shallow depths. Among ancient environments deep waters are probably most securely deduced from a combination of sedimentary features and a position at the edge of a continental plate margin.

Disturbed bedding, slumps and gravity slides

In most rocks that are unaffected by folding the bedding planes are horizontal or nearly so, excluding such obvious cases as current-formed foresets or accumulations on the flanks of a reef. Occasionally however, a normally bedded sequence contains units which are strongly convoluted, or altogether disrupted, resulting in balled-up pillow-like structures. These are slumps, or slump units, and are most common in sandstones though occasionally found in siltstones and carbonate rocks. Some of the less disrupted types result from the curving over of the tops of foreset beds. It seems that most slumping took place on a subaqueous slope while the sediment still held enough water to be unstable, and the dislodged mass crumpled up as it slid. The slope, however, need not have been more than a few degrees and depth of water does not appear to be crucial. Some kind of triggering mechanism seems likely and earthquakes have been suggested. Slumping in this sense is largely a sedimentary structure, but tectonically induced gravity sliding may occur on a larger scale, as is evident in the Alpine chain. Here there are large transported blocks (making up mélanges or wildflysch) or great rafts of strata (olistostromes). A modest British example is mentioned on page 35.

Unconformities and other breaks in sedimentation

The most obvious result of a period of uplift and erosion is the major unconformity – an angular unconformity where there is a discordance of dip above and below, and a disconformity without that discordance. There may also be smaller breaks, variously known as diastems, lacunae or

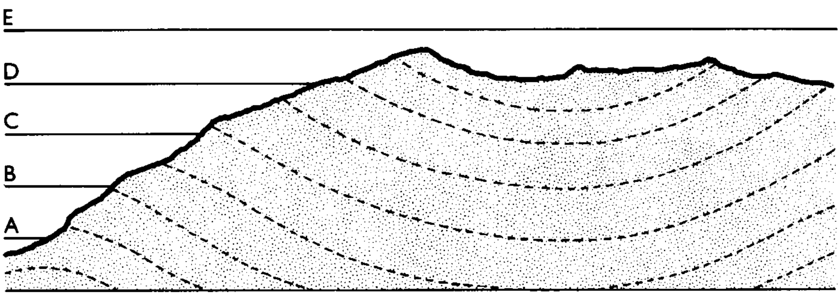

Fig. 3. Overlap and overstep. A sequence of beds, A–E, is laid down against an older landmass. Beds A–C show overlap against that landmass and the successive positions of their shores can be determined. Bed D, however, oversteps several of the older gently folded strata, and the shore positions for upper D and bed E cannot be determined.

hiatuses, in which evidence of erosion is inconspicuous or may only result from a prolonged dispersal of sediment. A non-sequence is a gap that is only detectable by palaeontological means, usually the absence of a zone or zones. It has also been suggested that sedimentation itself is a somewhat spasmodic process, and that the common bedding plane represents a lull in sedimentary influx.

Unconformity with overlap (Fig. 3) results from uplift and erosion, followed by a marine transgression farther and farther over the submergent land surface. Overlap is valuable in palaeogeographic reconstruction because it presents evidence of this retreating shoreline. No such inference can be drawn from unconformity with overstep, where a single formation oversteps or transgresses more than one underlying one, because there is no indication of how much farther the sea extended before the shoreline was reached. In both situations later folding or faulting may complicate the final structure. As a result the distinction between overlap and overstep may be debatable, and the palaeogeography similarly uncertain. An example affecting Ordovician geography is given on page 91.

Cyclic sedimentation

This refers to the vertical repetition, several times over, of a few sedimentary types, each cyclic unit being known as a cyclothem. An 'ideal cyclothem' may be either of the following:

d	coal		*a*	limestone	
c	sandstone	cyclothem	*b*	sandstone	
b	shale		*c*	shale	
a	limestone		*d*	coal	cyclothem
d	coal		*c*	shale	
c	sandstone	cyclothem	*b*	sandstone	
b	shale		*a*	limestone	
a	limestone				

Cyclic sedimentation is most conspicuous among various types of shallow water rocks, some of the best examples resulting from fluctuations between the sharply distinct environments just above and below sea-level – such as the simplified versions set out above. However other cycles have been recorded from evaporite, turbidite and fluviatile associations and overall there is much variation in scale and complexity. In practice the ideal type is nearly always complicated by minor omissions or small-scale oscillations.

The causes of cyclothems have been much debated and probably there are several. Some seem to reflect largely sedimentary events within the basin or catchment area, such as changes in river position or regime. For others a tectonic cause has been invoked or, particularly in recent years, a eustatic one. Eustatic changes are strictly those of ocean-level (not uplift or depression of the land) and the clearest mechanism is glacio-eustatic, or the withdrawal of water during major glaciations to form ice-caps. Outside periods of known glaciation – and several cyclic sequences belong to such periods – the most plausible cause is variation in ocean-floor topography related to changes in the rate of sea-floor spreading, so that with an enhanced rate there is overspill of oceanic water onto the edges of the continents. This view is supported by some degree of correlation between spreading rates and transgressions or regressions in late Cretaceous and Tertiary times. Unfortunately such comparisons cannot be directly sought in earlier systems, because no sea floor earlier than the Jurassic survives.

Stratal thickness, isopachs and rates of deposition

Thickness of strata is a basic component in stratigraphic syntheses, linked with lateral variations in thickness and changes in sedimentary type. Sedimentation is commonly accompanied by subsidence, and if there is evidence throughout that the formation was deposited near sea-level, then the thickness becomes also a measure of the subsidence. It is often convenient to summarize changes in thickness in the form of an isopach map (cf. Figs. 46, 67); the isopachs, being lines denoting equal thickness, form a system of contours based on points where the total thickness of the unit can be measured.

Thickness measurements normally refer to the rock unit as it exists at present, but there may have been changes during its history, for various reasons. Compaction is a diagenetic process affecting clay rocks. In their earliest stages these contain a large amount of water, which is gradually expelled by the weight of superincumbent layers so that the thickness is much reduced, perhaps to a quarter or less. Tectonic stress may also alter rock dimensions, either by lateral extension, with consequent thinning, or by compression to give 'tectonic thickening'. It has been estimated that in parts of North Wales the Lower Palaeozoic rocks have been about doubled in thickness owing to deformation during the Caledonian orogeny.

In favourable circumstances rock thickness may give an indication of subsidence but it is far more difficult to relate it to duration or the time taken to deposit a certain number of metres. Rates of sedimentation are elusive and the best attempts at measurement are either on a very small scale or a very large one. Certain laminated beds are formed by annual variations in the sedimentary regime, the best known being the periglacial varves; these may be counted in years. Then since isotopic determinations supply the duration of several of the later systems and of their major subdivisions, it is possible to deduce, for instance, that 300 metres of Lower Jurassic shales were laid down in about 10 million years. But to extrapolate within this span and assume that a certain 3 metres represent

100 000 years is hazardous, in that it presupposes a regularity of deposition that is hard to justify. Nevertheless, with some assistance from an exceptionally detailed zonal scheme, it has been suggested that a typical marine band of Namurian age (Upper Carboniferous), six to nine centimetres thick, represents no more than 10 000 to 12 000 years. This would be a very unusual degree of precision and probably should be taken only to indicate the *order* of time involved.

Lateral variation, facies and environments

In the same way that there are variations in thickness there are lateral changes in rock types with their associated structures and fossils. Such associations introduce the concept of facies – one essential to stratigraphy but not easy to define concisely. 'Facies' was first used in 1838 in a description of the Swiss Jura mountains, more or less in its modern sense, to describe lateral changes in lithology within a stratigraphical unit. Bound up with this is the concept of analogous lateral changes in the environments that gave rise to the sediments and which have their counterparts at the present day.

It is thus postulated that during a certain period the conditions of deposition varied from place to place – sand banks and channels here, mud-flats there, and shell banks a little way offshore; or perhaps a mangrove swamp, a lagoon, a barrier reef and a steep outer slope to the ocean floor. Each of these environments produces different bottom sediments, characterized by different faunal and floral assemblages; when 'fossilized' they become different facies – sandy, muddy, or calcareous, all of the same age and passing laterally into one another. The names given to the facies may pick out one aspect of the whole – the lithology (carbonate facies), fauna (shelly and graptolitic facies), tectonic or regional setting (trough and shelf facies), environment (lagoon facies). The lithological version is the most factual, universal and probably the most satisfactory, while the last two contain a greater element of deduction, but all have been used.

A detailed analysis of facies is an essential component in the assessment of past environments, and in favourable circumstances these may be set in a larger frame, in relation to contemporary seas, lands and mountain ranges, climatic belts and latitude, continental and oceanic margins, island arcs and vulcanism. A review of past and present environments shows that the major types are relatively few, though naturally with enormous