PALAEOLATITUDINAL CONTROLS OF PHANEROZOIC SEDIMENT-HOSTED MINERAL DEPOSITS

by

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I declare that the contents of this thesis have not been previously presented for a degree at this or any other University.

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ABSTRACT

The distribution of Phanerozoic sediment-hosted mineral deposits appears to be influenced by latitudinal zoning. The palaeolatitudes of the host rocks were determined using standard palaeomagnetic procedures - the most reliable results being for the Mesozoic and Cenozoic with early Palaeozoic palaeolatitudes least reliable. The palaeolatitudes derived from Tarling and BP palaeogeographic reconstructions are in general agreement i.e. +/- 10°, although greater discrepancies occur for India and Central America. It is shown that some types (e.g. sandstone copper, sandstone lead) have a preference for low latitude arid regions whilst conditions in the equatorial and temperate rainfall belts were more favourable to the formation of other deposits (e.g. sandstone uranium-vanadium, oolitic ironstone). Using climatic modelling assuming uniformitarianism of the principles governing the Earth's atmospheric and oceanic circulation patterns, the climatic conditions affecting the distributions of sediment-hosted deposits were evaluated. It is concluded that local climatic effects are influential in the genesis of limestone base-metal, oolitic ironstone, sandstone copper, sandstone lead, shale base-metal, sedimentary exhalative, sandstone uranium-vanadium, manganese, laterite and phosphate deposits. These climatic conditions affect the nature and degree of chemical weathering, erosion, abundance of organic matter, ground water chemistry and volume in a particular region. However in some instances, such as placer deposits, the major control on deposit distribution was the availability and distribution of source rocks. Such palaeolatitudinal/palaeoclimatic control on the distribution of some deposit types places genetic constraints upon their formation. It also has obvious implications in the evaluation of potential sites for exploration and development.

CONTENTS

CHAPTER ONE

INTRODUCTION

A number of Phanerozoic sediment-hosted mineral deposit types have a close spatial and temporal association with certain types of so-called climate-sensitive lithologies (e.g. evaporites and red beds). An examination of the present and palaeo-distributions of such climate-sensitive lithologies has suggested that they show a strong preference for certain latitudinal belts, presumably a response to the climatic conditions prevailing during deposition. It follows that there may also be a palaeolatitudinal control upon the distributions of some sediment-hosted mineral deposit types. Such a hypothesis will be tested here with the view to limiting the various theories as to their genesis.

The hypothesis described above has been proposed by numerous authors but it has not yet been fully tested and justified. Strakhov (1970) related the origin and distribution of copper-lead-zinc deposits to arid regions. Renfro (1974) suggested that sabkha-type environments produce conditions conducive to the formation of some mineral deposit types (e.g. sandstone coppers) which obviously indicated a strong climatic influence. This sabkha model has been taken up extensively by other workers e.g. Smith (1976) in relation to North Texas copper deposits and Rawson (1976) for uranium exploration. In addition the importance of a particular climatic regime (warm, tropical to sub-tropical) as one control in the formation of some primary mineral deposits has been proposed by numerous workers e.g. Wopfner and Schwarzbach (1976); Van de Poll (1978); Heckel and Witzke (1979); Cronin, Cannon and Poore (1983).

Sediment-hosted mineral deposit types were selected for examination as the host rocks for the mineralisation may also show similar palaeolatitudinal controls upon their formation as those described above. Frequently the host rocks are massive limestones, channel sandstones with abundant organic debris (e.g. fossil log jams) or evaporites thought to have been formed in coastal sabkha environments. The processes involved in the formation of sediment-hosted deposits listed below are also thought to be the result of latitudinal position, at least in part. At the Earth's surface the action of both the atmosphere and the hydrosphere alters minerals and forms new ones that are more stable under the existing conditions. Sediment-hosted deposits are derived from pre-existing material by processes of weathering, erosion, transportation, deposition, diagenesis and consolidation. A different source of materials and variations in the processes of formation may yield different types of deposit. Also the process of sedimentation itself may involve the concentration of metals into mineral deposits. The relative importance of climatic control is therefore relevant to the evaluation of ore genesis models.

The relative economic importance of deposits was not used as a constraint on data selection. Such a factor would have introduced a political/economic bias that may have influenced the results. It would also have been very difficult to assess the economic worth of deposits as factors which make them economic in one instance may make them uneconomic in another. The research has been confined to deposits of Phanerozoic age as the palaeomagnetic data which provide the basis of this work is extremely unreliable for the Proterozoic era. Although Palaeozoic palaeomagnetic data are also questionable it was decided that the level of reliability was sufficiently high for tentative conclusions to be drawn.

The determination of palaeolatitudes involves the use of standard palaeomagnetic studies based upon the permanent magnetisations in rocks (Chapter Two). The remanent magnetisation preserves the direction of the geomagnetic field at the time magnetisation was introduced into the rock. From this the palaeolatitude can be calculated.

It has already been mentioned that there is a connection between latitude and climate. The validity of making such a connection will be examined in detail (Chapter Four). There are many basic assumptions concerning the Earth's magnetic field, its dynamics and its patterns of circulation which must be made before such conclusions can be drawn. Despite these problems if a latitudinal control upon the formation of some mineral deposit types can be shown to exist, then it may be possible to explain anomalous occurrences by local variations in the predicted climatic conditions that prevailed during the formation of that deposit.

One of the consequences of this project will be the assessment of whether or not the classification of deposits can be elucidated by their apparent palaeolatitudes of formation. If it is not possible to classify deposits in such a way then their genesis cannot be related to palaeoclimatic factors and this thesis will be independent of the debate about the origins of these deposits so the classification will not actually matter. However if different groups of deposits show different palaeolatitudinal distributions, or if no palaeolatitudinal control is observed, then it will have implications for the genetic debate. Of course it is not possible to predict the outcome of the work so the classification of mineral deposits is a fundamental part of the thesis and will be developed with the exclusion of genetic constraints.

The palaeolatitudinal distributions of two varieties of volcanogenic mineral deposits (porphyry copper and epithermal gold) were determined in an effort to ascertain whether there was a bias inherent in the methods so that all mineral deposits show a latitudinal control upon their formation, regardless of origin. These deposit types were chosen as test subjects as there is no obvious reason why their distribution should be climatically controlled. The results for these volcanogenic deposits may test the reliability of the data collection and processing as the same methods were used as for the sediment-hosted deposits.

In conclusion the main purpose of this research is to determine whether a relationship exists between latitude, climate and the distribution of Phanerozoic sediment-hosted mineral deposit types. To reiterate: latitude has an effect upon climate (influencing both the atmospheric and the oceanic circulation patterns). It has also been shown by several workers that there is a latitudinal control upon the distribution of certain sediments. From the association between latitude and climate it follows that the distribution (and therefore formation?) of these sediments must be climatically dependent. There is an association of some mineral deposit types (e.g. sandstone copper, sedimentary exhalative, sandstone uranium-vanadium) with these so-called climate-sensitive lithologies so the distribution of these mineral deposits may also be climatically controlled. The testing of this hypothesis forms the basis of this research.

This work aims to determine whether any Phanerozoic sediment-hosted mineral deposit types lay within specific latitudinal zones. If such a latitudinal control is shown to exist, possible reasons for this will be proffered. It will also be noted if there is a significant shift in the distributions with time which may represent variations in global climatic conditions.

CHAPTER TWO

PALAEOMAGNETISM AND PALAEOMAGNETIC METHODS

The permanent magnetisation of igneous, metamorphic and sedimentary rocks has enabled inferences to be made concerning changes in the direction and intensity of the geomagnetic field in past geological times and also the movements of the continents relative to the poles. The basic principles of palaeomagnetism and palaeomagnetic methods are well known and documented in several textbooks (Irving, 1964; McElhinny, 1973; Tarling, 1983). A brief outline is given here to assist understanding of the methods and discussions to follow.

2.1 Naturally Occurring Types of Remanent Magnetisation.

The natural remanent magnetisation of rocks (NRM) is due to their content of accessory ferromagnetic minerals, as much as a few percent in certain igneous rocks but often less than half of one percent in many sediments. Two groups of ferromagnetic minerals are especially important, the anhydrous ferric oxides magnetite and haematite and their mutual solid solutions with titanium: that is the magnetite (Fe_3O_4)-ulvospinel (Fe_2TiO_4) and the haematite (Fe_2O_3)-ilmenite ($FeTiO_3$) solid solution series. It has been demonstrated experimentally that on cooling from high temperatures, an igneous rock acquires a magnetic moment, termed thermoremanent magnetisation (TRM), which is normally aligned in the direction of the ambient field. The critical temperature at which this moment begins to be acquired is known as the Curie temperature and varies according to the composition of the ferromagnetic minerals involved. Therefore, the TRM of igneous rocks (e.g. lavas, sills and dykes) is due to the magnetisation of ferromagnetic minerals acquired during cooling in the geomagnetic field from the temperatures at and below their Curie temperatures.

Depositional or detrital remanent magnetisation (DRM) is acquired by the physical rotation of magnetic particles during deposition as part of a sediment. In addition gravitational and dynamic forces also operate on all detrital particles and may cause deviation away from the ambient field direction.

Chemical remanent magnetisation (CRM) is acquired as a magnetic mineral nucleates and grows in a magnetic field. The acquisition of a remanence by means of crystal growth at constant temperature is exactly analogous to the acquisition of a thermoremanence. The nature of the chemical remanence will be affected by the temperature at which the chemical growth takes place and the duration for which the field is applied over any given volume range. A well-known and widespread occurrence of CRM is in the diagenetic growth of haematite grains in red sandstones.

The real surprise is not that rocks acquire a primary magnetisation when originally forming, but that they are able to retain it over tens, hundreds or even thousands of millions of years. The basic reason for this is that many rocks contain very small magnetic grains with very high coercive forces (the reverse magnetic fields needed to demagnetise the material) - high enough to protect the original magnetic directions against many of the vagaries of subsequent geologic history. Rock magnetisations may be very weak but they can also often be very stable and hence may divulge information on the direction (and in fewer instances the strength) of the Earth's magnetic field in the past. Geological tests have been devised to determine whether or not these original magnetisations had been altered. One of the main stages in the demagnetisation procedure is the assessment of the stability of the specific components of remanence which have been identified. Various stability indices (e.g. Tarling and Symons, 1967; Briden, 1972; Stupavsky and Symons, 1978) were amongst the earliest statistical methods developed for distinguishing specific vectors and establishing their stability.

These primary magnetisations generally decay very slowly, depending on the composition and grain sizes of their magnetic minerals. Over geological time the magnetic particles within a rock also gradually acquire new magnetisations in the direction of the prevailing geomagnetic field. In general, such secondary viscous magnetisations are mostly directed along the present geomagnetic field and can be effectively removed by partial demagnetisation through heating or placing samples in alternating magnetic fields. However a major problem arises when secondary magnetisations are due to prolonged CRM. The process of diagenesis (during which most consolidated sediments acquired a magnetic remanence) is variable, even for similar rock types. For example in red sandstones it can apparently occur rapidly after deposition or it may be delayed for 100 million years after deposition. Also, the passage of groundwaters through permeable rocks may lead to the continual re-setting of CRM as magnetic minerals are removed and new ones are crystallized. So studies of remanence of sediments, particularly consolidated ones, must be combined with a petrological investigation of the relationships of the magnetic minerals present. Even then the interpretation of the time of acquisition of the remanence may be unclear.

2.2 Separation and Identification of Remanent Components.

In order to dissect the natural remanence (NRM) of a sample into primary and secondary components partial demagnetisation techniques must be used. These methods require precise controls that are difficult to obtain in the field, so it is usual practice to collect samples and to undertake measurements of their susceptibility and remanence in the laboratory. The remanence must be examined for its stability and the possible age of its components, thus defining the directions and intensity of the geomagnetic field for some specific time, usually that at which the rock formed.

Palaeomagnetic analyses involve dissecting the NRM - the total of all magnetisations carried by a rock - into its components and establishing their sequence of acquisition. Standard methods of dissecting the natural remanence include partial incremental (stepwise) demagnetisation by heating (TH) or in alternating magnetic fields (AF). The precision with which the components are defined will obviously vary according to their magnitude relative to all other components and the noise level of the demagnetisation procedures. It is generally expected that the directional definition within any one specimen will usually be within a few degrees, about 2-3° (Tarling, 1983).

Once the NRM has been analysed, the next problem is to obtain a relative or absolute age for each magnetic component. This may become increasingly more difficult with increasing geological age as the NRM may become more complex. The stable magnetic component of remanence may now be defined, but it is still necessary to assess the likely time at which it was acquired. None of the tests (section 2.4,iv) can be considered conclusive on their own. Most are negative tests in the sense they may exclude certain components which do not fulfill the requirements. Components that do fit them may not automatically be considered primary.

2.3 Definition of a Palaeomagnetic Pole.

The remanent magnetisation in a rock is described by the basic parameters of direction and intensity. The magnetic vectors include the declination with respect to true north and the inclination from the horizontal. The average vectors in rocks of any specific age can be used to calculate the position of their corresponding palaeomagnetic pole on the standard geocentric axial dipole model for the average geomagnetic field. Although this model is itself derived from palaeomagnetic data it has been accepted as axiomatic. (This is discussed in greater detail in Chapter Four, section two). All rocks magnetised at the same time should have the same palaeomagnetic pole position if geomagnetic secular variations have been averaged out.

The palaeomagnetic pole is calculated from the direction (irrespective of polarity) and geographical co-ordinates using formulae which can be found in standard text books such as McElhinny (1973) and Tarling (1983). As the palaeomagnetic and palaeogeographic poles essentially correspond, the palaeolatitude can be derived by measuring the angular distance between the site and the palaeomagnetic pole. However because the geocentric axial dipole field is rotationally symmetrical about the palaeogeographic pole, the absolute palaeolongitude is indeterminate. Thus from the basic remanent magnetisation parameters of direction and intensity, the site palaeopole and corresponding palaeolatitude are derived. The poles are then evaluated according to the criteria outlined in the following section.

It is a generally accepted hypothesis that the tectonic plates comprising the Earth's surface move relative to each other. Thus although the Earth's axis of rotation remains fixed in space, the pole position appears to change relatively to each tectonic block through time. Successive determinations of the positions of the poles relative to any one block therefore lie on a curve - referred to as the apparent polar wandering path (APW). Each block will have a unique APW path which represents its movement relative to the Earth's axis of rotation. The change in position of the block through a given time period can be defined by a single angle with a single axis, the axis being specified in terms of latitude and longitude co-ordinates of its pole. This is in accordance with Euler's Theorem which states that the motion of a tectonic plate can be represented by rotation about an axis through the Earth's centre and a point on the surface.

The palaeomagnetic data are sparse and unevenly distributed in space and time, leaving doubt over the positions of the major plates for some time periods, especially those of Mesozoic age and older. Where palaeomagnetic data are sparse for a continent for a particular time period, positions are necessarily inferred through interpolation between older and younger data and are thus relatively approximate. There are a number of geological criteria which must be adhered to when such continental distribution and orientation is determined. These include;

i) the correlation of faunal provinces,

ii) the identification of accretionary flysch wedges to determine continental outlines as accurately as possible for a specific age,

iii) care is taken to prevent overlap of the plate in question with others,

iv) the correlation of climatic belt patterns and tectonic trends.

Examples of the application of these constraints are found in Ziegler et al (1977 and 1979).

2.4 Selection of Palaeomagnetic Data.

The problems of selection are outlined in Tarling (1985b) and are only briefly summarized here. The selection of palaeomagnetic data is critical to the polar wander paths produced for a tectonic block, so certain criteria are used in an effort to minimize the degree of subjectivity involved in the evaluation of the data. The following eight points constitute the criteria which palaeomagnetic data would ideally fulfill.

(i). The magnetisation isolated comprises only one component of remanence.

The presence of one component is accepted when an identical direction is recognised on two or more successive increments of the partial demagnetisation process. A certain degree of caution must still be exercised as even this single vector could still be a summation of vectors.

(ii). The magnetisation is stable to thermal (TH) and/or alternating field (AF) demagnetisation.

The component of remanence defined above must have a coercivity or temperature spectrum which corresponds to a relaxation time at least comparable to the geological age of the rock. This is the length of time each element (domain) within a grain takes to acquire a magnetisation with a component in the direction of the external field. In other words the magnetisation must be capable of having retained a record of the geological field since the rock was formed. The growth of authigenic, secondary minerals (e.g. haematite or magnetite) long after rock formation may produce a chemical remanence at the time of the exsolution or oxidation. This could possess an even greater stability than that in the pre-existing magnetic minerals and hence produce inaccurate results.

(iii). The samples are magnetically isotropic and homogeneous, that is, the observed vectors are true reflections of the geomagnetic field in which they were acquired.

The occurrence of anisotropy in rocks with a natural thermal remanence i.e. igneous and some metamorphic rocks, can be determined by giving the specimen a laboratory thermal remanence in a known field direction. The acquired direction is then compared with the applied field direction. Unconsolidated sediments may possess a magnetic anisotropy associated with the alignment of larger grains during deposition, although the stable remanence may well be carried by the smaller, interstitial grains.

Consolidated sediments may undergo major chemical changes during lithification which could erase all pre-existing depositional features and then they may acquire a chemical remanence relating to the time of lithification. It is thus difficult to ascertain the magnitude of any anisotropic corrections which should be applied to the remanent directions in sediments.

Inhomogeneity (i.e. uneven magnetisation) can occur on many levels from the atomic scale e.g. as impurities and defects in magnetic grains to uneven distribution of such magnetic grains in rocks. The effects of the inhomogeneities depend on their scale. If this is small relative to the size of individual samples, then these are generally reduced by spinning during measurement. However the measurement 'error' will be greatly increased if the inhomogeneity is on a scale comparable to the sample size.

(iv). The age of remanence is known.

It is necessary to attempt to date the time of acquisition of the different components for any interpretation of past field strengths and directions. Recent viscous magnetisations are easily identified as they have directions similar to the present geomagnetic field and they generally have lower magnetic stability to demagnetisation. The high temperature and high coercivity components are generally considered the most likely to represent the oldest geomagnetic field direction in the rock but this is only true if haematite was not produced at some later time. Haematite normally has both high coercivities and blocking temperatures, if of common grain size dimensions.

The most effective tests to determine the age of magnetisation are local consistency in fold, tilt, contact and conglomerate tests (Graham, 1949) in which magnetisations of samples taken from rocks in different attitudes are compared. If the components are of the same inclination, relative to bedding, then this suggests that the components were acquired before the rocks were disturbed. The fold/tilt test can only be carried out where identifiable tectonic deformation has occurred and it is essential that the magnetisations have not been affected by the actual tectonic processes themselves. In all of these tests it is important to establish that the consistent directions do not correspond with a known, but younger, direction of remagnetisation, such as may be caused by extensive penetrative oxidation or widespread burial without significant tectonic distortion.

(v). When sampling an adequate number of samples must be taken from a wide time range to reduce, by averaging, orientation and measurement errors and geomagnetic secular variations. Aberrant/anomalous samples must also be discarded.

It is essential to sample sufficient sites to average out all past changes in the geomagnetic field with time scales of less than 10^5 years. Examination of the present geomagnetic field (assuming this field is typical) indicates that normal secular variation amplitudes of the field cause a scatter which may exceed 20°. In order to reduce this scatter to an acceptable value the sampling must extend over several ten thousand years.

In consolidated sediments adequate sampling over a wide time range is relatively easily achieved as chemical changes during diagenesis are likely to be protracted, even within a single sample. In contrast, lava and dyke complexes generally acquire their remanence during cooling over short periods of time (maybe a few tens of years) and so will often provide only 'spot' readings of the geomagnetic field. Such rock types may require extensive sampling of a large number of different flows before the secular variations can be averaged out and a true determination of the average dipolar field achieved. However some lavas also acquire a chemical remanence over a longer period if deuteritic exsolution is prolonged. These can then be treated similarly to consolidated sediments.

(vi). The sites being combined are located in autochthonous areas if the tectonic interpretation is to be extended over a wide region.

Even if the observed remanence can be related directly to the time of formation of the rock itself, the results may only apply over a limited area. The basement rocks of the sites sampled must not have moved relative to each other. If they have moved their precise movements must be known. This problem is illustrated in a nappe where the remanent directions may be resolved to show the motion of the nappe whilst the tectonic movements of the basement rocks remain obscured.

(vii). Any tectonic change, since the acquisition of any magnetisation of known age, is known and can be corrected for.

This normally means that the bedding plane tilt of any shales closely associated with the rocks being studied has been determined and that the mechanism of tilting is known. If tilting is in the form of a single rotation about a horizontal axis, then the tilt correction is simple. However, it is often difficult to determine whether or not rotations about a vertical axis have also occurred and further complications arise if internal deformations (plastic or rigid) have taken place during deformation.

(viii). Samples which become magnetised during polarity transitions or at other times of non-dipole geomagnetic behaviour have been detected and eliminated.

The component isolated must have been originally present in the rock, prior to laboratory analysis. There are three main areas which may influence this;

a) For example, the use of mu-metal screening around magnetometers often results in a concentration of magnetic fields near the magnetometer entrance.

b) In thermal demagnetisation methods any stray magnetic fields during cooling must be sufficiently small (<3nT) that they have no detectable affect on the natural remanence. If new magnetic minerals are also formed as a result of the heating, then the presence of any weak fields may influence the measured results even further.

c) Alternating magnetic field demagnetisation methods are prone to produce a variety of laboratory induced components, such as anhysteretic, gyro and rotational remanences. However, they are free of the chemical changes affecting the thermal demagnetisation techniques.

In conclusion, if the samples are collected as described, properly tested and all the precautions listed are taken, then the palaeopole and palaeolatitude for a given site can be reliably determined and the apparent polar wandering path for a particular continent deduced. Within this study the site palaeolatitude is of major concern as it largely dominates the climate, sedimentation and erosion in a region and hence may affect the distribution of sediment-hosted mineral deposits. Therefore the fact that any errors in the reported palaeomagnetic studies will result in errors in the palaeolatitude determinations for that locality must be emphasized.

CHAPTER THREE

CLASSIFICATION OF MINERAL DEPOSITS

3.1 Introduction

To be useful a classification of mineral deposits must be as simple as possible and, perhaps most importantly, usable in the field. However, different types of deposits grade into each other, thus no classification can be either complete or inflexible. In the past the vogue, especially among American geologists, was to classify all deposits in terms of a magmatic solution origin (for example, Lindgren, 1933) as mentioned by Jacobsen (1975) with reference to copper deposits. When authors such as Garlick (1961) suggested syngenetic processes of ore formation it marked a significant change in thinking. This change in attitude was reflected in the new classifications that emerged which were based upon these `new' ideas of mineral deposit formation. In contrast the Europeans had long thought of the "bedded ores" (e.g. Falun, Rammelsberg, Bleiberg) as syngenetic.

The classification of mineral deposits is still a matter of great discussion. There is a considerable literature which either encompasses all mineral deposit types, for example, Wolf (1981) and Gustafson and Williams (1981), or specific metallic mineral deposit types, as in Dahlkamp (1978) concerning uranium deposits; Jacobsen (1975) with reference solely to copper deposits.

The general classification of mineral deposits used here (Table 3.1) includes both metalliferous and non-metalliferous groups but does not involve interpretative genetic constraints. A classification based upon the genesis of mineral deposits has to be evaluated with great care as discrimination between the interpreted ore-forming processes and the characteristics of the ore type itself is vital. Nonetheless some distinction is needed between primary and secondary ores is necessary although it is essentially subjective.

In the study of sediments, the separation of primary and secondary processes, and hence primary and secondary mineral deposits, is very difficult. This is partly because primary deposits are often in such chemical disequilibrium with their environment that they are radically changed during diagenesis. It is also due to the fact that there is still considerable debate as to the origin of many of these types of mineral deposits. The formation of a mineral deposit is largely reliant upon a wide variety of processes acting in concert. The individual processes involved are common e.g. precipitation, sedimentation, mechanical and chemical concentration of metals and replacement. But their concerted effort in a specific area in both time and space is not. These processes do not necessarily always occur in unison but even in sequential occurrence they constitute a remarkable coincidence. So it is imperative to decipher this sequence of events if the genesis of a mineral deposit type is to be determined. Once the genesis is deduced then a classification based upon genetic constraints will be of value. However, until these problems are resolved such constraints are excluded from this classification.

Once a mineral deposit has been defined as either primary or secondary, it is further classified according to the nature of the host rock and the type of mineralisation that is present. These constraints are objective and non-genetic so they invoke no controversy (Table 3.1). A primary sediment-hosted mineral deposit is the product of sedimentary processes at, near, or above the sediment-water interface. These deposits are themselves sediments so they can show all sedimentary features. Whilst some deposits may be confined to specific geotectonic settings, the main controls on this type of deposit are the physical and chemical conditions of sedimentation - hence palaeoclimate and palaeogeography may be particularly important.

A secondary sediment-hosted mineral deposit (Table 3.1) is one which is not obviously a sediment itself nor has it a clear igneous source for the mineralisation. Secondary sedimentary deposits appear to be more closely related to basin evolution and fluid migration than primary deposits and they may also be more closely controlled by geotectonic processes, particularly those which affect the temperature and routes of migrating fluids. Therefore geotectonics and environment may be particularly important constraints on secondary ore formation.

Table 3.1 CLASSIFICATION OF MINERAL DEPOSITS

PRIMARY DEPOSITS	SECONDARY DEPOSITS
PLAU, Placer Gold	SSCU, Sandstone Copper
PLDI, Placer Diamond	SHBM, Shale Base-Metal (Cu, Pb, Zn)
PLSN, Placer Tin	SSPB, Sandstone Lead
PLOX, Placer Oxide (Cr, Ti, Fe)	SHUR, Shale Uranium
PLOT, Placer Others (e.g. Platinum)	LSBM, Limestone Base-Metal (Pb, Zn, Ba, F)
PAPL, Palaeoplacer (Au, U)	SSUV, Sandstone Uranium-Vanadium
FEFM, Iron Formation (oxide - carbonate)	LATO, Laterites (Ni, Cr, Fe, Mn, Al)
MNFM, Manganese Formation	GOSS, Gossan
MNNO, Manganese Nodules	SUPE, Supergene Enrichment
OOFE, Oolitic Ironstone	SULF, Sulphur
SDEX, Sedimentary Exhalative (Pb, Zn, Ba)	
CALU, Calcrete Uranium	
PHOS, Phosphate	
MEVA, Marine Evaporites	
CEVA, Continental Evaporites (Li, B, Na)	

The following section is a summary of the main characteristics of each deposit type - as defined for this project. Discussion of their mode of origin will be attempted in a later chapter. There does appear to be a continuum of deposit types. Even if this ultimately proves to be coincidental it is still true that some examples may be classed in more than one group. Amongst these are included the Copperbelt ores of Zambia and Zaire (classed in this thesis as SSCU or SHBM); the uranium deposits of the Colorado Plateau (SSUV); the Kupferschiefer of Europe (SSCU and SHBM). Each of these deposits will be described in the subsequent sections. In an attempt to give an orderly description of the classification mineral deposit types have been grouped according to their mineralisation i.e. U, Al and Ni, Mn, Fe, Phosphate, Cu, Pb and Zn. Placers have been grouped in a separate section and each mineral type discussed separately.

3.2 The Classification in Detail.

3.2.1 Uranium

Abundance and Distribution

Table 3.2 shows uranium is present in low concentrations in most rocks. There is an enrichment in granites relative to other igneous rocks. Granites and rhyolitic ashes are commonly cited as sources of the uranium in orebodies. Among sedimentary rocks concentrations are also low, except in black shales where uranium can be concentrated up to ore grade. The greater mobility of uranium in comparison to thorium is evident in the values for Th/U in natural waters which are much lower than in most rocks. In general the uranium content increases with magmatic fractionation in the more silicic alkalic rocks and increases with organic matter in sedimentary rocks e.g. black shales. A more detailed synopsis on distribution and availability, is given by Bowie (1976, 1979).

Table 3.2. Abundance of U and Th in Common Rocks and Natural Waters.

Rock Types and Natural Waters	U (ppm)	Th/U
Igneous Rocks		
ultramafics	0.02	5.0
basalt	0.50	3.1
andesite	2.0	2.4
granite	4.0	4.9
Sedimentary Rocks		
quartz arenite	0.45	3.5
greywacke	2.1	3.2
arkose	1.5	3.3
shale		
grey and green	3.2	4.9
red and yellow	2.0	6.5
black		
average	53.0	
Chattanooga	79.0	
Alum Shale	168.0	
Ohio Shale	50.0	0.19
limestone	2.2	0.7
dolomite	1.0	
phosphorite	300 - 50	<0.1
Natural Waters		
seawater	6 - 0.3	<0.03
ground water	10 - 0.3	
river water	10 - 0.03	<0.03

Source: Maynard, 1983.

Classification of Uranium Deposits

Uranium deposits have been separated into four groups; palaeoplacer Au-U (PAPL), shale-hosted (SHUR), calcrete uranium (CALU) and sandstone-hosted (SSUV) deposits. Of these four groups, CALU deposits are Recent in age and have not been included in the thesis (section 3.4.1) and PAPL deposits are mentioned in the section concerned with placer deposits as the host rocks for the mineralisation are quartz-pebble conglomerates thought to have been detrital in origin. The remaining two classes were chosen to reflect the dominant host rock for mineralisation.

Sandstone-hosted Type (SSUV)

Two varieties of deposit are included within this group which are recognized by both the geometry of the deposit and the style of mineralisation.

a) Roll-front variety: These occur as small orebodies, for example, in non-marine Mesozoic to Cenozoic sandstones of Wyoming, Colorado, New Mexico and Utah.

b) Tabular Variety: These lack the definite oxidized/unoxidized boundary within the ore body which is obvious in the roll-front type. These orebodies are nearly concordant with bedding, containing copper and vanadium in addition to uranium. The uranium is associated with either woody material or pyrite in the rocks. These presumably acted as localized reductants, for example in the Grants Region, New Mexico, USA (Granger et al, 1961; Turner-Peterson, 1985).

Shale-hosted Type (SHUR)

Bell (1978) defined a black shale as a dark coloured, very fine-grained sediment without reference to fracture or fissibility habit, or degree of induration. Highly organic-rich layers, lenses and veins, which may be remobilized organic material, are often present e.g. kolm in the Alum Shales. The dark colour of the shales that host the low-grade uranium is directly related to their organic content, which may be greater than two percent (Bell, 1978). The shales are also characterized by laminations and a lack of bioturbation, the latter only occurring in less uraniferous and carbonaceous layers. Examples of SHUR deposits include the Alum shales of Sweden and the Phosphoria Formation of Western USA. One of the more notable characteristics of SHUR deposits is that some of the shales possess phosphatic nodules or have thin, phosphatic horizons. In the Phosphoria Formation, as the name suggests, the phosphatic layers are well developed, up to one metre thick. They contain an average of 100-200 ppm U_3O_8 (Maynard, 1983). The palaeolatitudinal distribution of SHUR deposits as a single group has not been determined as many of the examples collected are Cambrian in age for which no palaeogeographic reconstruction is available here. Also a number of SHUR deposits have been incorporated in to the PHOS group due to their high phosphate content.

Some authors, for example, Nash (1981) and Maynard (1983), proposed that there are five distinct types of uranium deposit which have a sequential distribution with time. This is a popular view and a brief explanation follows as a modified version of such a classification has been adopted here. (i) The oldest group, Proterozoic conglomeratic gold-uranium deposits (PAPL), possess pyrite and uraninite which are regarded as detrital in origin. Well-known examples of major deposits of this type are found in the Witwatersrand of South Africa and Elliot Lake, Canada. (ii) The vein, or unconformity, deposits occur in younger Proterozoic rocks, such as the Athabasca Sandstone of Western Canada and in rocks of similar age in Australia at Ranger and Jabiluka in the Northern Territory. These are amongst the largest and highest grade uranium deposits known. These have been incorporated into the more general class, SSUV. (iii) In the Phanerozoic era uranium mineralisation occurs in black shales such as the Alum Shale, Scandanavia and the Chattanooga Shale, eastern USA. The black shales have large tonnages, but the grade is low - about ten percent that of other types - and are therefore uneconomic. These are obviously covered in the SHUR group in the classification here. The older SSUV deposits of the roll-front and tabular varieties are also included here. (iv) The fourth group are the sandstone-hosted deposits of Mesozoic to Cenozoic age in the USA. They are individually fairly small, and are also incorporated into the SSUV class. (v) The Tertiary to Recent calcrete deposits of Australia and South West Africa.

These are surficial accumulations which differ in having uranyl vanadates as the chief ore minerals.

The classification for uranium deposits outlined earlier was chosen in preference to that of Maynard as the latter relied too heavily upon the categorization of deposits according to their supposed genesis. This was in conflict with the caveat laid down in the introduction to this chapter that no genetic constraints would be introduced into the classification of any deposit type.

3.2.2 Aluminium and Nickel

The laterite group of deposits encompasses both nickeliferous laterites and aluminium karst bauxites. Nickel and aluminium are enriched to ore grade by soil processes in many parts of the world.

Abundance and Distribution

Virtually all of the world's aluminium and perhaps one third of its nickel comes from these deposits (Lelong et al, 1976). Indeed, Edwards and Atkinson (1986) suggest that lateritic nickel accounts for about sixty-five percent of the known land-based nickel reserves in the freeworld countries. The aluminium deposits overlie a variety of rock types, but lateritic nickel is associated exclusively with ultramafics, reflecting the relative abundances of the two elements (Table 3.3).

Deep weathering under humid tropical conditions is necessary for the formation of both of these types, and aluminium ores are widely distributed in parts of the world having such climates now, or which had them in the Tertiary. The main type (approximately forty percent) of aluminium karst deposits is that found in countries which border the Mediterranean Sea (Edwards and Atkinson, 1986) and Jamaica.

Most crustal rocks (Table 3.3) contain a high proportion of aluminium-bearing minerals, so bauxites are formed from many rock types e.g. limestones, Jamaica; kaolinitic sands and clays, Weipa, Australia; nepheline syenites, Arkansas, USA. Production of lateritic nickel is largely confined to New Caledonia, Indonesia and Greenvale, Queensland. Smaller deposits are found in Oregon, USA, Cuba and the Dominican Republic. It has been noted (Golightly, 1981) that the major lateritic nickel deposits have formed from the mid-Tertiary to the present, although some deposits which occur outside the modern laterite soil belt, including those of Greece and Yugoslavia, are older and lie beneath a Cretaceous and Tertiary cover.

Table 3.3. Abundance of Ni and Al in Common Rocks and Waters.

Rock Types and Waters	Ni (ppm)	Al_2O_3 %	Al_2O_3 % / Fe_2O_3
Igneous Rocks			
ultramafics	1450		
peridotite		4.0	0.3
basalt	130	14.1	1.1
andesite	18	18.2	1.9
granite	10	13.9	4.9
nepheline syenite		21.3	4.6
Sedimentary Rocks			
sandstone			
orthoquartzite		1.1	1.6
arkose		8.7	3.5
greywacke	40	13.5	2.1
shale	70	14.7	1.8
black shale	50		
limestone	5		
carbonate		2.5	5.0
Sediments and seawater			
Pacific Mn nodules	3120		
Pacific deep-sea sediment ($CaCO_3$ - free)	300		
near-shore clay	40		
seawater	0.0005		
Natural Waters		Al ppm	
seawater		0.001	
river water		0.240	
springs in granite		0.018	

Source: Maynard, 1983.

Classification

Aluminium Karst Bauxites.

The aluminium karst bauxites (Bardossy, 1982) overlie highly irregular, karstified limestone and dolomite surfaces. Bardossy recognized six different types of karst bauxites, based on their depositional characteristics. All of these are included in this study.

Nickeliferous Laterite Deposits.

The nickeliferous laterite deposits of greatest economic importance have developed on peridotite bedrock. The nickel has been concentrated by the leaching of forsteritic olivine, serpentine or nickeliferous magnetite in the same rock. Golightly (1981) classified nickeliferous laterite deposits into four main types on the basis of the serpentine content of the host rock and the presence or absence of an intermediate zone in the laterite profile. Briefly, these classes are as follows;

Type Ia - laterites developed over unserpentinized peridotite,
Type Ib - laterites developed over partly serpentinized peridotite,
Type Ic - laterites developed over fully serpentinized peridotite.
The fourth class consists of profiles with silica boxwork or nontronite zones.

The latter typically occur in less humid tropical climates with a marked dry season, whereas Types Ia - c are characteristic of the humid equatorial zone or other locales with a very high rainfall and minimal dry season.

Within this report no such classification has been adhered to and all varieties of nickeliferous laterites have been included so as not to bias the results.

3.2.3 Manganese

Manganese is more abundant in basic than in acid igneous rocks (Table 3.4) as expected from its similarity to iron. With regard to sedimentary rocks, $CaCO_3$-free deep-sea clays contain considerably more MnO than any other sediment.

There are a number of types of manganese deposit which are thought to have different origins and are hosted by a variety of rock types. In this study, attention was only centred on two classes, the clastic-hosted and carbonate-hosted syngenetic groups as there may be a palaeolatitude control on their formation. They are reminiscent of some metallic mineral deposits in that they formed syngenetically in sediments.

One type of deposit, the volcanic-sedimentary accumulations, have not been included here as they are outside the scope of this investigation. Neither have their metamorphosed equivalents, for example, the Kalahari Desert deposits (Roy, 1981). The manganese deep-sea nodules, which are a special type of syngenetic manganese deposit covering the floor of many deep ocean basins, constitute another class which has been excluded from this discussion. These nodules are uneconomic at present, and their exploitation is riddled with technical, economic and legal problems (Archer, 1976), although they are a huge potential source of Cu, Ni and Co as well as Mn. Glasby (1977) and Glasby and Read (1976) gave an extensive summary of this particular deposit type.

The fourth class are of supergene origin in which the original type of accumulation of manganese is obscured by surficial oxidation. These are excluded as too few examples were collected to allow evaluation of any latitude control on formation. Lastly those deposits which occur in Proterozoic rocks and are similar to the large banded iron formations have been excluded on the ground of age of formation of these deposits.

Sediment-hosted manganese deposits include both carbonate and oxide minerals and lateral zoning of the sediment facies can be distinguished. Oxide ores occur nearest the continental margin, then the oxide-carbonate ore facies, and lastly, the carbonate ores furthest from the shore. Perhaps the most well-known examples of clastic-hosted manganese deposits are that at Nikopol and Chiatura in the USSR, which contain between seventy-five and eighty percent of the world's present proven reserves. The manganese horizon of this early Oligocene accumulation is oxide in the north changing to carbonate in the south, corresponding to an increase in water depth as described above (Varentsov and Rakhmanov, 1980).

Stratiform manganese deposits in carbonate host rocks are found in the Mesozoic of Morocco, Lower Cambrian carbonates of the Appalachians and Pre-Cambrian rocks of India (Roy, 1981). The Moroccan ore lies between clastics which are a coarse, near-shore to continental red-bed sequence and carbonates which are mostly fine-grained dolomites (Varentsov, 1964). An arid climate at the time of deposition has been inferred from the association with red-beds and gypsum.

Table 3.4. Abundance of Mn in Common Rocks.

Rock Types	MnO (ppm)	Mn/ Fe
Igneous Rocks		
granite	260	0.015
granodiorite	390	0.017
diorite	1390	0.019
gabbro	1390	0.016
peridotite	1050	0.016
Sediments and sedimentary rocks		
greywacke	690	0.020
quartz sandstone	170	0.030
shale	600	0.013
black shale	150	0.008
limestone	550	0.12
deep-sea clay ($CaCO_3$ - free)	5700	0.095
seawater	0.0013	

Source: Maynard, 1983.

3.2.4 Iron

Abundance and Distribution.

Iron is one of the most abundant metals, so it is a major constituent of most rocks (Table 3.5) and its ores are rock types in their own right. Igneous rocks, although varying in total iron, have about the same proportion of Fe^{3+} to Fe^{2+}. In contrast, sedimentary rocks show a wide range of oxidation states, indicative of the presence of environments of differing oxidation potential. Therefore it is in sediments that the greatest potential for iron enrichment exists, and most iron ores are sedimentary.

Deposits of sedimentary iron ore are copious throughout the stratigraphical column, occurring in every shield area of the world, and in all the Phanerozoic systems with the exception of the Triassic (Taylor, 1967). Among Phanerozoic formations perhaps the Ordovician-Silurian (e.g. Clinton Ores) and Jurassic systems (e.g. Northampton Ironstone) are the richest as they represent periods when conditions were especially favourable for the deposition and preservation of ferruginous sediments within a great number of isolated basins (James, 1966).

Table 3.5. Abundance of Iron in Common Rock Types.

Rock Types	FeO %	Fe_2O_3 %	$\frac{Fe_2O_3}{FeO+Fe_2O_3}$
Igneous Rocks			
alkali-olivine basalt	7.9	4.2	0.35
tholeiitic basalt	9.5	3.2	0.25
granodiorite	2.6	1.3	0.33
granite	1.5	0.8	0.35
Sedimentary Rocks			
sandstone			
quartz arenite	0.2	0.4	0.67
lithic arenite	1.4	3.8	0.73
greywacke	3.5	1.6	0.31
arkose	0.7	1.5	0.68
Shales and Slates			
red	1.26	5.36	0.81
green	1.42	3.48	0.71
black	4.88	0.52	0.10

Source: Maynard, 1983.

Classification.

There are numerous classifications of sediment-hosted iron deposits. For example Sokolov and Grigor'ev (1977) distinguished seven types of deposit and Kimberley (1978) formulated six. He produced a detailed classification of iron formations based upon the palaeoenvironment of their deposition. The palaeoenvironments were deduced from the characteristics of enclosing rocks and from sedimentary features e.g. textures and sedimentary structures, of the iron-rich rocks. An iron formation is described as a mappable rock unit composed mostly of ironstone, with the uppermost and lowermost beds being ironstone.

Kimberley (1978) noted that the distribution of various types of iron formation with time reflects the relative abundance of particular geological environments at the time of deposition. Volcanic environments were most common in the early Precambrian; continental-shelf environments dominated the Middle Precambrian (FEFM) and inland-sea environments in the Phanerozoic (OOFE) and this sequential distribution with time was evident in the collection of examples of sediment-hosted ferriferous chert-rich and chert-poor rock types.

Iron formations (FEFM) are those classified by Kimberley (1978) as Metazoan-poor, extensive, chemical-sediment-rich, shallow-sea iron formations (MECS-IF) and examples belonging to the oolitic ironstone group (OOFE) are those of the sandy, clayey and oolitic, shallow-inland-sea iron formation type (SCOS-IF). The production of iron from ore deposits is now almost entirely from these two types; as of 1970, they constituted more than ninety percent of the world's production (Maynard, 1983). The main characteristics of each type are briefly described below.

Iron Formations (FEFM)

These are the 'Iron Formations' of James (1954 and 1966) and they globally constitute significant proportions of the Lower to Middle Proterozoic sequences. There are two main varieties of FEFM, the Algoma type of volcanic association and the larger Lake Superior type which is considered here. Both have a shallow-shelf orthoquartzite-carbonate association. Well developed banding of light and dark coloured interbeds is characteristic of this deposit type. The lighter bands of micro- to crypto-crystalline silica, or chert, alternate with bands rich in ferriferous minerals such as haematite, magnetite, greenalite, siderite, stilpnomelane, minnesotaite, grunerite, pyrite and ankerite (Kimberley, 1978). An oolitic texture is occasionally found in North American FEFMs (Gross, 1965) but it does not predominate in any FEFM (Kimberley, 1978). These Precambrian banded, cherty FEFMs in the Lake Superior region were deposited on broad continental shelves, and deposits of the supratidal, intertidal and subtidal zones can be recognised in the well-known carbonate-oxide facies of the FEFMs (Lougheed, 1983).

The main examples are found in:-

Transvaal, South Africa	(Button, 1976)	
Lake Superior region	(Bayley and James, 1973)	2000ma
Sokoman Formation, Labrador	(Knoll and Simonson, 1981)	1900ma
Hamersley Basin, W.Australia	(Trendall, 1975)	

Oolitic Ironstones (OOFE)

The oolitic ironstones have a variety of names and have been described previously by many authors. They are the Clinton-type or Minette-type iron formation of Gross (1965); the Phanerozoic ironstones of James (1966) and the sandy, clayey, oolitic, shallow-inland sea iron formations (SCOS-IF) of Kimberley (1978). They are characterized by widespread ooliths - most commonly of chamosite or goethite. James and Vanhouten (1979) described OOFEs from northeast Colombia and Venezuela as multi-layered, essentially spherical, symmetrical ooids of goethite and rarely with small amounts of chamosite.

Occurrences of OOFEs are most common in the Phanerozoic, although some Precambrian examples have been described (Button, 1976a). However they are not evenly distributed throughout the Phanerozoic apparently being concentrated in two major time spans. The first spans the Ordovician to the Silurian; for example, Clinton deposits of USA and the Wabana beds, Newfoundland ores (Douglas, 1974). The second period is from the Jurassic to Early Cretaceous with the so-called Minette ores of Europe (e.g. Dunham et al, 1978).

OOFEs are fairly common with individual deposits usually smaller than Precambrian FEFMs (Maynard, 1983) from which they differ in possessing an oolitic rather than a banded texture, having a lower chert content and a different mineralogy. The most abundant iron minerals are goethite, chamosite and siderite and other minerals commonly found include apatite, kaolinite, dolomite, calcite, detrital quartz and phosphates. Locally, magnetite or pyrite may predominate (Kimberley, 1978; Ferguson et al, 1983; Maynard, 1983).

Kimberley (1979) noted the concentration of some elements which occur in certain iron-poor oolitic sediments as well as in iron-rich beds. He cited the manganiferous oolites with minor associated OOFE in Nikopol (Varentsov, 1964) and the oolitic phosphorite of the Permian Phosphoria Formation (McKelvey et al, 1959; and Sheldon, 1963) amongst others.

3.2.5 Phosphate Deposits.

Many phosphate deposits are local in character and are found as special phases within formations of a different nature, as described in the section on SHUR and OOFE deposits e.g. the Phosphoria Formation (McKelvey et al, 1959; Sheldon, 1963; Maynard, 1983). Examples of the SHUR variety have been included in the PHOS group as the phosphatic horizons and nodules are well developed. In contrast, other PHOS deposits occur on a much larger scale and constitute independent marine formations covering a considerable area, as described later. The phosphate content of a rock is expressed as a percentage of P_2O_5 and any rock which contains greater than 18% P_2O_5 is known as a phosphorite (Maynard, 1983); most of which are marine in origin or result from weathering of marine deposits.

Phosphorites contribute more than 80% of the world's production of phosphate rock (Maynard, 1983). They have therefore been the subject of numerous studies. The main publications are Bushinski (1969a) for Asian deposits; the British Sulphur Corporation (1964 and 1971) for a world survey of phosphate deposits and the U.S. Geological Survey Professional Papers , A-F (1959). More recent work has been carried out by Summerhayes et al (1972 and 1973) and Parrish (1982), Parrish and Curtis (1982), Parrish et al (1983 and 1986).

There are a variety of PHOS deposits. To illustrate the extent of this diversity five examples were selected at random and their characteristics summarized below.

1) The Miocene Sechura deposits of Peru are marine and pelletal in form with an average P_2O_5 content of 20% (Cheney et al, 1979) whereas

2) the seafloor phosphorites of Agulhas Bank are phosphatized limestones with carbonate and goethite zones in addition to phosphate. The latter is present as cement filling fractures and voids (Baturin and Dubinchuk, 1974).

3) The Ordovician Maquoketa Shale, Iowa has a basal, silty layer phosphorite layer with up to 22.5% P_2O_5, although the average is 13% P_2O_5. In places, however, the shale becomes a phosphatic dolomite containing 17% P_2O_5 (Brown, 1974).

4) Cathcart (1977) described a late Precambrian or early Cambrian deposit near Patos de Minas, Brazil where the phosphorite is laminated, consisting of black, elongated apatite pellets and quartz grains of average 13% P_2O_5.

5) D'Anglejan (1967) recorded a bedded phosphorite in recent continental shelf sediments of Baja California, Mexico. He suggests that the lithology present at Baja, that of carbonate-fluorapatite with opaline silica and reducing muds, is reminiscent of the well-known chert-carbonaceous shale-phosphorite association found in ancient deposits.

Not only are the phosphorites divergent in their nature, but so are the 'dilutents' (non-phosphatic matrix and their interbeds) and the associated sediments, as noted by Cook (1976). These include the following:-
a) mudstone and shale e.g. southeast Idaho; Maquoketa shale, northeast Iowa (Brown, 1974), b) chert e.g. northeast Utah, Karatau, c) limestone and dolomite e.g. northwest Queensland (Cook, 1972b); Agulhas Bank (Baturin and Dubinchuk, 1974), d) sand and sandstone e.g. central Florida (Riggs, 1979).

Abundance and Distribution.

Cook (1976) stated that phosphate ore originates from three sources:
a) igneous apatite accumulations such as those associated with the nepheline syenites and carbonates of South Africa and the Kiruna occurrence.
b) guano-derived deposits e.g. Nauru and Christmas Island,
c) marine sedimentary phosphate deposits.

Only the sedimentary PHOS deposits will be discussed here, as those of igneous association are not relevant to this research and land-based guano-derived deposits have their own characteristics related to a unique biomass influence. The principal deposits of guano are formed on rocky islands frequented by sea birds, the greatest accumulation being found in the dry Trade Wind belts e.g. the West Indies and islands of the Pacific Ocean. Although the distribution of these deposits is apparently constrained by atmospheric circulation patterns, and hence latitude (see Chapter 4, section 4.5.1) there are no fossil examples so they have not been discussed here. Marine sedimentary (sediment-hosted) PHOS deposits are known to occur in every continent except

Antarctica and range from Precambrian to Recent in age. A few attempts have been made to show there is a worldwide cyclicity in the deposition of phosphate. For example Strakhov (1969) suggested that worldwide phosphate maxima occurred during the Late Cretaceous to Early Tertiary, Permian, Early Cambrian, Sinian (Riphean) and the Proterozoic. Cook (1976) disputed these interpretations on the grounds of the huge disparity in duration of these so-called maxima. Bushinski (1969b) also disputed this concept and suggested that phosphate accumulates in response to localized phenomena.

There is, however, a general consensus amongst workers that phosphorites have been deposited in extensive phosphogenic provinces during specific time intervals (Figure 3.1). Bushinski (1969b) determined the following provinces:

1) Late Precambrian of central and southeast Asia,

2) Cambrian of central and southeast Asia, extending into northern Australia (Bushinski, 1969a; Howard and Hough, 1979),

3) Permian province of North America (McKelvey et al, 1959). There are relatively few Permian deposits but their total phosphate content is boosted by the enormous deposits of the western Phosphate Field of the U.S. (Cook and McElhinny, 1979),

4) Jurassic to Lower Cretaceous eastern European province,

5) Upper Cretaceous to Eocene Tethyan province of the Middle East (Sheldon, 1967) and North Africa extending to West Africa (Birch, 1979) and the northern part of South America,

6) Miocene province of southeast N.America (Gibson, 1967; Riggs, 1979).

There are apparent hiatuses of phosphate deposition during the Oligocene, Triassic and the Silurian to Lower Carboniferous periods (Cook and McElhinny, 1979).

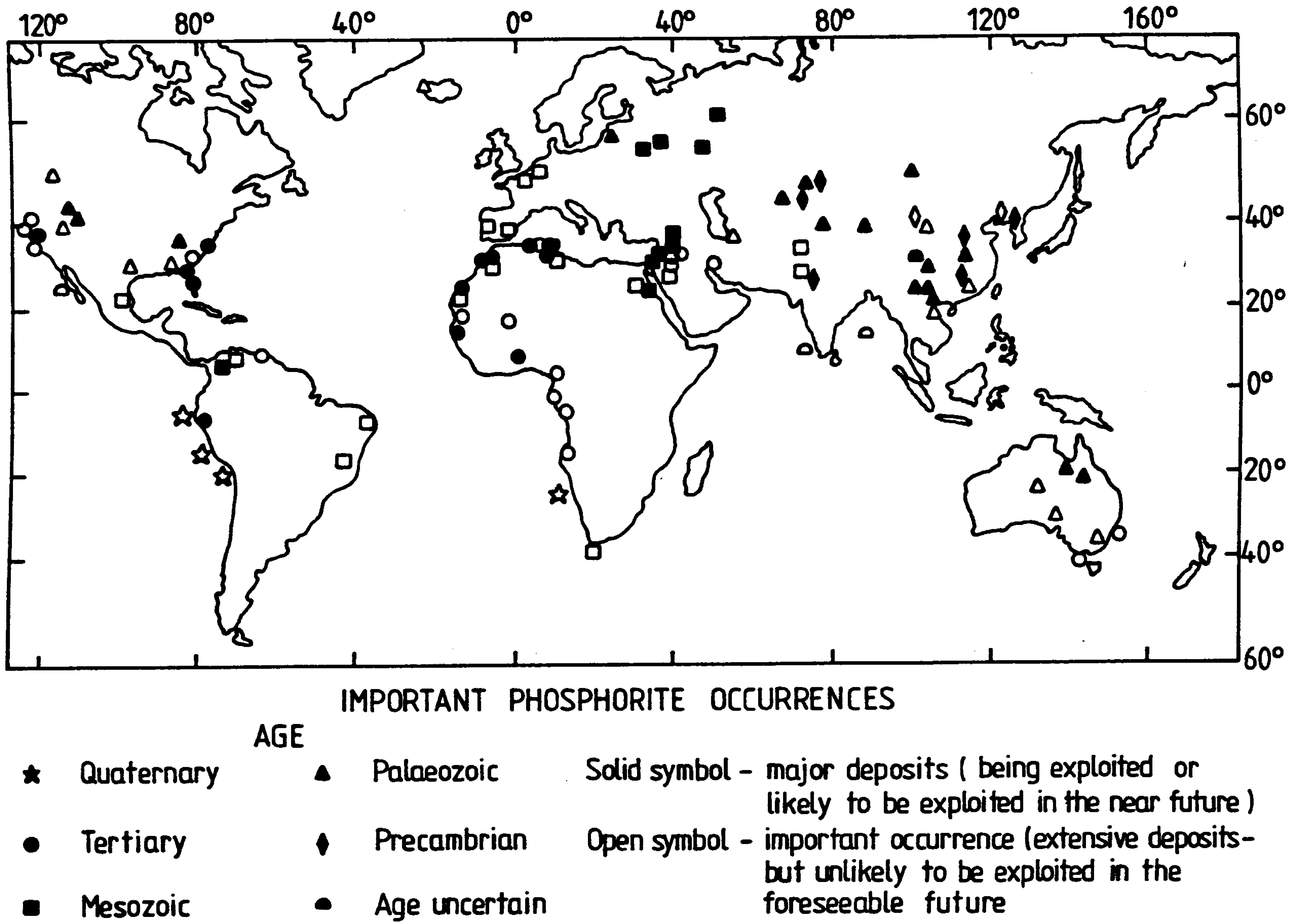


Figure 3.1. Distribution of major sediment-hosted phosphate deposits. (After Cook, 1976, Figure 1).

Classification.

A number of common features of the PHOS deposits may be used as a basis for a classification scheme although no two deposits are precisely the same. One of the most widely used classifications is that devised by Cook (1976) which is as follows;

1) Geosynclinal or West-Coast Type e.g. the western phosphate field of the U.S. (McKelvey et al, 1959),

2) Platform or East-Coast Type e.g. North Carolina (Gibson, 1967); Baja California, Mexico (D'Anglejan, 1967),

3) Weathered or Residual Type e.g. brown rock deposits of Tennessee; Northern Pakistan (Ghaznavi et al, 1983).

The difficulty with such a classification is that deposits commonly have characteristics of more than one type and the genetic implications in the three terms are undesirable, as mentioned with reference to some uranium deposits. An alternative scheme to that presented above was developed by Parker (1971) and comprised the two conglomeratic and three non-conglomeratic classes briefly described below;

Non-Conglomerates

NI Phosphatized glauconite- and quartz-poor microfossiliferous limestone.

NII Phosphatized microfossiliferous limestone rich in goethite.

NIII Phosphatized microfossiliferous limestone, highly glauconitic and quartz-rich in a micrite/fluorapatite cement.

Conglomerates

CI Rich in glauconite with pebble inclusions of phosphatized foraminiferal limestone and a cement similar to a NII rock.

CII Low in glauconite, highly microfossiliferous, with abundant goethite.

It is important to stress that these rock types form a continuous series. Birch (1979) successfully used this scheme to classify the phosphatic rocks off the western margin of South Africa and noted that these classes were merely designed to group together rocks which were deposited under similar conditions and to indicate whether there had been any reworking or transportation.

The PHOS deposits used in this study have not been subdivided (other than to determine whether or not they were marine sediment-hosted) for two main reasons. Firstly the dilemma of which of the major types of classifications (Parker, 1971 or Cook, 1976) to choose had to be overcome. Secondly once one had been chosen it would be extremely difficult to follow such a rigid classification when there are so many varieties of PHOS deposits. However care was taken to note if the different groups within the classifications were reflected in the results i.e. were there marked ranges of palaeolatitude for a particular PHOS deposit type?

3.2.6 Placer Deposits.

Placer deposits comprise alluvial, eluvial and colluvial material which contain economic quantities of some valuable material. Those containing gold, platinum, diamonds and tin (cassiterite) are amongst the most important types. These surficial deposits are formed by the separation of light, friable minerals and heavy, chemically resistant minerals from weathered debris and the mechanical concentration of these mineral particles. The mechanical agent is usually alluvial (streams) although it may also be marine, aeolian, lacustrine or glacial. Therefore for a mineral to occur as a placer mineral it must be highly resistant to water and abrasive action and be a heavy mineral (see Table 3.6). The latter is a term generally applied to minerals which sink in bromoform (S.G. 2.9).

Placer formation is favoured by prolonged sediment reworking causing heavy mineral concentration while transporting the remaining sediment downstream.

Placers might therefore be expected to be associated with surfaces of fluvial degradation in drainage basins and in alluvial fan sequences of depositional basins (Schumm, 1977) which is frequently the case.

Table 3.6.Physical Properties of the more common Placer Minerals.

MINERAL	HARDNESS MOH'S NO.	SPECIFIC GRAVITY	PRINCIPAL PLACER ENVIRONMENT
Cassiterite SnO_2	6 - 7	6.8 - 7.1	Eluvial, fluvial, marine
Diamond C	10	3.52	Beach, fluvial, eluvial
Garnet $R''_3R'''_2Si_3O_{12}$	6.5 - 7.5	3.5 - 4.3	Fluvial, eluvial
Gold Au	2.5 - 3	19.3	Fluvial, eluvial, beach
Ilmenite $FeTiO_3$	5 - 6	4.5 - 5	Beach sand
Magnetite Fe_3O_4	5.5 - 6.5	5.1 - 5.18	Beach sand
Monazite (Ce, La,Y,Th)PO_4	5	4.9 - 5.3	Beach sand
Platinum Pt	4 - 4.5	14 - 19	Fluvial
Ruby & Sapphire Al_2O_3	9	3.95 - 4.10	Fluvial, eluvial
Rutile TiO_2	6 - 6.5	4.2	Beach sand
Zircon $ZrSiO_4$	7.5	4.5 - 4.7	Beach sand

Modified from Hails (1976), Table 1 and Edwards and Atkinson (1986), Table 5.2.

Placer deposits are formed by natural surface processes so they have a wide geographical distribution. The majority are of Tertiary and Quaternary age which may be due to poor preservation of older deposits and the changes in sea level since the Quaternary.

Classification.

There are various classifications of placer deposits. Hails (1976) used a division which combined the mode of occurrence with depositional environment so placer deposits were divided into the following groups: marine (offshore), alluvial (stream) including river terrace, beach, eluvial (slope), residual and fossil. Emery and Noakes (1968) devised a classification consisting of three groups, each one being characterized by a combination of its physical properties and environment of deposition. These were the heavy heavy minerals (such as gold, tin and platinum) in stream deposits; light heavy minerals (such as ilmenite, rutile, zircon and monazite) usually occurring as beach deposits and the gem class (mainly diamonds) commonly found in alluvial environments.

The classification presented here is not based upon the geological environment of a particular deposit. It is solely concerned with the dominant mineral (e.g. gold, diamonds etc.) or mineral type (e.g. oxides) present in the placer deposit. However Edwards and Atkinson (1986) suggested that it is common for some minerals to be concentrated in a specific placer environment whilst others (e.g. gold) occur within the entire spectrum of placer deposits (refer to Table 3.6).

Placer Gold Deposits (PLAU).

Gold is the most ubiquitous of metallic placer minerals and is found in many parts of the world. Young PLAU deposits are composed of unconsolidated or semi-consolidated sand and gravel possessing small quantities of gold or other heavy minerals. Most are stream deposits which occur within present day valleys, or on beaches or terraces of pre-existing rivers (Hails, 1976). Giant PLAUs (e.g. Otago and Westland, New Zealand; Colombia; California) occur on Pacific Coast margins and appear to have formed during the Tertiary in similar tectonic and sedimentary environments (Henley and Adams, 1979). Other PLAU are formed by the secondary mechanical concentration or reworking of earlier, extensive auriferous placers e.g. those of southeast Australia (Sutherland, 1985).

Placer Diamond Deposits (PLDI).

Diamantiferous gravels (e.g. Vaal River, South Africa) are of considerable economic importance although they are not common. The primary source rocks for diamonds are kimberlites which themselves have a strong geographical association with cratonic areas (Dawson, 1980) so the distribution of PLDI deposits is restricted to these source areas. In some regions (e.g. Ghana and Brazil) there are economic PLDI deposits but the primary sources for the diamonds have not been established. However the association with Precambrian shield areas still remains (Wilson, 1982). Although most diamantiferous placers are alluvial or eluvial (e.g. Minas Gerais, Brazil; Venezuela; Namibia), there are major marine concentrations off the west coast of southern Africa.

Placer Tin (PLSN).

This category is dominated by detrital cassiterite deposits and by far the largest proportion of tin ore is obtained from alluvial deposits, for example South Island, New Zealand and the southeast Asian Tin Belt (Hails, 1976). Approximately 65% of the world's tin is produced from PLSN deposits in Malaysia, Thailand and Indonesia where alluvial deposits predominate, although some of eluvial origin are found (Edwards and Atkinson, 1986). Other deposits of residual origin are found in the Rondonia district of Brazil which provide up to 60% of that country's production.

Cassiterite (SnO_2) is the most important ore mineral being both very hard and heavy (see Table 3.6). It tends to concentrate naturally in surficial deposits derived from tin-bearing granites (in which it occurs as a primary mineral) and their associated metamorphic rocks.

Placer Oxides (PLOX).

This group consists mainly of deposits with titanium and tungsten ore minerals. The titanium-bearing minerals include;

a) Rutile (TiO_2) which is found as an accessory mineral in igneous rocks, pegmatites and metamorphosed limestones e.g. the Eastern Australian rutile province (McKellar, 1975); Richard's Bay, South Africa; Eneabba deposit, Western Australia (Lissiman and Oxenford, 1975).

b) Ilmenite ($FeTiO_3$) which occurs as an accessory mineral in basic igneous rocks and veins e.g. as Holocene beach deposits, West Coast, New Zealand (NZ DSIR 1969/70); Capel area, west Australia (Welch et al, 1975); Lakehurst area, New Jersey, U.S. (Puffer and Cousminer, 1982).

c) Scheelite ($CaWO_4$), the tungsten-bearing placer mineral, is derived from veins and contact-metamorphic deposits e.g. Otago, New Zealand where detrital scheelite is frequently associated with detrital gold (NZ DSIR 1969/70).

Placer "Others" (PLOT).

This class consists mainly of deposits containing minerals of the platinum group. For example osmiridium is found with alluvial gold in Tertiary conglomerates of New Zealand (NZ DSIR 1969/70). It also includes deposits with gemstones (excluding diamonds) e.g. those deposits occurring in the Mato Grosso and Minas Gerais districts of Brazil (Franco, 1981).

Palaeoplacer Gold-Uranium (PAPL).

Palaeoplacer gold-uranium deposits may be defined as the lithified equivalents of placer deposits (Utter, 1980). The principal examples are located in the Witwatersrand Basin of South Africa (Au-U); the Blind River-Elliot Lake district of Ontario, Canada (U); the Jacobina district, Bahia, Brazil; the Tarkwa region of southwest Ghana (Au).

The lower age limit for the occurrence of these giant PAPL deposits is 3100 m.y., the youngest age is 1900 m.y., and it is notable that Upper Proterozoic conglomerates (that is 1600-1700 m.y.) are conspicuously devoid of PAPL ore deposits (Pretorius, 1981). The South African deposits are of Archaean age (2500-2750 m.y.), the Tarkwa deposit, Ghana is Lower-Middle Proterozoic (1900 m.y.) and the Canadian and Brazilian deposits are considered to be Lower Proterozoic (Pretorius, 1976). These economic deposits are of limited geographical distribution and are confined to the age ranges given above so they appear to represent a rather specific type of preserved placer.

A notable feature of PAPL deposits is the resemblance between the host rocks from the principal districts. These oligomict conglomerates consist predominantly of quartz and chert pebbles in a sericite/chlorite matrix so they are mineralogically mature although texturally immature (Clemmey, 1981). Thin seams of carbonaceous material, thought to represent organic plant remains (Pretorius, 1975), are common and may be associated with high values of uranium and gold (Simpson and Beales, 1981).

Although the mineralogy of the PAPLs is dominated by quartz, they contain a diverse assemblage of accessory resistate and sulphide minerals e.g. platinum group metals, silver and thorium (Pretorius, 1981). Pyrite is the dominant sulphide in all but the Ghanian deposit where haematite, ilmenite and magnetite constitute the assemblage. Fresh, unoxidized pyrite is present as coarsely crystalline grains and as fine-grained concretions in the Witwatersrand example (Simpson and Bowles, 1977).

The sedimentary facies of PAPL sequences is generally accepted as parts of a fluvial fan which prograded into a water-filled intermontane basin (Pretorius, 1975). A detrital origin for uraninite grains in Canadian deposits has been demonstrated (Roscoe, 1969) and for a variety of ore particles of the Witwatersrand (Utter, 1980).

The position of the ore-bearing conglomerates within the sedimentary sequence has similarities in the Brazilian, Canadian and Ghanian deposits as they all possess three mineralised conglomerate horizons which are found near the base of the sequence. However in the Witwatersrand Basin mineralised sediments occur throughout the succession from bottom to top with the main proportion of ore-bearing horizons occurring in the upper part of the Witwatersrand Supergroup (Edwards and Atkinson, 1986).

The source of the detrital material for the PAPLs is thought to be the Archaean granite-greenstone basement terrain on which the sedimentary sequences rest, as suggested by Pretorius (1976) with regard to the Witwatersrand deposits. He considered the gold was derived from ultramafic and mafic igneous rocks of the greenstone belts whilst the uranium was drawn from the granites that intruded them.

3.2.7 Sediment-Hosted Stratabound Base-Metal (Cu, Pb, Zn) deposits.

There have been a plethora of classifications for sediment-hosted stratiform base-metal deposits beginning with Stanton (1972) and others in the early 1970s and continued by Wolf (1976 and 1981), Samama (1976), Gustafson and Williams (1981), Morganti (1981) and Bjorlykke and Sangster (1981).

Morganti (1981) recognized a number of mineral deposits containing zinc, lead, copper, barium and/or precious metals which occur in predominantly clastic sedimentary sequences where volcanic rocks are not demonstrably related to ore formation. These are referred to as sedimentary-type stratiform deposits e.g. Meggen, McArthur River, Mufulira. These were then further divided into three sub-classes based on gross sedimentation related to major tectono-stratigraphic environments, as follows;

1) intracratonic basin sulphide deposits,
2) flysch basin sulphide and barite deposits,
3) platform-marginal basin sulphide deposits.

Other workers, for example Bjorlykke and Sangster (1981), classified stratiform sediment-hosted deposits according to the dominant type of mineralisation i.e. copper, lead or zinc, the major host rock and its associations. A modified version of this classification has been adopted here. Bjorlykke and Sangster presented the subclasses which have been used in this classification.

1) Red-bed copper deposits	(SSCU) and (SHBM)
2) Sandstone lead deposits	(SSPB)
3) Carbonate-hosted lead-zinc deposits	(LSBM)

Cupriferous shales were incorporated into group one together with the sandstone-hosted copper deposits whereas they have been placed into a separate class in this thesis i.e. as shale-hosted base-metal (SHBM) deposits.

Sediment-hosted copper deposits (SSCU and SHBM) appear to have sufficiently distinctive characteristics to separate them from sediment-hosted lead-zinc deposits. The term sedimentary-exhalative is used here for those lead-zinc deposits of syngenetic or diagenetic type (SDEX) whereas epigenetic deposits are commonly described as being of Mississippi Valley-Type i.e. carbonate-hosted lead-zinc group (LSBM). A description of the major characteristics of each of the five deposit classes will be given later.

It has been suggested by some workers, for example Rose (1976), that different deposit types may be related e.g. the Kupferschiefer, Germany, the Zambian Copperbelt and some copper-uranium deposits in the United States. He suggested that the same chemical relationships may apply in view of their low temperature of deposition and certain geological features (e.g. sulphides, organic matter, occurrence near evaporites and red beds). In contrast, Bjorlykke and Sangster (1981) concluded that SSCU, SSPB and LSBM deposit types were clearly separate entities in terms of their tectonic and sedimentary environments and also that they were probably formed at different stages of continental evolution. In this instance each mineral deposit type has been regarded as a separate entity to avoid the application of genetic constraints upon the classification. However it will be noted in a later chapter if any relationship between deposit types becomes apparent from scrutiny of their latitude of formation.

Availability and Distribution of Copper.

Copper has a fairly uniform distribution in intermediate to basic igneous rocks (see Table 3.7). Among sedimentary rocks carbonates are noticeably low in copper which is reflected in the nearly exclusive association of commercial deposits with clastic rocks. Note that the carbonaceous shales have nearly three times the amount of copper found in other shales, showing the importance of reducing conditions for precipitating copper in sediments. By far the largest concentration is in pelagic clays where copper is associated with cobalt and nickel in iron-manganese nodules. This is likely to be an important source of copper along with a number of other metals in the future (see section 3.2.3). Silver has a similar distribution as copper but it is not as strongly depleted in seawater, carbonate rocks and quartz arenites.

Table 3.7. Abundance of Copper and Silver in Common Geological Materials.

Rock Types and Sediments	Cu (ppm)	Ag (ppb)	Cu/Agx1000
Igneous Rocks			
peridotite	47	60	0.78
basalt	90	100	0.90
andesite	53	80	0.66
granite	13	37	0.35
Sedimentary Rocks			
shale			
average	35		
red		150	
green		190	
black	95	290	0.33
sandstone			
quartz arenite	30		
arkose			
greywacke	11	250	0.044
limestone	6	125	0.048
Sediments, seawater			
pelagic clay	251		
seawater	0.0015	0.32	0.0047

Source: Maynard, 1983.

Distribution and Availability of Lead and Zinc.

Lead is uniform by distribution in sedimentary rocks while zinc is somewhat enriched in carbonaceous shales (see Table 3.8). Both are remarkably low in carbonate rocks suggesting that carbonate-hosted ores are externally derived. The concentration of lead in most natural waters is exceedingly small, except in highly saline ones. Therefore it would seem unlikely that lead-zinc deposits formed by direct precipitation from seawater or other normal surface waters.

Table 3.8. Concentration of Pb and Zn in some Rocks and Natural Waters.

Rock Types and Natural Waters	Zn (ppm)	Pb (ppm)
Igneous Rocks		
peridotite	56	0.3
gabbro	100	3.2
diorite	70	5.8
granodiorite	52	15.0
granite	48	24.0
Sedimentary Rocks		
sandstone		
quartzose, arkose	30	10.0
greywacke	95	20.0
shale		
average	100	-
carbon-rich	200	24.0
carbon-poor	-	23.0
carbonates	20	5.0
Modern Sediments		
marine mud	90	23.0
pelagic clay	140	55.0
Natural Waters		
seawater	0.005	0.00003
interstitial water	0.012	-
Salton Sea brine	780	80.0
deep formation brine, Canada	750	-
Atlantis II Deep brine	5.4	0.6
deep formation brine, Mississippi	155	30.0

Source: Maynard, 1983.

Sandstone Lead Deposits (SSPB)

SSPB deposits are relatively rare as lead and zinc are usually found in carbonates and shales although in countries such as Sweden they constitute an important resource. The Laisvall deposit of Sweden is perhaps one of the most well documented examples of this type and a considerable amount of work has been done on many aspects of this deposit (Rickard et al, 1975, 1979 and 1981).

In SSPBs the host rocks to the lead are basal quartzitic sandstones which usually constitute part of a sedimentary sequence lying on Precambrian basement as at Vassbo, Sweden (Christofferson et al, 1979).

SSPBs are characteristically low grade, lead dominant, pyrite free and silver poor (Bjorlykke and Sangster, 1981). Minerals of the ore association are usually galena, sphalerite, calcite, fluorite and barite and they occur as cement, infilling the pore spaces in the sandstone. A common feature of SSPBs is a high grade core surrounded by a lower grade halo and, if zinc is present, it occurs in a position stratigraphically higher than lead. This is most obvious at Laisvall with zinc being dominant in the Upper Sandstones to the northwest and very minor in the galena-rich Lower Sandstones (Rickard et al, 1979).

The depositional environments for these deposits range from continental to shallow marine. It is considered to have been an open tidal beach environment at Vassbo (Christofferson et al, 1979) and on a stable platform at the shallow tidal margin of the Proto-Atlantic at Laisvall (Bjorlyyke and Sangster, 1981). The L'Argentière deposit and related occurrences are found in a detrital terrigenous complex (Samama, 1976) and the Moroccan deposits occur in a sequence of coarse arkosic detrital sediments.

Sandstone-Copper (SSCU) and Shale Base-Metal (SHBM) Deposits.

Sediment-hosted copper deposits are fairly common, contributing about 30% of the Earth's copper reserves (Jacobsen, 1975). Production is dominated by two major cupriferous provinces, the Zambian Copperbelt of Upper Proterozoic age and the German-Polish Kupferschiefer of Lower Permian age. About one-third of the deposits are hosted by sandstones and the remaining two-thirds by calcareous shales (Edwards and Atkinson, 1986). They range from Precambrian to Recent in age and the most important deposits are those lying in Late

Proterozoic and Late Palaeozoic rocks which correlate closely with widespread desert environments of sedimentation (Kirkham, 1986).

Both SSCU and SHBM deposits are considered in this section as a number of the ore deposit examples used in this thesis may be classified into either group e.g. the Kupferschiefer, White Pine, Michigan and the Zambian Copperbelt. Gustafson and Williams (1981) noted that in some of the established SHBM deposits (e.g. the Polish and East German Kupferschiefer) the majority of copper is actually sandstone-hosted and about 40% of the ore in the Zambian Copperbelt is also sandstone-hosted.

A list of five varieties of sedimentary copper deposits was presented by Maynard (1983) and it illustrates the problem of classification of these copper deposits. In the following list the classes into which the individual examples have been placed in this thesis are given in brackets to show the disparity between this classification and that chosen by Maynard.

Type (Maynard)	Type (Grainger)	Example
Supergene	Not included	Chincarilla, Chile
Epigenetic in Sandstone	SSCU	New Mexico (SSCU)
Red-Bed Evaporite	SSCU	Creta, Oklahoma (SHBM)
Epigenetic in shale	SHBM	White Pine, Michigan (SSCU)
		Nevada Black Shales (SHBM)
Controversial	SHBM &/or SSCU	Kupferschiefer (SHBM, SSCU)
		Zambian Copperbelt (SHBM, SSCU)

Shale Base-Metal Deposits (SHBM)

A number of examples of SHBM deposits are briefly described below to illustrate the diversity of deposits included within this group.

Johnson (1976) reported on stratiform Permian copper deposits of 0.5-4.5% Cu grade in southwest Oklahoma hosted by laminated silty shales. The major examples are at Creta and Mangum where the mineralisation (chalcocite and malachite) occurs in two horizons within the Flowerpot Shale. The host rocks form part of a near-shore red-bed evaporite sequence with the highest copper concentrations found in grey shales immediately overlain by gypsum deposits (Hagni and Gann, 1976). Smith (1976) considered that most of the copper-bearing shales had been deposited on tidal flats.

Poole and Desborough (1981) described a black shale facies in Ordovician and Devonian rocks of Nevada consisting of mudstone, siltstone, chert and minor carbonate strata. These beds are highly metalliferous with up to 5000ppm vanadium, 350ppm selenium, 10ppm silver and 500ppm chromium residing in organic matter - the shales contain up to 20 weight % organic carbon. They have a high molybdenum content (1000ppm) which is generally found as molybdenite or in organic matter. Zinc (1800ppm) occurs in sphalerite. Many of the metalliferous beds are oil shales and it is assumed that anoxic environments of deposition and diagenesis were necessary for the accumulation of such significant metal and oil concentrations. These host shales were deposited on a continental rise along the western margin of Palaeozoic North America.

'Kupferschiefer' is the stratigraphic name given to a mineralised marl which covers most of northern Europe extending from England to Poland. It is extremely thin (generally less than 1m) and is rich in lead and zinc as well as copper in contrast to the Permian shales of Oklahoma described earlier. The Lower Zechstein shales of southwest Poland host lead mineralisation (Haranczyk, 1970) whereas the Kupferschiefer in other deposits in the Sudetic Foreland is copper-bearing (Preidl and Metzler, 1984). Copper concentrations higher than 0.3% are thought to be mainly restricted to near-shore regions of the Zechstein Sea, whereas zinc concentrations are more abundant and occur in a belt some distance from the coast. Lead is intermediate in abundance and occurrence (Wedepohl et al, 1978). They suggested that a considerable distance from the former shoreline (> 150km) the metal accumulations are more typical of black shales with reasonably high V, Mo, U etc. in bituminous marls low in Zn, Pb and Cu.

The Rotliegendes sands comprise the units beneath the Kupferschiefer. Where these sands are red they are known as Rote Faule - a term which is generally used to describe the red colouration of rocks upwards to the Werra Anhydrite (Maynard, 1983). The Rote Faule contains varying quantities of haematite (where the Fe^{3+} ion is responsible for the colouring) and it is notable for its extreme deficiency in copper (Jung and Knitzschke, 1975). In areas where the Rotliegendes is white (due to the presence of the Fe^{2+} ion, or the iron has been removed by leaching) it is referred to as the Weiss Liegende and copper mineralisation is prevelant. For example in certain areas of the Kupferschiefer of Poland economic concentrations of base metals are present in white sandstone (Rotliegendes; 50% Cu), black shale (Kupferschiefer; 20% Cu) and in the base of the overlying carbonates (Zechstein limestones; 30% Cu) - Kucha and Pawlikowski, 1986.

The Zambian Copperbelt can be divided into two sub-provinces, Northern Zambia and Southern Shaba. In Zambia the predominantly clastic host rocks are metamorphosed from chlorite to garnet grade and are often highly deformed (Fleischer et al, 1976). The sequence was deposited over a much rougher palaeotopography than the Zairean succession and the sediments unconformably overlie prominent granite hills, some of which are capped by stromatolitic bioherms (Garlick, 1981). The Zairean sub-province has a mainly dolomitic lithology, the host rocks being unmetamorphosed and relatively flat-lying although they are cut by numerous thrust faults. Both regions show evidence of hypersalinity during deposition; magnesite is common in Zaire dolomites with anhydrite common in Zambia. Annels (1974) considered this to be important in ore genesis. The deposits of the two provinces have similar mineralogy with copper and cobalt sulphides abundant while lead and zinc are conspicuously rare (Bartholome, 1974).

Within the Zambian province mineralisation is hosted by two different lithologies one of which constitutes the "ore shale trend" to the southwest of the region with mineralisation hosted by fine grained clastics (SHBM deposits). For example the Nchanga area where a marine transgression over conglomerates and arenites is manifested by a black, carbonaceous shale. At Nkana North Limb the mineralisation lies within siltstones, argillites, dolomitic argillites and argillaceous dolomites (Annels, 1974). Evaporites, particularly gypsum and anhydrite, are frequently found in close proximity to the ore-bearing horizon. Sedimentary structures such as dessication cracks are also found in many areas e.g. Chibuluma (Edwards and Atkinson, 1986).

SSCU deposits are found around the Mufulira region where three stratabound ore bodies are hosted by a fine grained carbonaceous wacke interbedded with barren dolomites (Van Eden, 1974). The copper-bearing sediments of this deposit type sometimes have a significant carbon content and, in keeping with this, the graphitic carbon content of the Mufulira greywackes varies from 1 to 2 % (Annels, 1979). Algal structures, such as those of the cupriferous dolomitic shale of the Copperbelt, are also characteristic and they have been found as biostromes and bioherms in the Mufulira deposits (Garlick, 1981).

Sandstone-hosted Copper Deposits (SSCU)

In addition to the group of SSCU deposits which could also be classed as SHBM deposits there is a group which appears to be similar to SSUV roll-front deposits. They generally tend to be confined to shallow lagoonal or lacustrine environments restricted to continental basins as described by Caia (1976) for Lower Cretaceous deposits of Africa. One region where these deposits occur has been described in detail and other regions mentioned briefly.

There are numerous small copper deposits associated with fluvial and transitional marine environments of Permian age in New Mexico e.g. the Scholle and High Rolls districts, USA (Lapoint, 1976). Copper-bearing minerals (typically chalcocite) replace woody, organic debris in fluvial sandstones which represent small channel deposits. Those in transitional marine environments are found as a replacements of algal mats and mineralisation occurs near oxidation fronts showing transitions from red to grey sediment. The Triassic deposits of New Mexico (e.g. Nacimiento Mine) are much larger and less arkosic than those of Permian age and copper-bearing iron sulphides in the Chinle Formation have partly replaced, or are closely associated with, carbonaceous fossil material, mainly logs (Woodward et al, 1974). Red-beds of early diagenetic age are always associated with these SSCU deposits (Lapoint, 1976) but the genetic relationships between the deposits in New Mexico and those, such as the Kupferschiefer, are not known (Lapoint, 1986).

The mineralisation in Lower Cretaceous conglomerates, sandstones and siltstones is thought to be controlled by palaeochannels of variable size and is closely associated with residual organic matter (Caia, 1976) as described above for occurrences in New Mexico. The late Cretaceous to mid-Tertiary deposits of the Corocoro Basin, Bolivia are also characterised by the replacement of organic material by copper-bearing minerals (Entwistle and Gouin, 1955) and their distribution and orientation have been compared to those of SSUV deposits in the past (Ljunggren and Meyer, 1964). The last group of deposits of this type used as examples in this thesis are the copper-silver occurrences in Carboniferous strata of southeast New Brunswick and northwest Nova Scotia. They are found in grey channel deposits and grey-green interbeds in the mainly arkosic Hopewell Group and the presence of organic debris again appears to be the controlling influence upon mineralisation (Van de Poll, 1978).

Finally, a few more examples should be added to the list of SSCU deposits which appear to be related to SHBMs, especially with reference to their mode of origin. Renfro (1974) proposed a sabkha (intertidal and supratidal) model for the depositional environment for these deposits to explain the sedimentary associations of red-bed - dark shale with copper - evaporites that are frequently seen. This model has been accepted by Smith (1976) for the Permian deposits of Texas and by Chartrand and Brown (1984) for Coates Lake deposit, Redstone area, NWT, Canada.

Sedimentary Exhalative Deposits (SDEX).

Large sediment-hosted lead-zinc deposits form a distinctive group characterized by stratiform, syngenetic ores. This group has been discussed in relation to SSCU and SHBM deposits by many authors who suggested the entire group were varieties of one class incorporating all sediment-hosted stratiform deposits of copper, lead and zinc (e.g. Morganti, 1981 and Gustafson and Williams, 1981). However other authors such as Badham (1981a) and Large (1980 and 1981) considered SDEX deposits as a separate group. Only deposits with dominant lead-zinc mineralisation have been included in this class. Those which possess more copper than lead and zinc and are hosted by argillaceous rocks have been classed as SHBM deposits and those hosted by coarser clastics have been classed as SSCU.

SDEX deposits are of restricted geographical and temporal distribution and there appear to be two main periods of SDEX mineralisation, the Middle Proterozoic (1400-1600 m.y.) and the Lower Middle Palaeozoic (320-500 m.y.). The Proterozoic deposits are generally larger than those of Palaeozoic age (Large, 1981).

The general characteristics of SDEX deposits have been presented by a number of authors e.g. Large (1980 and 1981); Gustafson and Williams (1981); Morganti (1981) and a summary of the salient features of SDEX deposits given by Large (1981) is listed here with additional references.

1) Stratiform sulphides occurring in one or more lens-like bodies e.g. McArthur River, Russell et al (1981).

2) Stratiform sulphides are concordantly interbedded with marine sediments of different lithologies e.g. Meggen and Rammelsberg, Krebs (1981).

3) Stockwork or vein-type mineralisation is common e.g. Tom deposit, Carne (1976); Meggen, Krebs (1981).

4) Distinct lateral and/or vertical zonation of sulphides is often seen e.g. McArthur River, Russell et al (1981); Meggen and Rammelsberg, Krebs (1981).

5) Sulphides may be in massive and laterally persistent beds e.g. Rammelsberg and McArthur River, Large (1981).

6) Characteristically simple mineralogy : pyrite and/or pyrrhotite, sphalerite, galena, minor chalcopyrite; marcasite and arsenopyrite occasionally also present. High Fe and Ag sulphide contents common e.g. high Ag in Rammelsberg, Mount Isa and Lady Loretta (Large, 1981).

7) Barite often overlies mineralisation or occurs as a lateral stratigraphic extension e.g. Chaudfontaine, Dejonghe (1979).

8) Thin tuffite horizons are found within host rocks in many deposits e.g. Mt. Isa, Lady Loretta, McArthur River, Large (1980); Kanchanaburi Province, W. Thailand, Diehl and Kern (1981).

9) Many deposits are spatially associated with a growth fault e.g. Mt. Isa, Tynagh, Large (1980); Central Pyrenees deposits, Pouit (1978).

10) Host rocks are mostly shales (which may be carbonaceous), siltstones and to a much lesser extent, carbonates e.g. Irish deposits, Russell et al (1981).

11) Many are associated with evaporite sequences, massive limestones or have evidence of a past arid climate e.g. Mt. Isa, Lady Loretta, Pyrenees, McArthur River, Large (1980); Kanchanaburi Province, W. Thailand, Diehl and Kern (1981).

12) Mineralisation is associated with carbonaceous matter in many deposits e.g. Dzhezkazgan, Meggen, Mt. Isa, Large (1981); Lady Loretta, Loudon et al (1975).

SDEX are not classed with LSBM deposits as they are considered to be syngenetic in nature. They have an early timing of mineralisation relative to deposition of host rocks, a greater conformity of mineralisation with the hosts (lacking in LSBM deposits) and higher iron and silver contents.

Carbonate rocks hosting the Irish deposits of Lower Carboniferous age are included in the SDEX class. It was thought by workers such as Large (1981) that they are characterised by submarine exhalative features e.g. their conformity with host rock stratigraphy (Sangster, 1976). However it must be noted that deposits such as Silvermines have been regarded (Taylor and Andrew, 1978) as partly exhalative (Upper G zone) and partly replacive (Lower G ore zone). Also the main part of the Tynagh orebody has been considered to be replacive but the iron formation and surrounding manganese halo have been accepted as clear evidence of syngenetic precipitation (Badham, 1981a). The fact that Irish deposits possess some characteristics which may be indicative of a secondary sedimentary origin e.g. a dominant cross-cutting mineralisation, has led to the notion that these deposits may represent a transitional type between SDEXs and LSBMs (e.g. Edwards and Atkinson, 1986).

The Irish SDEX deposits differ from another controversial type, the Alpine LSBM deposits, in having a more complex mineralogy with the addition of silver and copper to the extractable metals. They also have a distal, tuffaceous Fe-Mn deposit which is lacking in the Alpine deposits.

Although Proterozoic deposits are not within the scope of this research it is worth noting that many of the most important SDEX occurrences are Proterozoic in age (Gale, 1983 and Lambert, 1983).

Limestone Base-Metal Deposits.

Carbonate-hosted strata-bound lead-zinc deposits make up some of the world's major ore deposits and have long been considered as a single class, the Mississippi Valley type, named from the region where some of its principal examples are located. Limestone base-metal deposits mainly occur in the Tri-State, southeast Missouri and Upper Mississippi Valley districts of the United

States and some Alpine and Silesian areas of Europe. As a group they exclude skarns and other replacement deposits e.g. the lead-zinc bodies at Bingham, Utah which may be related to nearby intrusions (Sangster, 1976).

These deposits are characteristically of Phanerozoic age. Both Stanton (1972) and Sangster (1976) mention the paucity of LSBM sulphides in Proterozoic rocks. This is particularly significant in view of the relatively long time range of this era.

LSBM deposits are stratabound and generally thought to have been emplaced after lithification of the host rocks, mineralisation having been largely controlled by pre-ore structures. The ores normally consist of sphalerite, galena, fluorite and barite in varying proportions and contain minor pyrite, Sb, Cu, Ag and Cd in dolomitic carbonates (Badham, 1981a). The common characteristics of LSBM deposits are given by Jackson and Beales (1967), Heyl et al (1974) and Sangster (1976). Only a brief summary of their salient features is given here i.e. those points which have been used to classify examples as LSBM deposits. These are:

a) The host rocks commonly form part of a thick carbonate pile, close to facies fronts with basinal argillaceous rocks (Badham, 1981b) and with a close spatial association to evaporite sequences.

b) They consist primarily of bedded replacements and veins. The ore fills former open spaces that took the form of karst cavities, high porosity zones in bioherms, or faults (Maynard, 1983).

c) There is an association between ores and dolomitization, although the relationship of these two events with time has not been satisfactorily established (Kyle, 1981; Bass-Gustkiewicz et al, 1982).

d) Many of the deposits are known to be associated with petroleum, oil and H_2S (sour gas) which are thought to have a control on the environment of deposition of the metals (Dozy, 1970; MacQueen and Thompson, 1978).

e) Their mineralogy is simple and the precious metal content is low i.e. galena, sphalerite, pyrite, fluorite with minor Co, Ni, Ag, Cu, Cd, In, Ge, Ga (Kyle, 1981).

f) There is a general absence of igneous rocks as potential sources of ore solutions (Jackson and Beales, 1967).

g) Ore is frequently related to positive structures including basement "knobs", calcareous sand banks and algal reefs (Sangster, 1976).

h) Solution activity, brecciation, slumping, collapse and thinning are common (Ohle, 1985).

There has been much debate over the classification of deposits in the Alpine Mesozoic Geosyncline of Central Europe known as the Alpine-type deposits (Brigo et al, 1977). They possess many features generally considered to be epigenetic e.g. vein fillings, breccia cementation and massive replacement bodies. However some authors such as Sangster (1976) considered Alpine-type deposits to be more akin to SDEXs than LSBMs with regard to their form relative to the host rocks e.g. lenticular bodies containing low grade ore are found in various host rocks. Despite this they have been classified as LSBMs here in agreement with Large (1976) who assessed that the deposits do not share many of the features which are generally accepted as characteristic of submarine exhalative mineralisation. He interpreted many of the syngnentic features as deposition of sulphides in karst and large solution cavities e.g. bedded ores at Bleiberg have been shown to be cave-filling and are clearly epigenetic.

3.3 New Approaches to the Problem

Eugster (1986) presented a new approach to the problem and suggested that copper-lead-zinc deposits could be classified with respect to their geochemical environment as represented by the pH, the temperature of the mineralising fluids and the nature of the host rocks. He proposed four major groups of deposit types, each of which is described briefly below, with examples;

a) MVT. Mississippi Valley Type. Hot acid brines, carbonate host rocks. e.g. Tri-state District, USA. Classified as Limestone Base-Metal (LSBM) deposits here.

b) KST. Kupferschiefer Type. Cool acid brines, shale host rocks. e.g. Kupferschiefer; Zambian Copperbelt; Creta, Oklahoma, USA; Cretaceous deposits of Angola. Classified as Shale Base-Metal (SHBM) deposits here.

c) GRT. Green River Type. Alkaline solutions, lacustrine environment. e.g. McArthur River; Mt. Isa; Dugald River. Classified as Sedimentary Exhalative (SDEX) deposits here.

d) SHT. Sandstone-hosted Type. Placed in a class related to KST. Classified as Sandstone Copper (SSCU) deposits here.

Eugster admitted that much more work needed to be done to refine this approach which is still in its infancy as it required a reasonably large number of deposit groups to accomodate all the different variations upon a deposit type. The classification used in this thesis bears more than a passing resemblance to the groups of Eugster. It is still, in part, based on the temperature of the metal-bearing solutions i.e. cool (primary deposits) or hot (secondary deposits) and the nature of the host rock to the mineralisation. The similarities continue in that some individual deposits have been placed into equivalent classes (although the titles of those classes are different).

Brown (1986) pointed out that efforts to determine the genesis of stratiform sediment-hosted copper deposits are frustrated by the improper usage of terms when describing mineral deposits. He inferred that when workers allude to a type locality, such as 'the Kupferschiefer-type' of Eugster, it is inadequate to do so when a particular deposit is dominated by features not seen at that type locality as this is potentially misleading and confusing to other workers.

3.4 Omissions

A number of mineral deposit types shown in the classification in Table 3.1 have not been considered in the thesis. These are given in the following sections, with reasons for their omission from the discussion.

3.4.1 Paucity of examples.

MNNO	manganese nodules	CALU	calcrete uranium
GOSS	gossan	SUPE	supergene enrichment
SULF	sulphur	CEVA	continental evaporites

The mineral deposit types listed above have not been included because too few examples were collected during the initial literature search.

3.4.2 Marine Evaporite Deposits (MEVA).

MEVA deposits are those whose origin can be directly related to processes of precipitation and recrystallization from saturated solutions. There is a great deal of literature on the subject of the palaeolatitude of evaporite deposition (e.g. Irving and Briden, 1962; Briden and Irving, 1964; Gordon, 1975; Meier, 1981) and MEVA in general (e.g. Stewart, 1963; Davison, 1966; Sonnenfeld, 1984) so only a brief precis of the considerable work which has already been carried out will be given here.

Most modern evaporites are accumulating within fifty degrees of latitude north and south of the equator (Gordon, 1975) and the distribution within this belt is bimodal. Irving and Briden (1962) stated the full range of MEVA occurrence is from 31°S to 83°N, but that 73 % of deposits lie between 30° to 60° N (Figure 3.2). With regard to this particular phenomenon, evaporite deposits therefore have a special significance as palaeolatitudinal indicators because of their limited distribution. From this palaeoclimatic conditions can

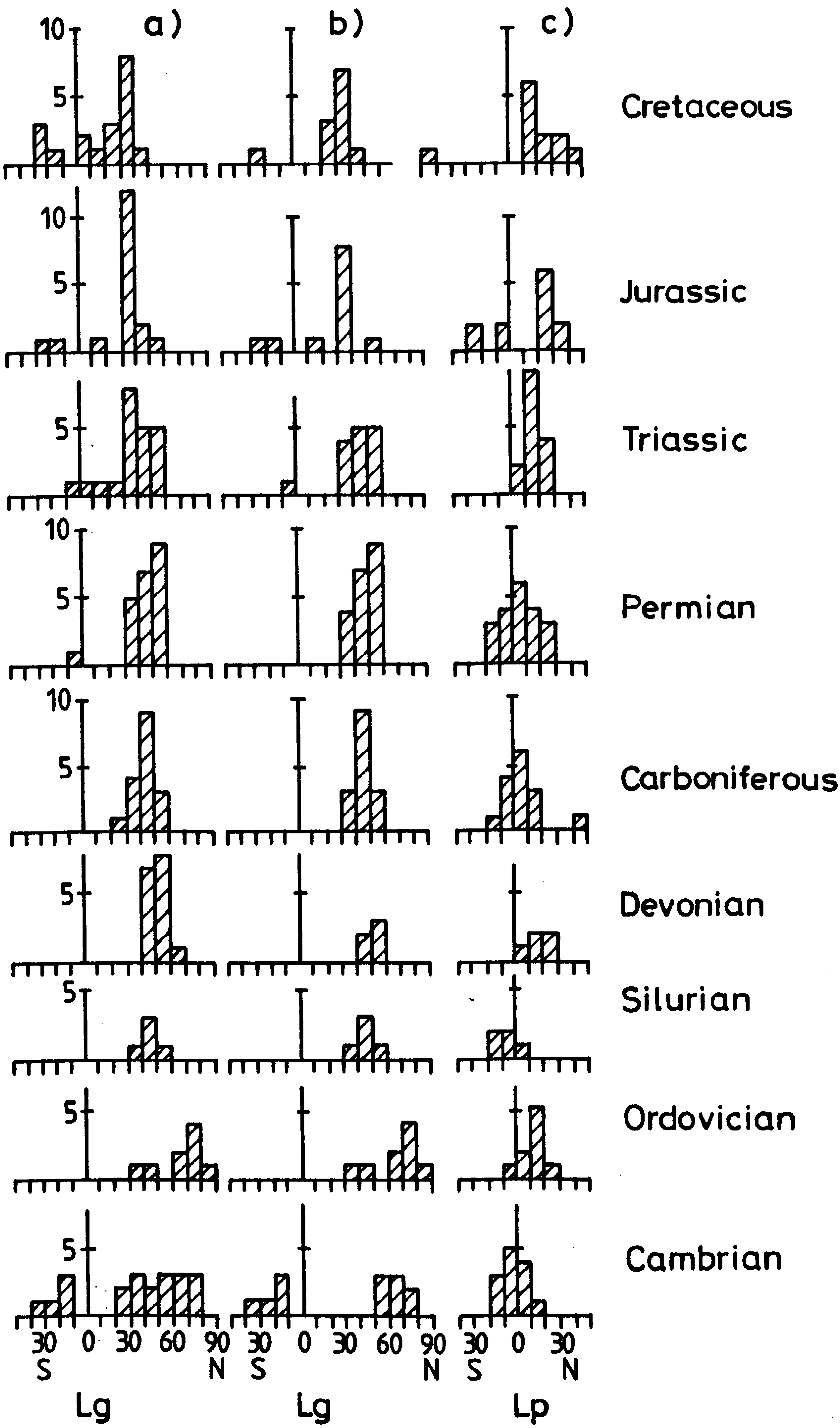


Figure 3.2. Column a)=present latitude (Lg) of all evaporite deposits. Column b)=Lg of all evaporite deposits for which palaeomagnetic data are available. Column c)=the palaeolatitudes (Lp) of deposits in column b. (After Irving and Briden, 1962, Figure 2).

be inferred due to restricted conditions under which evaporites develop. Although a number of local factors can cause excessive evaporation, atmospheric circulation has an overwhelming influence on the global distribution of modern evaporites and there is no reason to assume that it was otherwise for ancient evaporite deposits - assuming, of course, the principle of uniformitarianism (Gordon, 1975; Ziegler et al, 1984). (See also Chapter Four, section 4.6.1).

When ancient evaporites are analyzed in terms of latitude of formation their original distributions are found to follow the present day model very closely, at least as far back as the Permian, and possibly to the beginning of the Phanerozoic (Irving and Briden, 1962; Gordon, 1975; Meier, 1981). There is a pronounced asymmetry in evaporite occurrence during later Phanerozoic times, with more prominent deposition in the northern hemisphere attributable to the greater proportion of land mass relative to the southern hemisphere during this time (Meier, 1981). However these results may also be influenced by better preservation of, and more widespread data on, younger rocks. In contrast, Gordon (1975) noted that during earlier Phanerozoic times there was more prominent evaporite deposition in the southern hemisphere, again a reflection of the greater concentration of land. But in both cases the land area per hemisphere has been deduced from palaeomagnetically guided reconstructions with their inherent assumptions and errors. Therefore such interpretations of the data must be regarded with some caution.

The Permian to the Present

Gordon (1975) emphatically stated that evaporites developed between latitudes 50°N to 50°S during the period from the Permian to the present. Only one percent of Triassic and two percent of Cretaceous occurrences are exceptional and extend respectively to about 52° to 54° from the equator at that time. Meier (1981) noted that during the Triassic to Jurassic, the main evaporite belt lay between 20° and 40°N and concluded that from the beginning of the Jurassic the northern evaporite main belt was positioned south of the 50°N latitude. A zone of low evaporite frequency lay at, or near, the equator and the majority of deposits are distributed as follows (Gordon, 1975).

0° - 10° = 13.5% : 10° - 20° = 31.4% : 20° - 30° = 24.2% .

Cambrian to Carboniferous.

For this period Gordon (1975) found that deposition occurred within 50° of the equator with the exception of about 2% of Devonian evaporites which were laid down between 50° and 53° from the equator and 16.5% of Carboniferous evaporites deposited between 50° and 62° from the equator. Cambrian to Ordovician results are unusual in that 4% of occurrences were south of 80°. However, Meier (1981) also admitted difficulty in producing a useful interpretation for the Cambro-Ordovician part of his histograms and found latitudes limiting Palaeozoic salt deposits ranged from 80°N to 40°S. According to Meier (1981) the main maximum of salt deposits is found between 30° to 50°N, whereas the majority of evaporite deposits occur in a zone between 30°N and 70°N from the Ordovician to the Permian.

In general similar conclusions can be drawn from the Cambrian to the Carboniferous and the period from the Permian to the Recent, although the southern hemisphere belt does not stand out so prominently as the northern one. The weakness of these southern belts tends to obscure the equatorial low-frequency zone in the evaporite distribution which is a distinct feature (Gordon, 1975). The poorer definition of the pattern is probably due to poorer quality palaeomagnetic data and uneven land distribution between the northern and southern hemispheres.

3.4.3 Proterozoic Deposits.

PAPL palaeoplacer (Au - U)

FEFM iron formation

SHUR shale uranium

The mineral deposit types listed above are all from the Proterozoic era and as such should not be covered in the scope of this project.

3.5 Conclusion

Finally as every author seems to develop his own particular classification, it would be hoped that such a subjective approach would have no influence upon the conclusions drawn from this research. When the project was originally formulated one of the aims was to assess whether or not it is possible to clarify some of the confusion that exists around the classification of deposits by classifying them by their apparent palaeolatitude of formation. If it is not possible to classify deposits in such a way this thesis will be independent of the debate about the origins of these deposits and so the classification will not actually matter. However, if different groups showed different results, then this would have implications for the genetic debate. Initially, however, it was not possible to predict such an outcome so the classification was an important part of the ground work done at the onset of the project.

CHAPTER FOUR

BASIC ASSUMPTIONS

The basic principles of palaeomagnetism have been outlined (Chapter Two) and the classification of mineral deposits has been discussed (Chapter Three). It is now appropriate to examine the assumptions concerning the Earth and its dynamics which were made at the onset of the research as any subsequent conclusions will be drawn in the knowledge of the caveats that follow. These factors have to be assumed before the relationship between palaeolatitude (and so palaeoclimate) and the formation of mineral deposits can be assessed.

4.1 Uniformitarianism

Gould (1967) divided uniformitarianism into two separate varieties, substantive and methodological. The former demanded a "uniformity of material conditions or of rates of processes" which he dismissed as untrue and so ignored. Methodological uniformitarianism comprised the two assumptions;

a) "natural laws are constant in space and time and

b) no hypothetical unknown processes can be invoked if observed historical results can be explained by presently observable processes."

These assumptions were considered by Gould to be ineffective as they were too simplistic and based on induction.

Uniformitarianism in its methodological form is still used today and has been assumed in this thesis in the absence of evidence to the contrary. This is not an entirely naive approach as the concept has been frequently applied to palaeoclimatology and "the present as the key to the past" has been widely used as the basic premise upon which a range of research has been based. Some degree of success has been achieved in climatic modelling using this method, for example Barron (1985) and Scotese and Summerhayes (1986). However Frakes (1979) considered that while the principles governing the motions of the atmosphere and oceans must have been extant since the Proterozoic, the gross physical changes in the Earth and possibly the Sun, demand that rates of heat flux have been characterised by change. He argued that the evidence for that change existed in the abundantly fluctuating nature of the palaeoclimatic record and hence the Uniformitarian Principle had been invalidated. It is appreciated that questions can be raised as to the validity of using this principle so the factors that may temper its usefulness are discussed later. However it has been assumed here that the principles have remained constant throughout the Phanerozoic although their effects probably have not.

4.2 The Earth as a Geocentric Axial Dipole.

The conclusions which can be drawn from any application of palaeomagnetic methods to determine palaeolatitudes and the relative displacement of different tectonic blocks must be dependent, at least in part, upon the model for the average geomagnetic field which was initially adopted. In this case the generally accepted hypothesis that the time-averaged palaeomagnetic field is that of a geocentric axial dipole (GAD) has been adopted. To determine this field the actual variation in the geomagnetic field must be averaged for a length of time comparable to any secular variations that may have occurred. It is known that this model is probably invalid during polarity transitions, or attempted transitions, as the dipole component diminishes at this time (Park, 1983; Tarling, 1985) but such intervals are usually eliminated from the data used to calculate the pole positions.

The simplest model for the Earth's present geomagnetic field is a dipole field. This hypothetical dipole is geocentric, but inclined at about 11.5° to the geographic axis so the geomagnetic and geographic poles do not coincide at the present day. Nonetheless this model has been assumed for progressively earlier time periods. The wisdom of accepting that this model for the Earth throughout the Phanerozoic can be tested in a number of ways.

1) The geocentric dipole aspect of the model can be examined by showing the consistency between pole positions calculated with such an assumption and those without (Tarling, 1983), but it is more difficult to test for the axial nature of the model (see (3) below). It has been shown that good agreement exists among poles from rocks of roughly similar age across continental areas, indicating that the model is correct back to at least the Carboniferous period (Morel and Irving, 1981). They argued against the use of a non-dipole geomagnetic field to explain possible discrepancies in apparent polar wander paths for the latest Carboniferous to the early Triassic.

2) World-wide field measurements relating to the last few tens of millions of years, when plate motions can be considered to have been insignificant, may be used to authenticate the GAD model (Tarling, 1975). Park (1983) felt the Earth's field of the past 20 million years approached the GAD model when it had been averaged over periods of several thousand years.

Smith (1981) considered a number of widely separated sites less than 7,000 years old and plotted the palaeomagnetic north poles so determined. They were grouped around the geographic pole rather than the present geomagnetic pole and so demonstrated that for the past 7,000 years at least the Earth's magnetic field had been mainly dipolar. But in order to calculate the pole positions from declination and inclination measurements the initial assumption of field dipolarity was essential. Circularity in this argument was avoided by the fact that the dipole assumption led to a close grouping of the palaeomagnetic poles and so proved the point on grounds of consistency. If the field had not been mainly dipolar for that period of time the calculated poles would have been scattered all over the globe. However calculated poles from terrestrial igneous rocks erupted during the past 25 million years have shown a tendency to occur on the far side of the present rotational pole and also slightly to the right of it (Wilson and Ade-Hall, 1970; Wilson, 1970 and 1971). This can be modelled by a northward displacement of the dipole 200 to 300 kilometres along the Earth's rotational axis. This departure must be accepted for it has been determined from rocks of vastly different ages e.g. 2 and 30 million years, but the age range is unlikely to have been allowed for significant plate motions. But it does only correspond to a net shallowing of the inclination at the locality by 1-2° relative to the expected dipole inclination. The observed dipole off-set is therefore small and it is thought by some (e.g. Tarling, 1983) that it does not seriously detract from the GAD as a working hypothesis and could be due to a slight, but systematic, shallowing of the inclination in palaeomagnetic studies.

3) Comparison of palaeolatitudes calculated on the geocentric axial dipole model with those indicated by other palaeolatitude estimators (Briden and Irving, 1964; Tarling, 1975) also provides a test for the GAD model. The consistency of many palaeoclimatic indicators with palaeomagnetically determined palaeolatitudes suggests that the average field has been axial, probably within 5° to 10° for most of geological time (Tarling, 1983). Drewry et al (1974) were convinced of the GAD model for the Earth's field for Permo-Triassic times by comparison of the dipole axis with independent estimates of the spin axis. Their results were obtained from the long-term positions of the Earth's climatic belts, which in turn were inferred from the distribution of climatically-controlled sediments. They concluded that the excellent agreement between the expected positions of such climate-sensitive sediments as evaporites and tillites and their inferred positions shows that, to a first approximation, the Phanerozoic geomagnetic field had been a geocentric dipole whose axis was coincident with the Earth's rotation axis. However if the position of the axis had changed relative to the ecliptic the palaeoclimatic

significance would be less. For example if the pole axis was in the equatorial plane then six months summer and six months winter would be experienced throughout the globe.

The GAD assumption is unaffected by the possible presence of true polar wander (TPW) e.g. the relative movement of the lithosphere as a whole with respect to the mantle, or of the mantle relative to the core. Such movements would affect equally all the calculated poles of a given age and so would not affect geological conclusions based on these poles (Park, 1983).

To summarise, it has been assumed that the GAD model for the Earth's geomagnetic field is acceptable for the Phanerozoic. However this assumption has been the crux of this thesis as all the palaeomagnetic methods used have been based upon it. If this model were proven to be incorrect, then all conclusions drawn from these methods would be invalidated or would require re-evaluation. It is possible that the GAD model is invalid at some times, but this has been assumed to be sufficiently rare as not to effect the conclusions.

4.3 Uniformitarianism and the Earth's Heat Budget.

The solar energy input to the Earth's atmospheric system is not constant as daily and seasonal cyclic variations in incident radiation are known to occur. Variations extending over periods of years e.g. sunspot and Milankovitch cycles have also been recognised. The latter are thought to be the best known causes of climatic change over thousands of years (e.g. Herbert and Fischer, 1986). Although the sun's radiation was not necessarily constant over geological time, it is considered to be sufficiently invariable to be included in a climatic model in this thesis. From this model the energy transfer necessary to explain the Earth's present climate can be inferred (Frakes, 1979). The transfer of this heat energy from the Earth to the atmosphere and then from the equatorial zone to high latitudes has probably varied only within narrow limits throughout geological time. To support this assumption concerning the solar constant, in agreement with Bambach et al (1980), it is argued that as the Sun is a stable main-sequence star so it is not unreasonable to assume that its radiation intensity has not altered greatly in the last 600 million years. (However a variation of +/- 10% is probable, Tarling, pers. comm.). Frakes (1979) held a similar view and considered the radiant energy of the Sun to have been maintained, within relatively narrow limits, near its present level since the beginning of the Phanerozoic. This is shown by similarities of the earliest life forms to modern species. If it is assumed that the Earth has always been spherical, that its axis of rotation has been approximately perpendicular to the ecliptic and that it has always orbited at a similar distance from the Sun then the heat budget of the earth can be regarded as essentially fixed. This heat budget is characterised by excess heating in the equatorial region, heat transfer by atmospheric and oceanic circulation towards the poles, and cooling in the polar regions. Although the Earth's present climate is probably atypical of past geological periods, particularly those with vast expanses of shallow seas and no ice sheets, the equator-to-pole temperature differential which drives the atmospheric circulation must always have been in existance. So the rainfall belts have probably remained at relatively constant latitudes while surface temperatures probably have not (Habicht, 1979; Ziegler et al, 1981). Some workers (e.g. Tarling pers. comm.) believe that if no ice sheets were present the rain belts would have a different latitudinal extent, although latitudinal zoning would remain a feature.

In conclusion the heat budget of the Earth, and its distribution through the circulation patterns, have been assumed to have been approximately constant throughout the Phanerozoic. However it does not necessarily follow that climate zones have also remained constant during that period. Nonetheless, it still seems probable that climatic latitudinal zoning existed although the precise latitudinal extent of climatic belts may have varied significantly. The influences upon climate of geography and changes in the circulation patterns are also very great and will be discussed in section 4.5.

4.4 Uniformitarianism and the Earth's Rotation.

The constancy of the Earth's obliquity and its speed of rotation, have already been used to question the validity of accepting the Uniformitarian Principle. It has been assumed that neither have affected atmospheric and oceanic circulation patterns (and hence climatic regimes) too drastically despite the fact that they must have altered slightly during the Phanerozoic e.g. the Earth's rotational period was probably some 400 days per year 400 m. y. ago. The reasons for these assumptions have been given below.

4.4.1 Rotation.

It is thought that the Earth's rotation was faster in the past (Wells, 1963; Scrutton, 1965 and 1978; Mohr, 1975; Rosenberg and Runcorn, 1975) and has slowed through time by tidal friction (e.g. Sundermann and Brosche, 1978). Scrutton (1978) suggested a rotational speed for the Palaeozoic some 15 to 20 % faster than it is today. The resultant stronger Coriolis force would have lead to a compression of the zonal pattern of circulation equatorward by about 4° to 6° latitude in each hemisphere. At a particular rotational speed more than three zonal cells would be required for the heat exchange between the equator and each pole to be effected. However Hunt (1979) suggested the rotational speed would have to be considerably higher than that postulated for the Palaeozoic in order for the present zonal pattern to break down. Scotese and Summerhayes (1986) suggested that an increase in the number of cells would only be produced by an increase in the Earth's rotational speed in excess of one and a half times its present rate.

The compression of the zones has probably not exceeded 20 % since the beginning of the Palaeozoic era (Ziegler et al, 1979) which equates to about 6° latitude assuming a Hadley cell of 30° width. As the error in positioning the continents on palaeogeographical reconstructions is probably 5° to 10° it was felt that little would be gained by trying to modify the zonal pattern for each of the Phanerozoic periods with this compressional factor. On this basis the present regime of climatic zones has been accepted here as a model for past periods.

4.4.2 Obliquity.

The atmospheric pressure system of the Earth derives from both the uneven incidence of solar radiation on the spherical Earth and the inclination of its axis of rotation. At present, this always points in the same direction in space, relative to the ecliptic, irrespective of the Earth's position in orbit, so the amount of radiation striking the outer atmosphere is variable both with latitude and time of year. Variations in the Earth's obliquity and other orbital parameters have been recognised by workers such as Milankovitch (1941), Hays et al (1976) and Berger (1976, 1978 and 1980) who acknowledged their potential influence on palaeoclimate. However cycles resulting from these variations are short-lived ($<10^4$ years) on a geological time scale and their effects may be difficult to recognise in the geological record largely due to imprecision in correlation (Parrish, 1982).

As there appears to be no physical or astronomical evidence why obliquity should have changed significantly it has been assumed here that the tilt of the rotational axis (at 23.5° to the orbital plane) was the same in the past as it is at present . However some workers have used large scale variations in obliquity to explain the apparently anomalous distributions of climate-sensitive lithologies. Wolfe (1978 and 1980) proposed an obliquity near to zero to explain the high latitude occurrence of some apparently warm climate plants in Eocene sediments while Williams (1975 and 1976) suggested an obliquity of 90° to account for Eocambrian tillites with an apparent low latitude distribution at the time of deposition. These appear to be drastic solutions to problems which may be explained in terms of distribution of continents, variations in palaeogeography and orography, or in the actual age of the sediments.

4.5 Climate.

It has been mentioned that all climates are a result of the way solar energy reaches the Earth and is distributed by the oceans and the atmosphere. These circulation systems are driven largely by the thermal gradient causing heat exchange between the equator and the poles. Climate can be described as the average weather conditions of a region throughout the seasons. It is expressed in terms of the means, extremes and frequencies of weather elements such as atmospheric pressure, temperature, humidity, rainfall, cloudiness, wind speed and direction. The climate of a particular area is governed by its latitude, its position relative to continents and oceans and by local geographical conditions. There are numerous local exceptions to predicted climates at certain latitudes as these climates can be modified by such factors as altitude and proximity to mountains.

Nairn (1973) distinguished three main climatic variations;

1) short term climatic fluctuations, about some mean value, attributable to astronomically controlled insolation differences.

2) longer term trends in mean climatic values (usually defined by temperature) e.g. the onset of glaciation brought about by the movement of a land mass to a polar location.

3) regional climatic changes that do not significantly alter the average global climate. For example the Cretaceous opening of the Atlantic profoundly modified

the climate in that region but may not have caused an appreciable variation in the climate of eastern Asia.

It is not the purpose of this chapter to investigate fully the complexities of weather and climate. General books on the subject include Sellers (1965), Lamb (1977), Critchfield (1983) and Houghton (1984). However, for the purposes of this thesis it is essential to evaluate any relationship that may exist between latitude and climate. A brief description of the present climatic regimes has thus been given, together with some explanation of atmospheric circulation patterns and the consequent climatic zonations.

4.5.1 Present Climatic Regimes.

On an Earth with a homogeneous surface, the wind systems would be wholly zonal i.e. parallel to latitude (Lorenz, 1967). This circulation pattern is shown in Figure 4.1, with hot air rising at the equator and cool air sinking at the poles. Complete direct heat exchange is prevented by the Coriolis force caused by the Earth's rotation which deflects moving air or water to the right in the northern hemisphere and to the left in the southern hemisphere. The meridional circulation shown in the diagram can be described as tricellular (Smith, 1981). The Hadley Cell comprises warm air rising from the equator which forms an equatorial low pressure zone known as the doldrums. This air cools and sinks near 30° latitude forming the subtropical high pressure belt. Air moving equatorward from this high pressure zone is deflected forming the Tropical Easterlies (the Trade Winds). The Ferrel Cell occurs in mid-latitudes and comprises air which moves poleward away from the mid-latitude high pressure belt forming the Westerlies. This cell is driven by frictional coupling between the Hadley Cell and the third cell. This is the weak Polar Cell of dry air sinking at the poles which forms a weak polar high pressure zone. Air flowing out of this zone constitutes the Polar Easterlies. Between these and the westerlies is a low pressure belt of rising air.

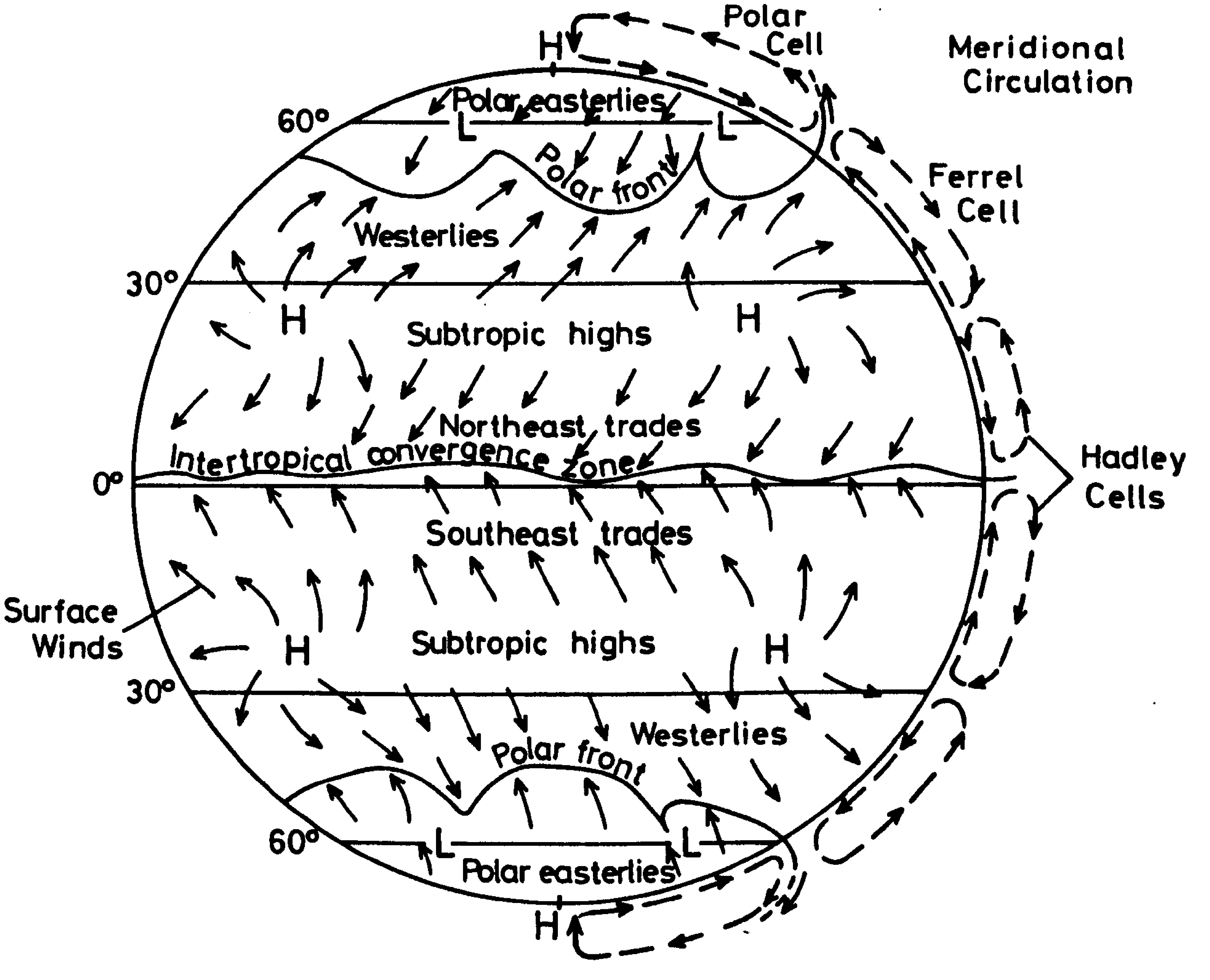


Figure 4.1. Schematic representation of the global wind and pressure belts. H=high pressure. L=low pressure. The effects of the differential heating of continents and oceans have been omitted. (Modified from Drewry et al, 1974, Figure 14).

This zonal pattern experiences slight seasonal shifts which displace the main pressure belts by approximately 10° latitude. Another important feature of the pressure system is the displacement of high pressure cells towards the eastern side of ocean basins. This results in important longitudinal variations in climate (Scotese and Summerhayes, 1986).

Oceanic and atmospheric circulation systems are related so it is to be expected that zonal oceans (i.e. those which girdle the Earth parallel to latitude) have zonal circulation; for example, the Circum-Antarctic Ocean. This is evident throughout present oceans although clustering of continents in the northern hemisphere obscures the pattern (Ziegler et al, 1981b).

4.5.2. Factors affecting Ideal Circulating Patterns.

The presence or absence of polar ice, the temperature at the sea surface, the equator-to-pole gradient and geography affect the transfer of heat from one place to another over the surface of the globe. The affect of regional and global geography upon climatic patterns has been described below in some detail. The geography of a particular period is vital to this research in view of its influence upon climate and in the determination of palaeolatitude.

4.5.2.1 Effects of Global Geography.

High and low pressure zonal belts may be disrupted as a result of the different thermal regimes of oceans and land masses. The high pressure systems centred at about 30° latitude are intensified over the oceans as water is cooler than the adjacent land at this latitude. The oceanic currents associated with these high pressure belts are composed of warm equatorial currents driven by the trade winds and the cool westerly currents at mid-latitudes. These are then linked to boundary currents adjacent to continents. The western boundary currents (e.g. the Gulf Stream) carry warm equatorial waters polewards and the eastern boundary currents (e.g. the Peru Current) carry cool water from mid-latitudes towards the equator. As a result the boundaries between the climatic zones occur at different latitudes on the opposite sides of oceans (Ziegler et al, 1981b). At the same latitude an ocean can be 10°C warmer on one side than the other (Parrish and Ziegler, 1980). For example, in July at a latitude of 20° south, the Atlantic is 23°C off eastern South America whereas it is 16°C off southwest South Africa.

A more extreme climatic asymmetry must occur when a large continental land mass exists e.g. during the late Permian when the presence of the supercontinent Pangaea resulted in an extraordinarily long fetch for the equatorial current. It is thought (Parrish, 1982) that when this current reached the western side of the Tethys it would have been forced into two Gulf Stream-type boundary currents flowing north and south. These would have carried very warm water into high latitudes, thereby creating a wide tropical zone on that side of Pangaea. The late Permian world was characterised by very great asymmetry, with tropical fauna extending to 40° to 50° latitude north and south in eastern Pangaea (Ziegler et al, 1981b) assuming correct interpretation of palaeomagnetic data for that period.

Low pressure systems are likely to form when a continental margin lies near to 60° latitude, for example around the southern tip of Greenland and the margin of Antarctica in the southern hemisphere at present. Land is cooler than the adjacent ocean at this latitude so air tends to sink relative to the air over the ocean. This effect locally intensifies the zonal belt of low pressure predicted for that latitude by the zonal model and creates cold water oceanic gyres.

In association with these different oceanic temperatures on opposite sides of continents at the same latitudes there are also quite different climates as summarised by Bambach et al (1980). On the eastern side of continents the tropical humid zone widens as the prevailing easterly winds bring moisture from the ocean. The zone is narrow on the western side confined to a region of intense heating where surface air is rising and losing moisture. In contrast, the wet belts in temperate latitudes are much broader on the windward west-facing margins of the continents.

Arid zones occur closer to the equator in the belts of easterly winds on the rain-starved western sides of continents and extend poleward in mid-latitude regions of prevailing westerly winds on their similarly rain-starved eastern sides so the limit of arid belts rise in latitude across the continents from west to east as shown in Figure 4.2. These features are all manifested in modern climatic regimes in the northern hemisphere. For example the arid belt rises from low latitudes in the Sahara in eastern Africa to high latitudes in the Gobi Desert of western Asia. France and Austria are at the same latitude as the Gobi Desert in the belt of the prevailing westerlies.

4.5.2.2 Effects of Regional Geography.

a) Continentality.

One of the major consequences of the difference in thermal regimes between land and sea is the greater range of surface temperatures experienced in the interior of large continents, both diurnally and seasonally (Frakes, 1979). A continental climate is characterised by extremes of temperature, relatively small rainfall and low humidity. It is experienced in the interior of large continents and by their neighbouring islands if they are exposed to prevailing winds from the interior during the winter. The term 'continentality' refers to the extent to which such conditions exist on any landmass and quantitative estimates of continentality are derived from recorded temperature ranges, humidity values and latitude. In this instance,the term 'continent' has been applied to a reasonably large land mass, for example the European, African or Asian continents of today and not to smaller continental areas such as the microcontinents Madagascar, New Zealand and Indonesia whose climates may be dominated by maritime influences.

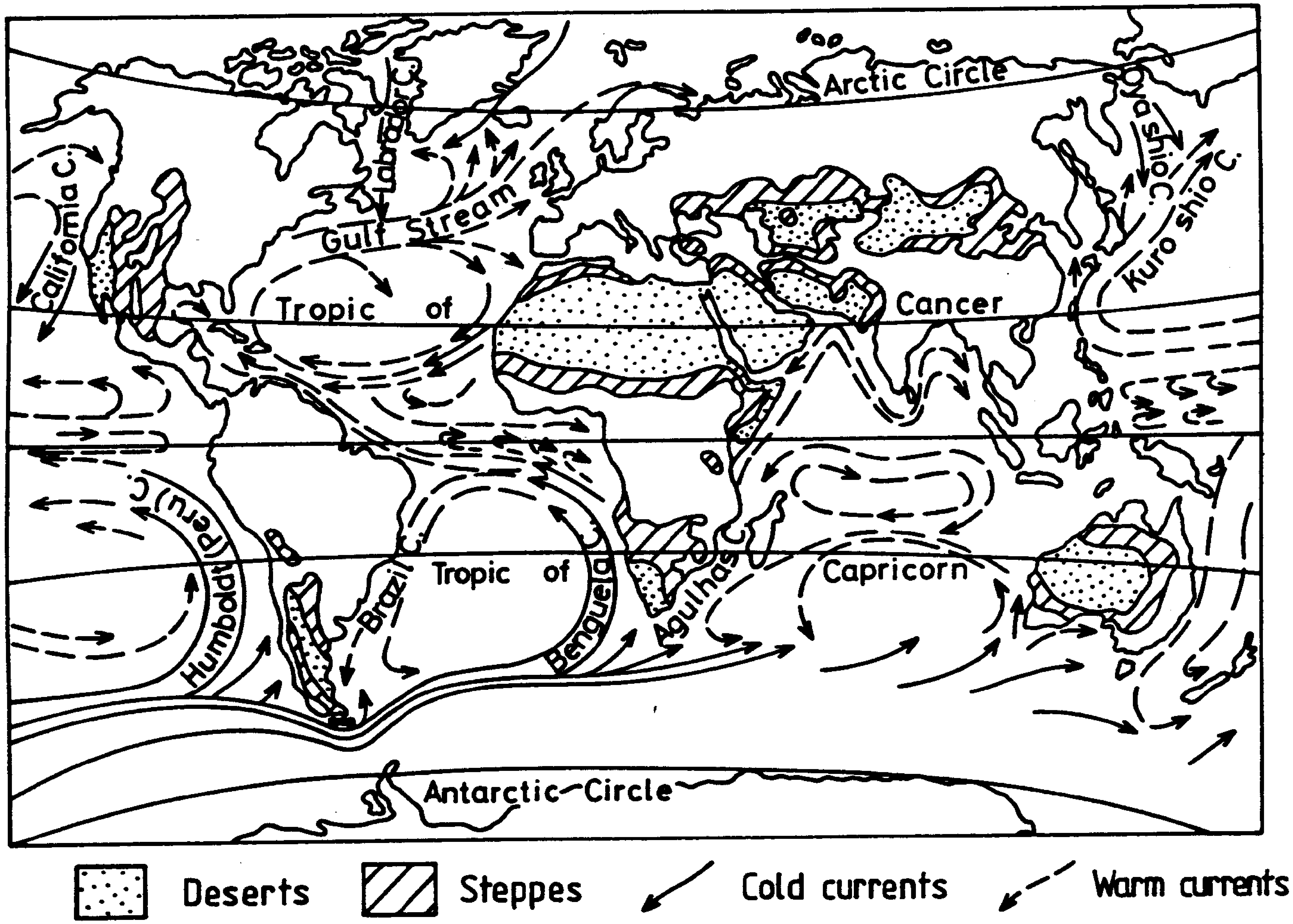


Figure 4.2. Present world distribution of arid zones and major ocean currents. (After Habicht, 1979, Figure 3).

b) Orographic Influences.

The presence of mountains can greatly influence the climate of an area for a number of reasons. For instance they control the precipitation in a region as the air is forced to rise over the mountains and so is able to hold less moisture. In the coastal mountains of middle latitudes the windward flank receives abundant rainfall whereas on the leeward side the passing air has so little moisture that deserts are frequently developed e.g. western North America, western South Argentina and parts of inner Asia (Rasool, 1984). The effect of mountain ranges upon precipitation patterns has been shown by comparison of Figure 4.3 (patterns on a continent of uniform low relief) with Figure 4.4 which shows the global precipitation patterns when mountain ranges have been added to the continental masses.

Desert conditions may also be caused by the absence of water vapour in the air, the absence of rain-inducing disturbances, or by a combination of both factors. The best examples of these circumstances are the down wind sides of cold oceanic currents, inner continental regions which are remote from sources of water vapour and regions such as central North America which are often humid, but rainless, because of the absence of cyclonic activity (Hare, 1977).

A pertinent example of the influence of mountains upon climate is that of distribution of climate-sensitive sediments off the coast of China described by Scotese and Summerhayes (1986). It also illustrates why caution must be exercised in the interpretation of such sediments. This particular effect of local geography upon climate was used by Hay et al (1982) with reference to late Triassic to Jurassic palaeoclimatology. Air rising over eastern coastal ranges cools, becomes saturated with moisture, rains result and conditions are then favourable for coal formation on the seaward/windward facing slopes. Once over the mountains the now dry air is heated as it descends into the rift valleys and is able to take up moisture from the valley floor, creating conditions favourable for the accumulation of evaporites - as in Death Valley,

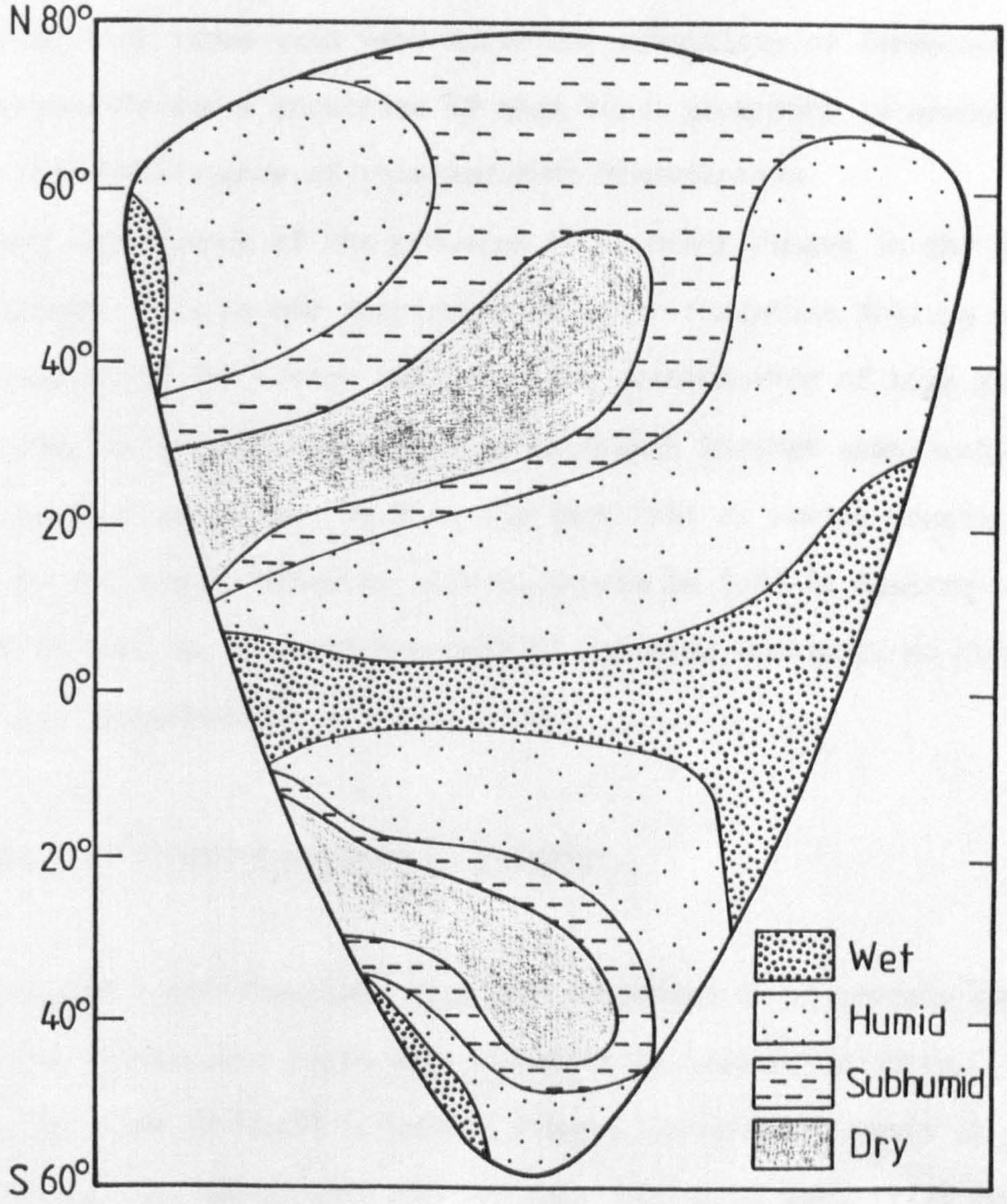


Figure 4.3. Schematic representation of the distribution of annual precipitation on a hypothetical continent of low and uniform relief. (After Robinson, 1973, Figure 1.b).

California at present. This hot dry air then rises over adjacent ranges, cools, becomes saturated and the cycle begins again. So as a consequence of the local geography two rock types with very different conditions of formation are found in close proximity and a knowledge of that local geography is needed to interpret the significance of this sediment distribution.

Another consequence of the presence of mountain ranges is the phenomenon known as albedo. This is the proportion of solar radiation falling on a non-luminous body which the latter reflects. The accumulation of high albedo (45 to 95%) snow and ice at high elevations on mountains further complicates the role of mountains in climatology. So does the fact that it varies considerably with the angle of the slope. Opposite effects should be felt in deserts but the bare ground may reflect up to about one-half of incoming sunlight, so this must also be taken into consideration (Frakes, 1979).

4.6 Climate and Climate-Sensitive Lithologies.

It has long been recognised that the accumulation of certain kinds of sediments and sedimentary rocks were favoured by specific climates. For example coals usually occur in humid climates; evaporites in arid; reefs in tropical, and tillites in glacial (Briden and Irving, 1964; Robinson, 1973; Drewry et al, 1974; Habicht, 1979; Ziegler et al, 1981 and 1984; Scotese and Van der Voo, 1982). Hence climate plays an important role in controlling the distribution of different rock types in sedimentary basins.

Previously in this chapter it has been assumed that the broadly zonal distribution of climates seen today has persisted throughout the Phanerozoic. It has also been assumed that the global, regional and local factors which affect the climate of today also influenced climates of the past. If this were not the case it would be very difficult to explain why tillites, carbonates,

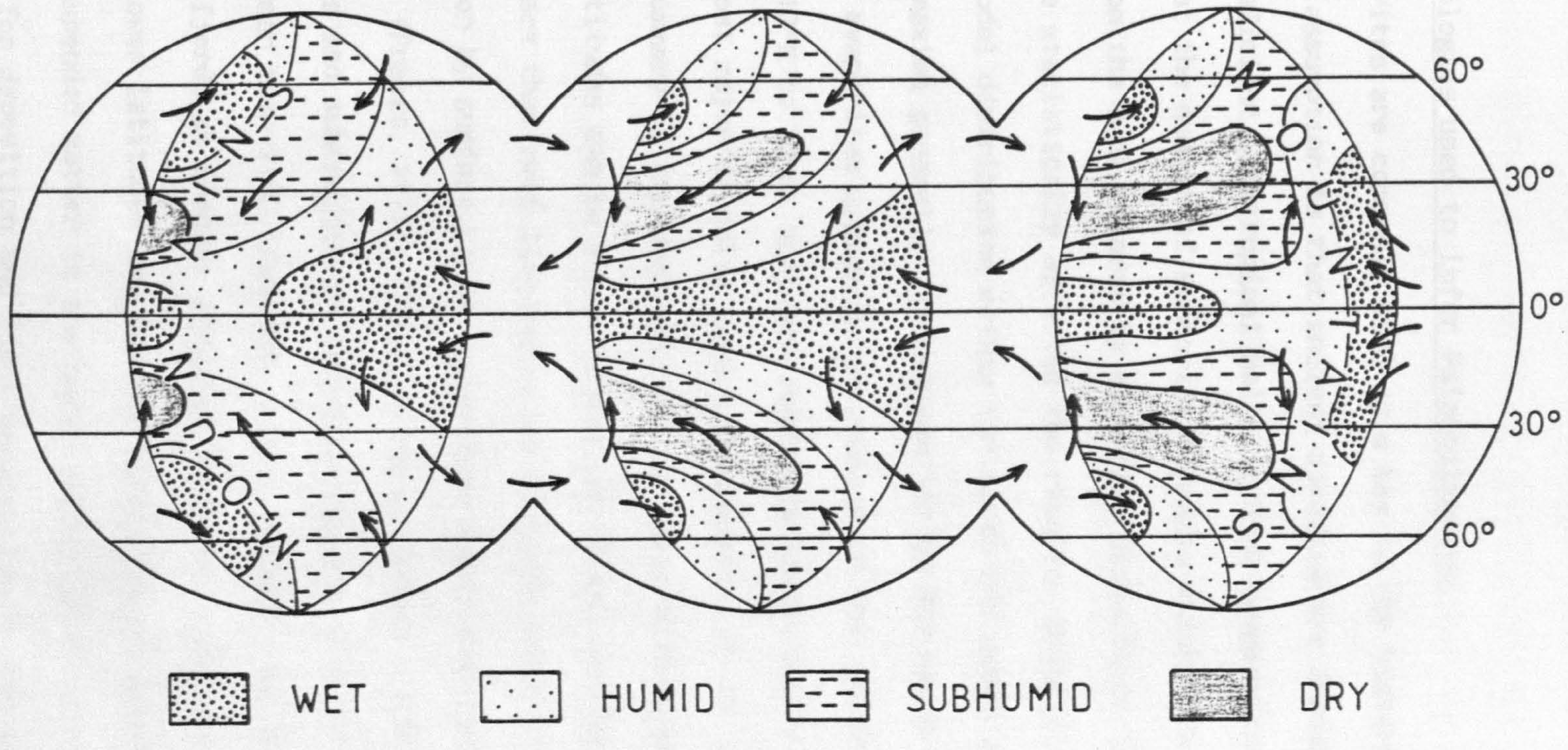


Figure 4.4. Schematic representation of the distribution of annual precipitation on a hypothetical continent as a function of latitude and geographical configuration. (After Bambach et al, 1980, Figure 3).

coals and evaporites were concentrated in the past in the similar narrow latitudinal belts to which they are confined today.

4.6.1 Lithologies used to infer Palaeoclimates.

Evaporites are commonly used as a key to the history of aridity on the globe. The assumption is that ancient deposits are analogous to modern rainfall-deficient, subtropical belts of high evaporation. It has also been assumed that the number of evaporite deposits which have been completely removed from the rock records by extensive dissolution is sufficiently small not to have statistically affected the results. Briefly, most modern evaporites have a bimodal distribution within 50° north and south of the equator and evaporite maxima generally lie between 10° to 30° north and south (Meier, 1981). Few evaporites appear to accumulate in the low pressure equatorial zone i.e. from 10°N to 10°S. Northern evaporite limits reach 50°N (Meier, 1981) although most major deposits have formed within 40° of the equator (Tarling, 1981). A summary of the work done on the value of evaporites in the inference of past latitudes can be found in Chapter Three, section 3.4.2.

The fact that reef limestones are strongly controlled in their global distribution by surface water temperature makes them excellent palaeoclimatic indicators (Frakes, 1979). The majority of modern bioherms are restricted to the tropics and subtropics at latitudes of less than 30° north and south and in waters warmer than 21°C (Habicht, 1979). However some reefs do occur in Norwegian fjords at present although these are single-species types whereas those in lower latitudes are usually multi-species types. Other factors, for example suspended matter in the water column, availability of suitable shallow platforms for deposition and light penetration to the sea floor (Ziegler et al, 1984) also limit the extent of reef distribution. However some of these may be more closely related to tectonics than climatic environment and may interfere with interpretations of palaeoclimate.

Coal is useful in recognising areas of high precipitation such as the equatorial rainfall belt and the modern distribution of peat bogs suggests generally high latitude and temperature sites of accumulation (Habicht, 1979). Low latitude occurrences are thought to result from the partial destruction of woody materials by `oxidation and bacterial activity' (Frakes, 1979). Coal may also accumulate in semi-arid areas of poor drainage given a high water table. These exceptions place limitations on the usefulness of peat and coal deposits as climatic indicators. Parrish et al (1982a) stated that temperature also modified coal distribution and that coals are under-represented at the equator. Habicht (1979) considered coals would only be relevant as temperature indicators when supplemented with botanical evidence.

Red-beds were among the first sedimentary deposits to be considered climatic indicators as their colour was assumed to reflect specific conditions of deposition i.e. a hot, arid, oxidizing environment. However the development of red pigment in most red-beds was complex and has proved difficult to decipher. The predominant clay minerals in red-beds are illite and chlorite and so provide no specific clue to the climate in the source area or at the place of deposition (see section 4.6.2). The palaeomagnetic evidence of the distribution of red-beds relative to their pole position corroborates the palaeogeographical data which suggest that most red-beds (evaporites and aeolian sandstones) accumulated less than 30° north and south of the palaeoequator (Briden and Irving, 1964; Kruseman, 1967; Schenk, 1969). One major problem in the evaluation of the palaeoclimate indicated by red-beds is that the reddening usually attributed to sandstones that have formed in hot desert environments may often have developed many millions of years after their deposits. The reddening can thus be a misleading indicator of the likely palaeolatitude at which such sandstones were deposited (Tarling, 1981).

In the late 1800s and early 1900s many geologists believed that ancient red sandstones were of desert origin because many of the world's tropical deserts are typified by red sand dunes (Crosby, 1891; Barrell, 1908). However later it was thought (e.g. Van Houten, 1948 and 1964; Krynine, 1950) that most red-beds formed in a climate of heavy, but seasonal, rainfall and tropical, lateritic weathering. Under this humid-tropical-red-bed hypothesis the presence of desert dunes at present had to be explained. They were interpreted as relics of past wetter climates or as having inherited their colour from older red sedimentary rocks. Walker (1967a, 1967b) claimed that red-beds could develop diagenetically in hot arid or semi-arid climates mainly by intrastratal weathering of heavy minerals during deep burial, aided by ageing. The hypothesis that the haematite pigment in red-beds forms in situ after deposition and is not derived from the erosion of red tropical (lateritic) soils is at variance with the views of Van Houten and Krynine. Walker (1974) later claimed that red-beds could form diagenetically in moist tropical climates by alteration processes similar to those producing red-beds diagenetically in deserts. But it was not possible to differentiate red-beds formed in moist climates from those formed in deserts. More recently Parron and Nahon (1980) considered Mesozoic-Cenozoic red-beds in France and some West African basins (Senegal and Ivory Coast) were the result of in situ lateritic weathering of glauconitic sediments.

It is well known that many major red-bed occurrences are associated with extensive evaporite deposits e.g. those of Permo-Triassic age in Europe and North Africa; those of Carboniferous to Triassic age in the western interior of the United States. The presence of these evaporites indicates that regionally arid climates prevailed during deposition (Walker, 1976). Thus the evaporites have important implications in the interpretation of the origin of the associated red-beds. This association does not prove that all red-beds form in desert environments (Walker, 1974) but suggests that arid climates are particularly favourable for their formation. The association of red-beds with sabkha evaporites is thought to be of significance in the search for sulphide deposits (Dean, 1979).

So red-beds per se are not reliable indicators of the climate either in the source area or in the depositional basin. The general consensus is that other characteristics of ancient red-beds e.g. fauna, flora and association with evaporite minerals or with aeolian sandstones should provide the most reliable evidence of the climate at the time of deposition.

4.6.2 The Effect of Climate on Sedimentation.

The lithologies usually found in particular climatic zones as seen today are mentioned below. These same climatic patterns are thought to be reflected in the distribution of lithofacies through the Phanerozoic as the continents migrated between climatic zones.

a) The Tropical Zone spans the area between 23.5° N and S and includes the equatorial wet zone and about half of the subtropical dry zone. The low latitude hot wet zone is represented by thick clastics, coals and carbonates. It is best developed along the east coasts where prevailing winds bring moisture and heated surface winds towards the continent. Desert zones occur on the western sides of continents centred at 20° N and S; dry belts are represented in the geological records by evaporites.

b) The Temperate Zone is commonly defined as the area between the tropics and the Arctic Circle and is generally wet except where mountains or broad land areas dissipate the moisture. Thick clastics and coals occur in temperate rainy belts, especially on the windward, western sides of continents above 40° latitude.

c) The Polar zone is the area above the Arctic Circle represented in part as tillites in the geological record.

Good consistency through the Mesozoic and Cenozoic in terms of latitudinal occurrence of coals and evaporites has been determined (Gordon, 1975 and

Parrish et al, 1982b). Although the latitude positions of continents during the Palaeozoic are less well constrained, similar lithofacies patterns have been demonstrated (Ziegler et al, 1981).

The distribution of the clay mineral content in sediments and soils shows a distinct global zonation which is related to climate. One particular type of clay mineral, montmorillonite, may be significant in the study of sediment-hosted mineral deposits because of its capacity for adsorption of moisture and its ability for base exchange. The kandite group (e.g. kaolinite, chamosite, greenalite) forms in areas where rainfall exceeds evaporation and leaching is intense. The latter is an important factor as excessive silica ions must be removed to maintain a high Al:Si ratio. This group is dominant in tropical and equatorial rivers and seas. The smectite group (e.g. montmorillonite, nontronite) occur in regions where evaporation exceeds precipitation and leaching is negligible. It is also essential that alkaline conditions should prevail to maintain a low Al:Si ratio. Clay minerals in sediments of middle latitudes are dominated by the illite group (e.g. illite, glauconite) where a non-acid, potassium-rich environment occurs. Rainfall and the consequent leaching should only be moderate and intermittent. The chlorite group of clay minerals are found mainly in high latitude sediments being relatively unstable and chemical weathering is more subdued in these regions (Greensmith, 1978 and Smith, 1981). In areas of subaerial weathering and soil formation the prevailing physical and chemical environments are important in determining the type of clay mineral which occurs. However changes in the climate of the source area and hence these environments may be ultimately reflected in the clastic clay mineral content of the derived sediments.

One of the major problems of using palaeoclimatic indicators to infer palaeoclimates is the grouping of data over too long a time interval, as pointed out by Robinson (1973). This problem is illustrated in Figure 4.5 showing the differing climatic zones of a continent of long latitudinal extent

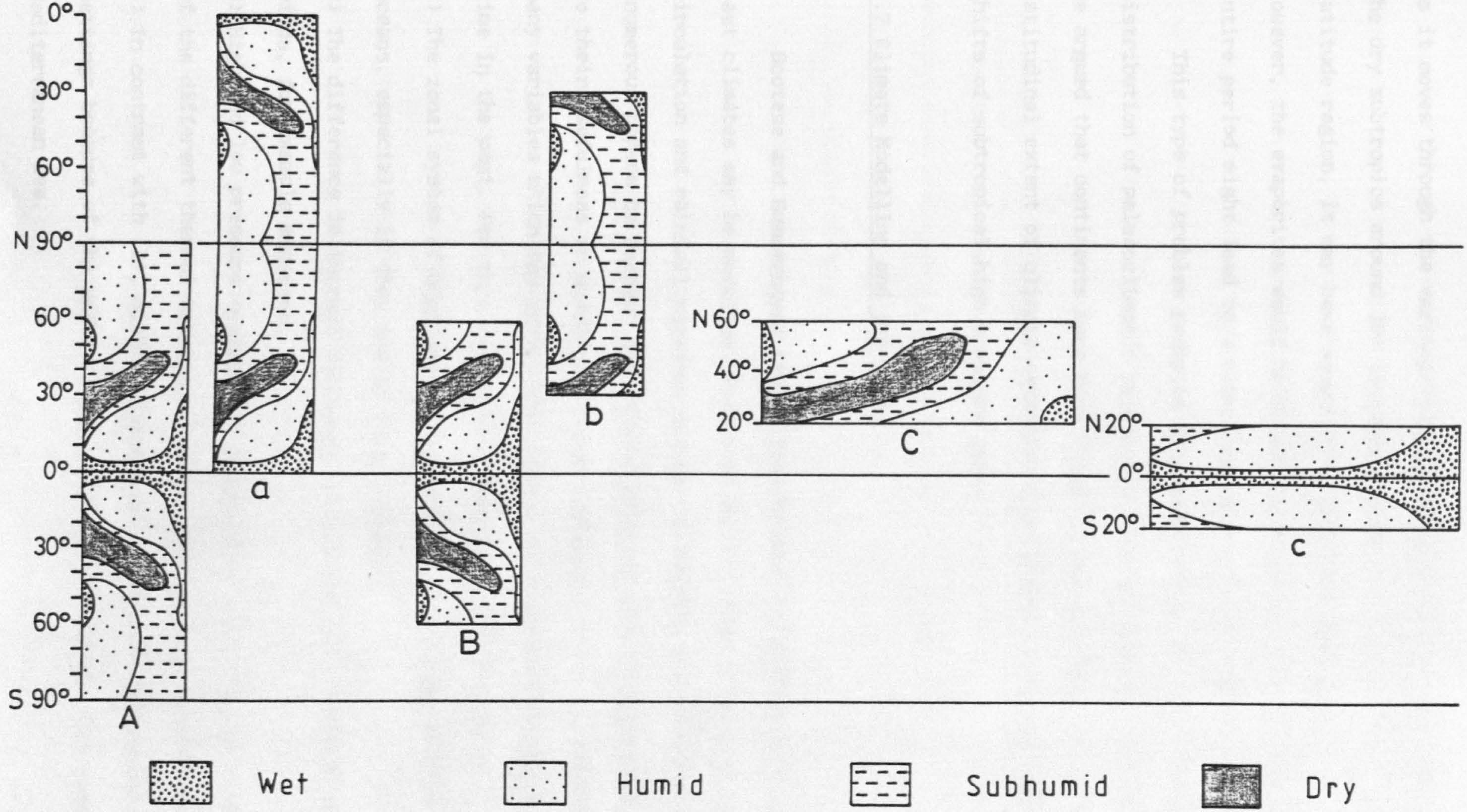


Figure 4.5. Idealized models of hypothetical, rectangular continents showing the distribution of annual precipitation regions as a function of latitude. (After Robinson, 1973, Figure 4).

as it moves through the various climatic regimes. Evaporites tend to form in the dry subtropics around 30° latitude and as a continent moves through that latitude region, it may have evaporites deposited over much of its area. However, the evaporites would be diachronous and the use of evaporites from an entire period might lead to a modelling of a dry zone that is too wide.

This type of problem prompted Meyerhoff (1970) to conclude that the distribution of palaeoclimatic indicators does not support continental drift. He argued that continents have been fixed during the Phanerozoic and changes in latitudinal extent of climate-sensitive lithologies simply reflected major shifts of subtropical high pressure zones.

4.7 Climate Modelling and its Uses.

Scotese and Summerhayes (1986) listed seven main parameters upon which past climates may be modelled. They were derived from models of atmospheric circulation and rainfall patterns devised by Parrish and others from Chicago (numerous papers by Parrish and Zeigler given in the bibliography). In addition to their usefulness in modelling, these parameters serve as reminders of the many variables which may affect the climate of a particular region at a certain time in the past. For this reason they have been included here.

1) The zonal system of high and low pressure belts is best developed over the oceans, especially if they are of large extent.

2) The difference in thermal behaviour of land and sea generates seasonal winds, for example monsoons.

3) High and low pressure systems are intensified over continents as the effect of the different thermal behaviour between land and sea becomes exaggerated.

4) In contrast with (3), milder climates occur in water-dominated continental settings because of the ameliorating effect of the ocean, for example the Mediterranean Sea.

5) High mountains may form effective barriers between pressure systems.

6) The Hadley cells are seasonally displaced due to the Earth's obliquity at the junction of the Trade Winds i.e. the Intertropical Convergence.

7) High pressure systems are intensified off the western coast of continents.

There are three types of climate model currently in vogue. Outlines of these approaches have been given to highlight the problems involved in the determination of palaeoclimates.

4.7.1 Direct Measurement Models.

Quantitative data on important climatic parameters can be entered into existing numerical climate models derived from an understanding of present atmospheric and oceanic circulation systems, for example the dynamic climate simulation models of Barron et al (1981). These models take into consideration such variables as the Earth's rotation, the sun's insolation and the amount of cloud cover in addition to other important physical factors, for example continent positions, continental area and topography. One of the more useful aspects of this model type is that it can simulate temperature as well as pressure. However, as Barron (1985) acknowledged, these models are still very much in the development stage and have a limited application.

For the Cretaceous and younger, interpretations from isotopic data suggest sea surface palaeotemperatures and palaeogeographical and biogeographical data are more widespread and reliable than are those for earlier periods. So research has been restricted to the latter part of the Phanerozoic, for example Barron and Washington (1982 and 1984) and Barron (1985). Unfortunately data for the Palaeozoic are so sparse and unreliable that the direct approach of dynamic modelling is not feasible at present.

4.7.2 Inference Models.

These models are based on the notion that the distribution of sediments and organisms may be used to infer palaeoclimatic patterns. Heckel and Witzke (1979) presented a Devonian palaeogeographical reconstruction based solely on the distribution of lithic palaeoclimatic indicators such as marine carbonate sediments and marine-derived evaporites. Supplementary data from phosphorites, bauxites and coal were used in conjuction with tectonic considerations. It was assumed that evaporites and reefs were climatically controlled deposits and that there was a constancy of general physical principles of atmospheric circulation patterns. Care must be taken when using these models as they are based largely on circular reasoning. There can be no real test of these models as data and model are combined.

4.7.3 Theoretical Models.

This approach does separate the model from the data, the former being based on palaeogeography and the more general principles of atmospheric circulation. It has been adopted by Parrish (1982) and Ziegler (1981a, 1981b and 1984) and by Scotese and Summerhayes (1986) for their parametric model of palaeoclimates. It is possible to independently test the model using palaeoclimatic indicators. A comparison is made between the observed distribution of climatically controlled sediments or organisms and that predicted from the model. An advantage is that the approach may be used when palaeoclimatic data are sparse but no predictions about the Earth's thermal regime are possible.

Generally there is good correspondence between the predictions of Scotese and Summerhayes and those of Barron et al despite the two approaches being radically different. This correspondence gives support to the use of such climate modelling to determine the climatic conditions of a region in the past and the atmospheric and oceanic circulation systems which may have created that climate.

4.8 Conclusion.

The intention of this chapter was to show that the Uniformitarian Principle can be applied to palaeoclimatology. However, it must be used with the proviso that numerous factors hinder uninhibited application of this principle and these need to be taken into consideration.

In accepting that the Earth's climatic belts have had a similar latitudinal extent throughout the Phanerozoic many assumptions have been made including the uniformity of atmospheric circulation, constancy of the Earth's rate of rotation and its obliquity. With the aid of such climatic assumptions it has been possible to predict regions of upwelling (Parrish and Curtis, 1982); the distribution of evaporites and coals (Parrish et al, 1982b); the distribution of petroleum source rocks (Barron, 1985); and they have been useful in the parametric approach to climate modelling adopted by Scotese and Summerhayes (1986). These successes lend credence to the assumption that uniformitarianism with regard to palaeoclimatology is, in part, acceptable. However the use of geological data is limited as there are frequently insufficient data to interpret climatic conditions fully. This shortage may result from poor preservation, alteration or dissolution of materials or it may arise from the fact that many climatic parameters (e.g. atmospheric pressure) leave no discernible mark on the rocks.

The relationship between palaeoclimatology, latitude and mineral deposits must now be examined. Workers such as Briden and Irving (1964) found a palaeolatitudinal control upon the distribution of certain sediments. It has been shown that latitude has an effect upon climate influencing atmospheric circulation (the position of an area in the cellular circulation system), oceanic circulation patterns, temperature and precipitation. It follows that the formation of these sediments must be palaeoclimatically dependent (Frakes, 1979; Habicht, 1979; Parrish et al, 1982; Ziegler et al, 1984). It has been noted (Chapter 3, section 3.2.7) that there is a close relationship between some mineral deposit types (e.g. SSCU, SDEX, SSUV) and some climate-sensitive lithologies (e.g. red-beds and evaporites). Therefore the distribution of such mineral deposit types must also be palaeoclimatically influenced. If this is so, then mineral deposit distribution (and possibly genesis) must also be palaeolatitudinally controlled and thus allows a more meaningful genetic classification to be considered. Hence the purpose of this research.

With the acceptance that palaeoclimate is directly related to palaeolatitude, it must also be accepted that this is a very complex relationship and many factors may affect this correlation; for example local geographical effects upon the distribution of climate-sensitive lithologies and the asymmetry of climatic conditions on opposite sides of a continent. However if a latitudinal control upon the formation of some mineral deposit types can be shown to exist, then it may be possible to explain anomalous occurrences by local variations in the predicted climatic conditions that prevailed during the formation of that deposit.

CHAPTER FIVE

METHODS

5.1.Literature Search.

Data collection was achieved through an extensive literature search from many sources including numerous journals, text books and BP Minerals Internal Reports. Information from the latter was restricted due to a confidentiality clause associated with the reports, but the data extracted, for example location of established ore deposits, were generally known so no confidentiality has been violated.

All the references used to compile the data-sets can be found in the bibliography flagged with a 'D'; those flagged with a 'T' were used for the text. Obviously, some references were used for both. This notation was used in order to facilitate the use of the bibliography for future workers.

Non-economic deposits have been included within this study as the grade and tonnage of deposits were not always reported within the references. It is, in any case, difficult to assess the potential economic value of individual deposits as parameters which may make a deposit economic in one instance may make it uneconomic in another. An economic constraint upon data selection would thus have introduced economic and political biases which would have influenced results. To set strict limitations on the size of deposits to be included within this study would have been extremely difficult and could have been potentially misleading. It is not the value of the deposit that is of interest in this work, it is its presence. Economic deposits are anomalous by their nature and to consider only these deposits may lead to the loss of important information regarding the genesis of some mineral deposit types. Nonetheless most data were available for deposits on a large and economic scale and so there was unintentional bias towards such deposits.

Figure 5.1 shows the present distribution of all the mineral deposits used in this thesis on a world map. It is worth noting the patchy, uneven distribution of the examples, illustrating the way in which exploration and research have been centred on some areas, such as North America and Europe, rather than others (e.g. areas of South America). It may also be a reflection on the amount and quality of information available, for instance, data from the USSR can be very sparse.

5.2. Data Assembled for each Deposit.

The following four sets of data were compiled for each mineral deposit.

(i) Ore Deposit Type.

Each example was classified according to the criteria defined in Chapter Three. The classification of each deposit was purely subjective and other authors have their own criteria for selection of data. However some grouping of deposits had to be made in order to facilitate handling of the data. The grouping chosen emphasizes the descriptive features of the deposit types and avoids, where possible, genetic presumptions.

(ii) Present Day Co-Ordinates.

The co-ordinates for each deposit were obtained from the Times Atlas or from the maps displayed in the original references.

(iii) Ages of Host Rock and Mineralisation.

The age of the host rock for deposits considered to be syngenetic is vital as this corresponds to the age of mineralisation. However the ages of the host rocks and the mineralisation should be very different for epigenetic deposits.

KEY: Figure 5.1.

SDEX	Sedimentary-exhalative (Pb, Zn, Ba)	1
LSBM	Limestone Base-metal (Pb, Zn, Ba, F)	2
SSCU	Sandstone Copper	3
SSPB	Sandstone Lead	4
SSUV	Sandstone Uranium-Vanadium	5
SHUR	Shale Uranium	6
SHBM	Shale Base-Metal	7
PLAU	Placer Gold	8
PLDI	Placer Diamonds	9
PLSN	Placer Tin	10
PLOX	Placer Oxide	11
PLOT	Placer "Others"	12
PAPL	Palaeoplacer (Au, U)	13
OOFE	Oolitic Ironstone	14
FEFM	Iron Formation	15
MNFM	Manganese Formation	16
MNNO	Manganese Nodules	17
PHOS	Phosphate	18
MEVA	Marine Evaporites	19
CEVA	Continental Evaporites	20
SULF	Sulphur	21
LATO	Laterite (Ni, Al)	22
GOSS	Gossan	23
CALU	Calcrete Uranium	24
SUPE	Supergene Enrichment	25

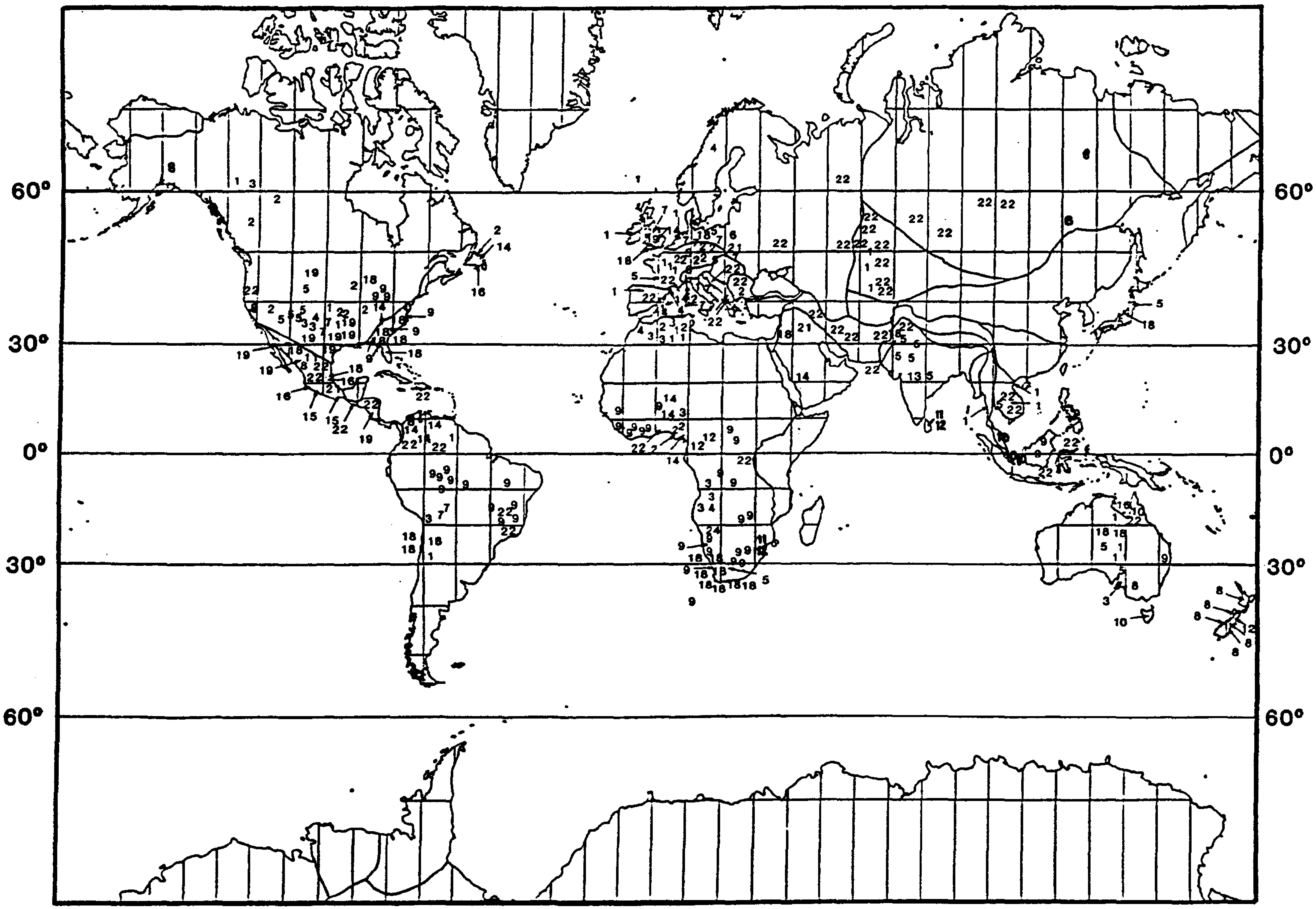


Figure 5.1. Present distribution of all mineral deposits collected for this project. Numbers represent mineral deposit types - key as in text (page 98).

In many of the cases only rarely were the age of both the host rock or the mineralisation cited so a compilation of data from several sources was often necessary.

(iv) Lithological Associations with possible Climatic Implications.

Any study of mineral deposits and their surrounding rocks rapidly shows an association between certain types of sediments. For example red beds and evaporites commonly occur in the host rocks of sedimentary-exhalative and sandstone-hosted copper deposits. This relationship is repeatedly found: Gustafson and Williams (1981) noted that of twenty eight deposits cited in their paper only five (Meggen, Rammelsberg, Laisvall, Howard's Pass and Broken Hill) lacked obvious evidence of evaporites or evidence of aridity in the associated sedimentary environments. Although some deposits are apparently devoid of an association with evaporites care must be taken to note any evidence of the presence of metamorphic evaporites. Such metamorphic evaporites are to be found at Broken Hill and are in close proximity to both the Howards's Pass and Laisvall deposits.

The main climate-sensitive lithologies which were noted specifically for their possible inference as to the climate at the time of their deposition were red beds, evaporites, coal and tillites. Other lithologies, such as limestones, may also be used as palaeoclimatic indicators. They do have a less restrictive climatic inference (see Chapter Four, section 4.6.1) but a valid one, nonetheless.

5.3 Preparing Data for Computation.

5.3.1 Co-Ordinates.

The present day co-ordinates of each example have been converted into their decimal equivalents to the nearest 0.1° for computer plotting and analysis. All co-ordinates (even those from present day examples) were subjected to this conversion to maintain a standard format.In accordance with convention, North and East Co-Ordinates are positive; South and West Co-Ordinates are negative. Therefore the co-ordinates 39° 46'N 116° 15'W become 39.8 -116.2.

A number of mineral deposit localities have been eliminated from the original list. Individual data points were not evenly distributed, some areas having received more attention from geologists than others. It was therefore necessary that the data from closely spaced localities should be filtered in order to avoid over-emphasising the well-studied regions. It was difficult to restrict sample size to kilometres (rather than a percentage of a degree latitude and longitude) while collecting the data at source, as the scale of maps displayed varied greatly between references and frequently no scale was shown on the diagram. Therefore all data within a twelve minute (12') latitude-longitude square were considered as one data point. The twelve minute filtering square was chosen so that any deposit within 0.2 of a degree latitude-longitude of another is eliminated, once the co-ordinates have been converted to their computer format. The area covered by such a filter square is approximately 400 sq.km. at the equator. The area encompassed by such a square will vary according to its latitude, but most deposits are in mid latitudes so it was considered to be a constant way to filter data once collection was completed.

The sample size of 0.2° is a fairly fine filter when considering mineral deposits, some of which may have extensive occurrences e.g. phosphates, sedimentary-exhalative or shale-hosted base-metal deposits (such as the Kupferschiefer). However other deposits, for example some sandstone-coppers, are found only within narrow environmental constraints such as palaeochannels, so the filter needed to be fine enough to show the incidence of these mineral occurrences.

Ore districts, such as Grants District, New Mexico (SSUV deposits) and Mississippi Valley District (LSBM deposits) usually contain many ore deposits, or mines, within a relatively small area. In such cases, only a few deposits were chosen to represent the extent of the area covered. Generally, an attempt was made to select deposits rather than individual mines. The latter are parochial economic conditions while the former are the results of geological processes which are considered here.

Any sampling bias present at this stage may affect the intensity of the distribution, but not its range.

5.3.2 Sorting of Data for Plate Movement Program.

Once the data had been converted to a format suitable for computer analysis, and deposits of the same age and palaeolatitude had been eliminated, the data were sorted into the age range categories shown in Table 5.1, based on age of mineralisation. Only one example of deposits with the same age and palaeolatitude was retained to prevent a bias in the results. The concentration will be affected but such filtering should have eliminated too close sampling of the deposits so giving a more unbiased representation of the distributions.

Table 5.1: Subdivisions of Phanerozoic Time used for Deposit Age Ranges.

AGE RANGE FOR THESIS (m.y.)	ROTATION POLE (MY)	GEOLOGICAL PERIOD COVERED (APPROX.)	ACTUAL AGE RANGE * (m.y.)
0 - 25	0	Present Day	0 - 25
26 - 75	50	Eocene	26 - 65
76 - 110	100	Cretaceous	66 - 144
111 - 160	130	Upper Jurassic	145 - 181
161 - 225	200	Lower Jurassic	182 - 213
226 - 275	250	Permo-Triassic	214 - 286
276 - 325	300	Upper Carboniferous	287 - 320
326 - 375	350	Lower Carboniferous	321 - 360
376 - 425	400	Lower Devonian	361 - 408
426 - 505	450	Ordovician and Silurian	409 - 505

*According to "Subdivisions of Phanerozoic Time", Cambridge University Press, 1982, 2nd. Edition. Based on information published in "A Geologic Time Scale" by W.B.Harland et al (1982), Cambridge University Press.

Nine datafiles were produced, each representing a specific age range from the Eocene (50 m.y. mid-point) to the Ordovician and Silurian (450 m.y. mid-point). The age categories are loosely based on 'Subdivisions of Phanerozoic Time'(British Petroleum Company p.l.c., 1982). They were also influenced by the reconstructions available on the Newcastle Plate Movement Program. The actual ages differ from the thesis age ranges in Table 5.1. The former are those of the "Subdivisions of Phanerozoic Time" whereas the ages represented by the geological periods have been adapted for the thesis ranging from 25 m.y. younger to 25 m.y. older than the plate rotations which are available.

Each of the nine datafiles were then divided into groups appropriate to the plates on which they originated, as in Table 5.2. For example, a mineral deposit located in Brazil with an age of mineralisation of 36 m.y. would be placed into the Eocene datafile (according to Table 5.1) divided as part of the S.American plate and rotated according to the values for the 'South America' plate, Eocene (50 m.y.) shown in Table 5.2.

5.4. Global Palaeogeographical Reconstructions.

5.4.1. Definition of Plate Boundaries.

The continental plates shown in the palaeoreconstructions presented in Figures 6.15 a-j are represented by their present day coastlines and major geological boundaries. In fact the shelf margins (loosely defined by the 2000m bathymetric contour) were used to "fit" the continents together. Some of the present day land masses, for example Australia, can be treated as a single block or plate, whereas others must be subdivided along sutures determined from geological evidence such as ophiolites, accretionary flysch wedges and other sediments indicative of a former ocean.

Table 5.2: Total Rotations for each continent.

PLATE	TOTAL ROTATION USED		
	LAT.	LONG.	POLE
EOCENE (50 MY)			
Europe	-59.4	101.8	18.0
North America	0.0	94.5	15.5
Central America	0.0	94.5	15.5
Africa	-42.0	117.1	21.6
India	-28.0	-175.5	36.5
Australia	5.7	-168.8	37.8
South America	19.2	88.3	10.2
Arabia	-43.7	133.7	25.2
New Zealand	-23.7	-166.9	37.1
Sundaland	-14.4	-78.8	40.6
Siberia	-59.4	101.8	18.0
CRETACEOUS (100 MY)			
Europe	-55.5	103.7	31.7
North America	0.0	95.1	21.6
Central America	0.0	95.1	21.6
Africa	-49.1	132.3	47.0
India	-24.0	173.2	90.3
Australia	-22.1	-168.6	42.3
South America	-20.3	89.0	10.8
New Zealand	-54.0	-172.8	60.6
Siberia	-55.5	103.7	31.7
UPPER JURASSIC (130 MY)			
Europe	51.2	132.3	10.2
North America	76.0	119.0	40.8
Central America	76.0	119.0	40.8
Africa	0.0	169.0	34.0
India	-2.7	159.0	95.7
South America	81.8	-23.5	40.7
New Zealand	-31.8	-174.9	32.8
Sundaland	51.2	132.3	10.2
Siberia	51.2	132.3	10.2

Table 5.2: Total Rotations for each Continent.

PLATE	TOTAL ROTATION USED		
	LAT.	LONG.	POLE
LOWER JURASSIC (200 MY)			
Europe	66.9	42.5	40.5
North America	76.8	44.2	65.5
Central America	62.8	109.7	49.5
Africa	0.0	169.0	23.0
India	-6.1	-156.3	86.6
South America	67.1	28.6	43.5
Arabia	-6.5	175.4	27.6
Sundaland	8.1	-106.9	38.7
PERMO-TRIASSIC (250 MY)			
Europe	0.0	76.3	45.6
Turkey	0.0	76.3	45.6
Laurentia	28.5	67.5	54.5
Gondwanaland	-21.8	130.0	63.4
Sundaland	-21.8	130.0	63.4
China/Japan	-21.8	130.0	63.4
UPPER CARBONIFEROUS (300 MY)			
Europe	0.0	72.7	50.7
Southern Europe	0.0	72.7	50.7
Central Europe	0.0	72.7	50.7
Laurentia	26.1	63.6	58.8
Gondwanaland	-12.7	128.4	74.5
Sundaland	-12.7	128.4	74.5
China/Japan	-12.7	128.4	74.5
Siberia	29.6	94.0	39.8

Table 5.2: Total Rotations for each Continent.

PLATE	TOTAL ROTATION USED		
	LAT.	LONG.	POLE
LOWER CARBONIFEROUS (350 MY)			
Europe	-13.6	58.0	148.3
Southern Europe	-16.7	79.0	74.8
Central Europe	-12.4	76.5	75.4
Turkey	-10.8	120.8	72.2
Laurentia	0.0	60.0	74.5
Gondwanaland	-15.8	111.1	90.2
China/Japan	-15.8	111.1	90.2
Siberia	0.0	62.0	62.0
LOWER DEVONIAN (400 MY)			
Europe	-9.7	66.0	86.5
Southern Europe	-9.4	66.3	87.5
Central Europe	-8.7	67.5	90.9
Turkey	-26.8	104.9	69.7
Laurentia	6.1	55.0	87.3
Gondwanaland	-7.7	82.5	89.0
Sundaland	-7.7	82.5	89.0
China/Japan	-7.7	82.5	89.0
Siberia	3.8	64.5	66.2
ORDOVICIAN & SILURIAN (450 MY)			
Europe	9.7	55.0	92.7
Southern Europe	13.5	75.2	85.9
Central Europe	22.0	58.5	99.4
Turkey	7.2	80.0	85.4
Laurentia	9.0	50.0	96.6
Gondwanaland	-17.7	75.0	124.6
Sundaland	-6.8	58.6	130.0
China/Japan	-17.7	75.0	124.6
Siberia	11.4	43.7	96.8

A complete list of the major continental units used for all the Phanerozoic reconstructions is as follows:

Laurentia	Siberia	Australia	Sundaland
North America	Europe	Arabia	New Zealand
South America	Central Europe	India	China/Japan
Africa	Southern Europe	Nafnam	Alaska
Madagascar	Nova Scotia	Greenland	Gondwanaland
Turkey	Central Asia	Newfoundland	Central America
E.Antarctica	W.Antarctica	W.Antarctic Peninsula	

Although a detailed description of each plate is not necessary,a number of the plates listed above require clarification as to their constituent parts, some of which may vary with time.

A number of the plate outlines are particularly poorly defined (e.g. Sundaland). However such uncertainties do not have a drastic effect upon the results of this research as only a small proportion of deposit examples occur in this region. In contrast the definition of the southern boundary of Laurentia is particularly relevant as a large number of examples occur on this plate. A detailed description of this boundary and the problems associated with its definition are given by King (1975a and b; 1977).

The plates for the Silurian and Devonian reconstructions can be found in Tarling (1985c); those for the Carboniferous in Tarling (1985a) and a discussion on the fragmentation of Gondwanaland in Tarling (1980). Other problems associated with the palaeogeographic reconstructions (e.g. the tectonic interpretation of the Mediterranean region during the Mesozoic and Cenozoic) may be found in Tarling (1983) together with possible solutions.

Laurentia (LAUREN)

Comprising North America, Greenland, Scotland, Mexico, California, Alaska and Florida.

Central America (CENTAM)

The microplates of Mexico, Yucatan, Honduras and Baja California are treated separately from North America from the mid Mesozoic onwards to avoid overlap with South America. This whole area is extremely complex geologically with some evidence of Cretaceous fault motion. However the actual coupling of plates is unclear.

N.America and N.Africa (NAFNAM)

Comprising the eastern coastline of North America, roughly the Appalachians and Caledonides, and part of West Africa. This region is Acadian in age, having been formed by the collision of N.America and Africa. Consequently it is only found in the Ordovician and Silurian reconstruction.

Northern Europe (EUR)

Comprising Northern Brittany, Sweden, Poland, Ukraine, Germany (excluding the Bohemian Massif), Norway, Finland, Denmark, Belgium and Holland. Also included in this plate are the British Isles excluding Scotland and Ireland north of the Caledonian suture.

Central Europe (CEUR)

Comprising the Armorican Peninsula, Massif Central, Vosges-Black Forest and the area leading south towards the Bohemian Massif. The Czechoslovakian data referring to the Bohemian Massif are of poor quality, so there is a tendency to leave this area as a separate block. The problem is its palaeoposition relative to other land masses.

Southern Europe (SEUR)

Comprising the Iberian Peninsula (Spain and Portugal), Sardinia, Corsica, Balkan Peninsula, Hungary and the Menderes block, the area to the south of Bulgaria, Kabylie block in Algeria, the area from the Bohemian Massif northwards to the region encompassed within Central Europe. It is questionable whether or not Italy and Sardinia should be included here or as part of Africa. The geology of Italy is extremely complex so the determination of whether blocks are allochthonous or autochthonous is very difficult. Hence the position of Italy during the Mesozoic and Cenozoic is a problem. The positions of Corsica and Sardinia are also difficult to determine as magnetic overprinting has obscured much of the evidence (Tarling, 1983). Italy has been rotated as if it were part of Africa in Cretaceous times.

Gondwanaland (GOND)

Comprising Africa, Madagascar, Arabia, South America, India, Australia, Antarctica, New Zealand and Iran.

Siberia (SIB)

This unit is quite well defined as being that part of Asia bordered to the west by the northern part of the Ural Mountains; to the east by the Verkhoyansk Mountains; the northern boundary being the Arctic continent and the southern one being the Kazakhstan block. The problem with this unit is the southern boundary, more specifically, its relationship with the Kazakhstan block. Major fault lines are found within this area, many of which were active in the Mesozoic and Cenozoic and which may also have been active in Palaeozoic times. This leads to complications in reconstructing the history of the area and the problem is confounded by the presence of other lineaments which may well represent previous oceanic areas.

Central Asia (CENTAS)

Comprising the area to the east and south of Kazakhstan, but excluding India.

"Sundaland" (SUNDAL)

Comprising Malaysia, Thailand, Vietnam, Cambodia, Kalimantan, Borneo and the western side of Papua New Guinea - the eastern part being considerably younger. This probably comprised separate units at different times.

China/Japan (CHI/JAP)

The southern and northern parts of China are treated as a single unit, although they were only joined together in late Palaeozoic or early Mesozoic times. There are numerous granite intrusions present in certain areas, thought to be indicative of previous sutures, implying the two regions of China were separate entities.

5.4.2. Phanerozoic Mean Rotation Poles.

A rotation pole has been defined in Chapter Two, together with an outline of the criteria for the selection of these poles. The rotation poles as shown in Table 5.3 constitute a fundamental dataset critical to the production of reconstructed maps, and thence palaeolatitudes.

The mean pole positions used for the reconstructions are those cited in Tarling (1983) from which Table 5.3 was taken. The poles upon which the rotations were based are given in the left hand column, while alternative poles are given in the right hand column.

Table 5.3: Phanerozoic Mean Pole Positions.

AGE	N	LATITUDE	LONGITUDE	*α95	N	LATITUDE	LONGITUDE	α95
NORTH AMERICA								
Cretaceous	5	68.4	185.1	5.8				
Jurassic	13	73.4	98.6	9.6				
Triassic	40	60.8	96.1	3.1				
Permian	43	44.5	116.3	2.7				
Carboniferous	33	37.6	125.1	3.4				
Devonian	4	-18.8	133.9	36.4	27	39.4	128.0	13.9
Silurian	E	0.0	120.0	-	13	13.6	127.8	16.0
Ordovician	4	-13.9	114.4	16.8	15	20.9	136.0	22.3
EUROPE								
Cretaceous	E	72.0	173.0	-	-	72.0	173.3	5.0
Jurassic	E	67.0	155.0	-				
Triassic	69	54.7	159.8	4.2				
Permian								
Upper	73	44.4	166.3	1.3				
Middle	21	39.1	163.3	3.9				
Lower	62	40.8	165.6	2.0				
Carboniferous								
Upper	37	39.3	162.7	2.8				
Middle	24	28.8	174.4	4.3				
Lower	17	12.9	137.8	16.1				
Devonian	E	10.0	145.0	-				
Silurian	E	5.0	150.0	-	-	-0.5	135.0	7.9
Ordovician	E	0.0	150.0	-	-	15.6	148.7	8.1
								(URA)
SIBERIA (ANGARA)								
Cretaceous	33	76.1	175.7	4.8				
Jurassic	16	76.7	143.2	13.3				
Triassic	67	49.3	147.4	2.6				
Permian	36	45.4	146.5	6.4				
Carboniferous	E	30.0	150.0	-				
Devonian	E	25.0	155.0	-				
Silurian	8	-4.3	121.2	19.0				
Ordovician								
Middle	11	-21.8	129.7	4.0				
Lower	12	-40.2	132.3	6.6				

* α95: THERE IS A 95% PROBABILITY THAT THE TRUE MEAN DIRECTION LIES WITHIN THE CONE OF CONFIDENCE AROUND THE OBSERVED MEAN. THE SMALLER THE VALUE OF α95 THE MORE RELIABLY THE MEAN HAS BEEN ESTIMATED.

Table 5.3: Phanerozoic Mean Pole Positions.

AGE	N	LATI-	LONGI-	α95	N	LATI-	LONGI-	α95
SOUTH AMERICA								
Cretaceous	12	85.6	203.0	4.6				
Jurassic	8	86.6	41.0	10.5				
Triassic	9	84.1	69.1	5.9	19	85.5	46.1	7.6
Permian	6	60.0	173.5	6.4	17	77.2	165.8	6.9
Carboniferous	7	60.7	169.2	11.5	8	58.3	175.8	30.6
Devonian	6	-4.5	135.5	27.6				
Silurian	-	-	-	-				
Ordovician	7	-16.6	159.4	27.4	7	-9.7	159.0	26.7
AFRICA								
Cretaceous	13	68.3	237.3	6.5				
Jurassic	17	67.9	251.8	5.8				
Triassic	10	66.0	246.9	9.0				
Permian	3	37.6	229.4	29.1	5	44.4	237.4	17.6
Carboniferous	E	20.0	220.0	-				
Devonian	1	0.5	205.0	-				
Silurian	-	-	-	-				
Ordovician	1	-14.0	156.0	-	3	60.5	159.8	>90.0
INDIA								
Lower Tertiary	12	29.3	275.4	13.9				
Cretaceous	26	20.6	282.3	11.2				
Jurassic	-	-	-	-				
Triassic	4	13.8	304.4	19.3				
Permian	7	-6.5	309.3	89.2				
Carboniferous	-	-	-	-				
Devonian	2	2.4	327.5	>90.0				
Silurian	1	-30.0	348.0	-				

Table 5.3: Phanerozoic Mean Pole Positions.

AGE	N	LATITUDE	LONGITUDE	α_{95}	N	LATITUDE	LONGITUDE	α_{95}
AUSTRALIA								
Cretaceous	4	53.7	336.8	10.1				
Jurassic	9	48.0	345.0	10.1				
Triassic	3	46.5	339.7	24.3				
Permian	3	41.7	307.4	10.5				
Carboniferous	2	33.7	315.0	47.3	6	69.4	337.0	27.9
Devonian	1	41.5	228.5	-	6	68.5	197.3	27.2
Silurian	2	36.0	219.1	40.7				
Ordovician	3	27.0	207.2	20.1	4	24.4	213.4	18.7
ANTARCTICA								
Cretaceous	5	83.7	130.3	8.4	6	85.1	93.8	10.1
Jurassic	7	51.6	39.1	7.5				
Ordovician	3	(13.8	202.4	25.7)				

URA = Urals

"......This data must be used with some reservations, as with most palaeomagnetic data, as no allowance was made for anisotropy and much of the data have not been adequately analysed for the presence of multicomponents. Additionally, thermal demagnetisation has not always been used, or when used, the effects of chemical changes are not always recognised in the original study. However, proper criteria would eliminate most of the available data." (Tarling, 1983, p223).

Tarling's pole positions and the palaeogeographies determined from them were used in preference to those of other workers for the following reasons. The palaeomagnetic data had been selected and processed using the methods outlined in Chapter Two when possible. Some of the pre-Permian palaeogeographies of other workers are based largely or entirely on interpretations of palaeomagnetic evidence with limited or no effective use of palaeontological and other types of geological evidence (e.g. Smith et al, 1973 and 1981). However the palaeomagnetic data may be unreliable as discussed more fully in Chapter Seven. To overcome any errors the Tarling palaeogeographies are based upon a combination of palaeomagnetic data, the distribution of climate-sensitive lithologies (Tarling, 1980) and the distribution of certain fossil species (Turner and Tarling, 1982). This introduces a degree of subjectivity into the construction of palaeogeographic maps which may lead to errors in the positioning of continents due to errors in the interpretation of the distributions of climate-sensitive lithologies. Conversely, if a number of geological factors are taken into account, the application of such geological constraints to palaeogeographic reconstructions may lead to a more sensible positioning of the continents. Hence the Tarling reconstructions were used.

In addition to the Tarling maps those of BP based on the palaeoreconstructions of workers in Chicago have also been used as a comparison. Ziegler et al (1979) give a full discussion of the choice of plate boundaries used in these palaeoreconstructions and possible queries in the position of some of the plates. Unfortunately only Mesozoic and Cenozoic palaeogeographies were available so a comparison of two complete sets of Phanerozoic results was not possible.

5.4.3. Map Projection.

The global reconstructions in Figures 6.15 a-j were generated by the computer program. The Plate Movement Program was written by Tarling and it provides most Phanerozoic palaeogeographic reconstructions, with the exception of the Cambrian period, which is currently under revision. In this program individual plates, or parts of plates, are defined in latitude and longitude co-ordinates and rotated according to the appropriate rotation into their previous positions and then replotted.

The reconstructions were plotted using a Molleweide Projection, an equal area map in which latitude lines remain parallel. It encompasses the entire surface of the globe, centred on the 0° meridian which is an arbitrary reference as absolute values of palaeolongitude are indeterminable. Palaeogrid lines have been removed from the maps in order to allow distinction of individual points more easily. However, for reference purposes, a list of the mineral deposit palaeolatitudes can be found in Appendix I. The continental outlines were based on present day coastlines and some major geological features for location purposes.

CHAPTER SIX

RESULTS

The purpose of this research was to determine if there existed a palaeolatitudinal control upon the formation of certain mineral deposit types and to use this to evaluate some of the more widely accepted models of mineral genesis. The main characteristic sought within the distribution patterns was a limitation of the range of a particular deposit type to low latitudes i.e. 30° (35° including error bars) north and south of the equator. Such a confinement to these so-called low latitudes was not chosen arbitrarily: it is generally quoted as that of the formation of evaporite deposits (Frakes, 1979 and Habicht, 1979). As many mineral deposit types are associated with evaporite deposits and other climate-sensitive lithologies thought to be related to low latitudes, this limit upon palaeolatitude was considered to be particularly important. Hence the range was tested for each of the deposit types described in Chapter Three. However any palaeolatitude control must be of interest so other obvious distribution patterns have also been noted.

In the following chapter the full ranges of mineral deposit distribution have been described noting whether palaeolatitudes are from north or south of the equator, rather than regarding the absolute palaeolatitudes irrespective of hemisphere. This method was chosen as it cannot be assumed that the geography and climatic conditions at a specific latitude are identical in both hemispheres as such conditions are greatly influenced by the distribution and relative proportions of land mass to oceans.

6.1 The Palaeolatitude Distribution of all Deposits Examined

The scatter graphs Figures 6.1a and 6.1b (minimum and maximum age-determined palaeolatitudes respectively) show the distribution of palaeolatitudes for all the mineral deposit examples used plotted against their age of mineralisation. Both graphs show the characteristics listed below.

a) The palaeolatitudes range from 73° south to 65° north.

b) There is a marked gap in the distribution between 30° north and 30° south from 200 to 150 million years, that is the Triassic to Lower Jurassic. Three main explanations are proposed for this hiatus:

1. the land area in the region during that period of time was a small proportion of the total land mass of the Earth,

2. few sediments were deposited in this region from the Triassic to the Lower Jurassic period,

3. the deposits which originally formed may have been eroded. The phenomenon of selective preservation of mineral deposits must always be taken into account in the consideration of distribution patterns.

c) There is a slight suggestion of clustering of the points;

(i) from the equator to 10° north between 280 and 220 million years,

(ii) from 200 million years to the present in the region 20° to 50° north. This probably represents the paucity of all sediments of this age in the southern hemisphere. It may also be due to more widespread exploration in most northern continents relative to those in the southern hemisphere.

d) In the southern hemisphere from 360 to 150 million years a marked decrease in the latitudinal range of mineral deposits is apparent i.e. the most southern position occurs between 30° and 35° south. In the northern hemisphere the most northern limit is 30° N within the age range 450 to 220 million years. This could be due to the influence of continental distribution; i.e. a greater

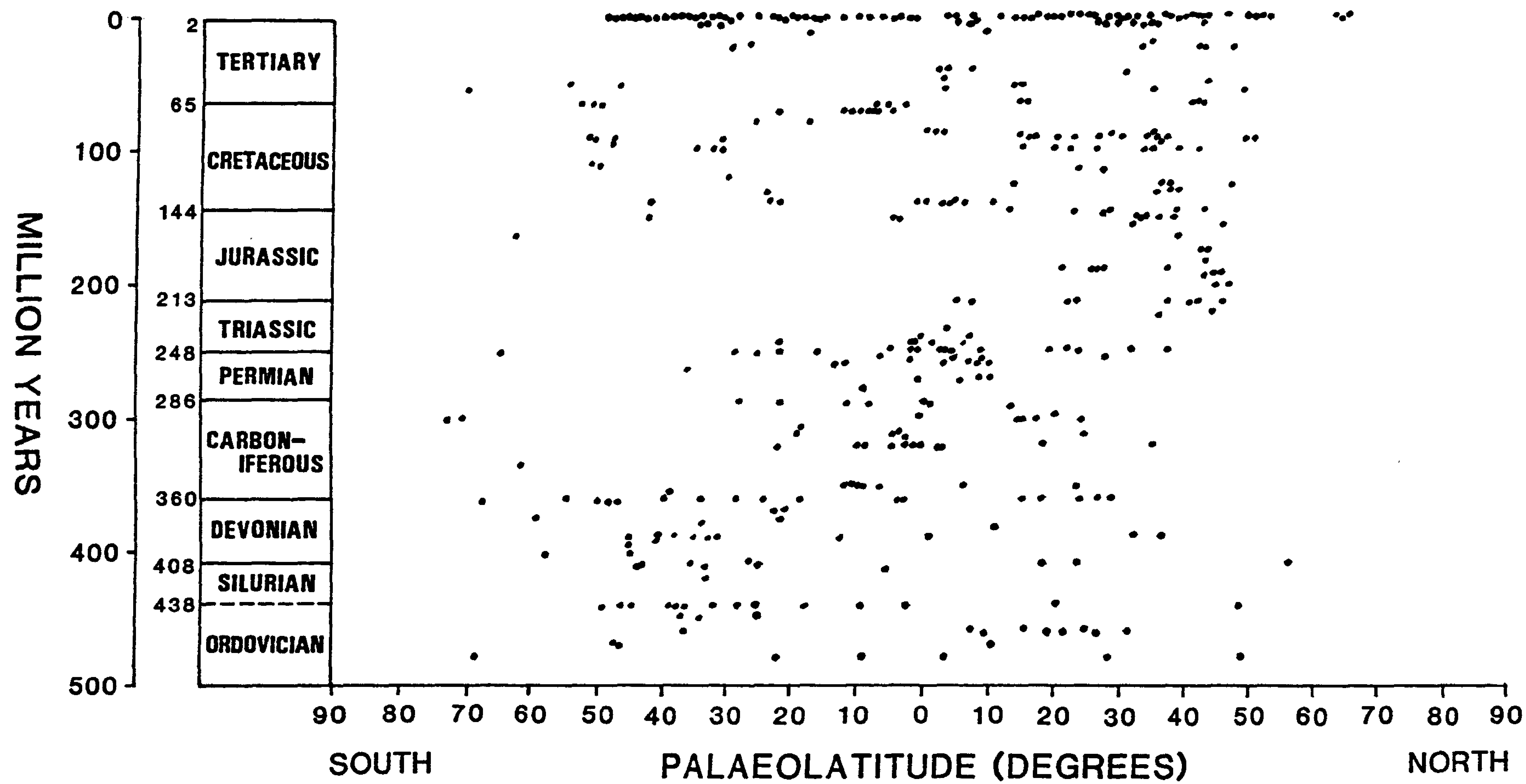


Figure 6.1a. Palaeolatitudes determined from minimum ages of mineralisation for all mineral deposit examples used. The points are not a representation of deposit size or of age range.

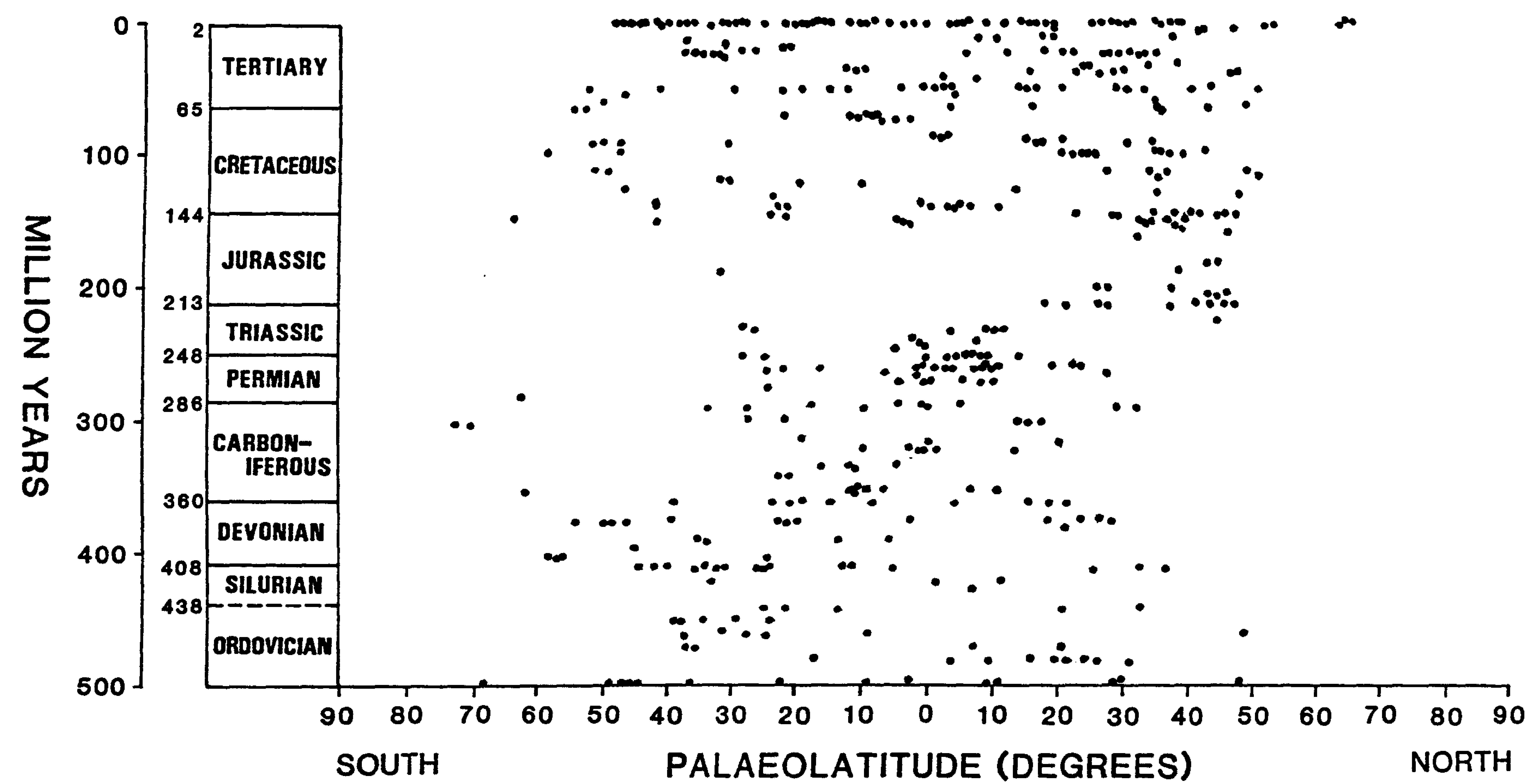


Figure 6.1b. Palaeolatitudes determined from maximum ages of mineralisation for all mineral deposit examples used. The points are not a representation of deposit size or of age range.

proportion of continental land area in the northern than in the southern hemisphere during more recent times.

e) Figure 6.1a shows two hiati in the deposit distribution pattern between 100 million years and the present in the two latitudinal bands; 15° to 30° north and from the equator to all southern latitudes. However these gaps are indistinguishable in the maximum plot, Figure 6.1b, so the maximum ages for these particular deposits may be more appropriate assuming a low latitude control upon the formation of certain deposit types. At this stage a new aspect of the possible palaeolatitude control on the formation of mineral deposits has been introduced. It may be that such a control could be used to solve certain geological problems - in this instance the correct age of mineralisation.

In conclusion there is a suggestion of a palaeolatitude control upon the mineral deposit types under consideration as these diagrams illustrate an uneven distribution of examples. In particular the deposits have a limited palaeolatitude range and there are areas (in both time and latitudinal extent) where either a paucity or a concentration of deposits occurs. An examination of the distribution of individual mineral deposit types may outline any possible palaeolatitude control hinted at in this more general approach.

The Form of the Diagrams

Before describing the results it is necessary to explain the form of some of the diagrams. A number of the examples had poorly defined ages of mineralisation: i.e. a wide range between the oldest and youngest possible ages. Such deposits had to be rotated more than once; for example the Irish SDEX deposits (e.g. Keel) dated as 360 to 320 million years old. Their palaeolatitudes were determined from both the 300 and 350 million year palaeogeographic reconstructions, giving values of 3.5°N and 19.2°S respectively. So the problem of assessing and illustrating two or more different palaeolatitude values had to be overcome. The histograms of Palaeolatitude versus Frequency (% of examples) for each mineral deposit type illustrate one possible solution to this problem. There are two diagrams, a histogram of the palaeolatitudes determined from the minimum ages of mineralisation and one of the palaeolatitudes determined from the maximum mineralisation ages. These will be referred to as the minimum and maximum histograms and both are displayed on a single page to aid comparison. In the majority of cases both hemispheres have been shown. However for some mineral deposit types the whole range of palaeolatitudes from the southern and northern hemispheres have been displayed to highlight patterns in the distribution.

The annotation for the histograms Figures 6.2 - 6.14 also needs some clarification. The number of deposits represented by each column is given at the top. The letter at the base of each column refers to the list of deposits which comprise that particular column - given on the same diagram or on the facing page. The figures in brackets shown with these column headings are the ages of the reconstructions from which the palaeolatitudes were determined (in millions of years). The striped section of a column denotes deposits with a palaeolatitude in the northern hemisphere, the shaded section represents palaeolatitudes in the southern hemisphere.

6.2 CARBONATE-HOSTED DEPOSIT TYPES

6.2.1 Limestone Base-Metal Deposits (LSBM)

The palaeolatitudes determined for LSBM deposits extend from 30° south to 50° north and more examples plot in the northern than the southern hemisphere (Figure 6.2.i). The two peaks in the distribution are more obvious from the histogram (b) than the histogram (a) (Figure 6.2.ii). These peaks are found in the equatorial rainfall belt i.e. between 10° south and 10° north and the temperate rainfall belt, 25° to 50° north. It appears that a latitudinal

Figure 6.2 (i): LimestoneBase-Metal Deposits

Minimum Age Histograms

a -	Austria	(250)
	Benue Trough, Nigeria	(50)
	Italy	(250)
	Kansas, USA	(300)
	Missouri, USA	(300)
	Nova Scotia, Canada	(300)
	Oklahoma, USA	(300)
	Yugoslavia	(250)
b -	Benue Trough, Nigeria	(50)
	Czechoslovakia	(250)
	Missouri, USA	(300)
	Oklahoma, USA	(300)
	Poland	(250)
	Kazakhstan, USSR	(350)
c -	Robb Lake, Canada	(350)
d -	Illinois, USA	(400)
	Pine Point, Canada	(300)
	Wisconsin, USA	(400)
e -	Bou Beker, Morocco	(0)
	Tunisia	(50)
f -	Illinois, USA	(100)
	Kentucky, USA	(100)
	Tunisia	(0)
g -	Bulgaria	(200)
	Czechoslovakia	(200)
	Italy	(0)
h -	Poland	(200)

Maximum Age Histograms

A -	Austria	(250)
	Italy	(250,300)
	Kansas, USA	(300)
	Missouri, USA	(300)
	Oklahoma, USA	(300)
	Yugoslavia	(250)
B -	Austria	(250)
	Benue Trough, Nigeria	(100)
	Czechoslovakia	(250)
	Kansas, USA	(350)
	Missouri, USA	(300,350)
	Oklahoma, USA	(300)
	Poland	(250,300)
	Kazakhstan, USSR	(350)
C -	Benue Trough, Nigeria	(100)
	Pine Point, Canada	(350)
D -	Robb Lake, Canada	(350)
E -	Nova Scotia, Canada	(350)
	Tunisia	(50)
F -	Bou Beker, Morocco	(200)
	Illinois, USA	(400)
	Wisconsin, USA	(400)
G -	Czechoslovakia	(200)
	Illinois, USA	(100)
	Italy	(200)
	Kentucky, USA	(100)
H -	Bulgaria	(200)
I -	Poland	(200)

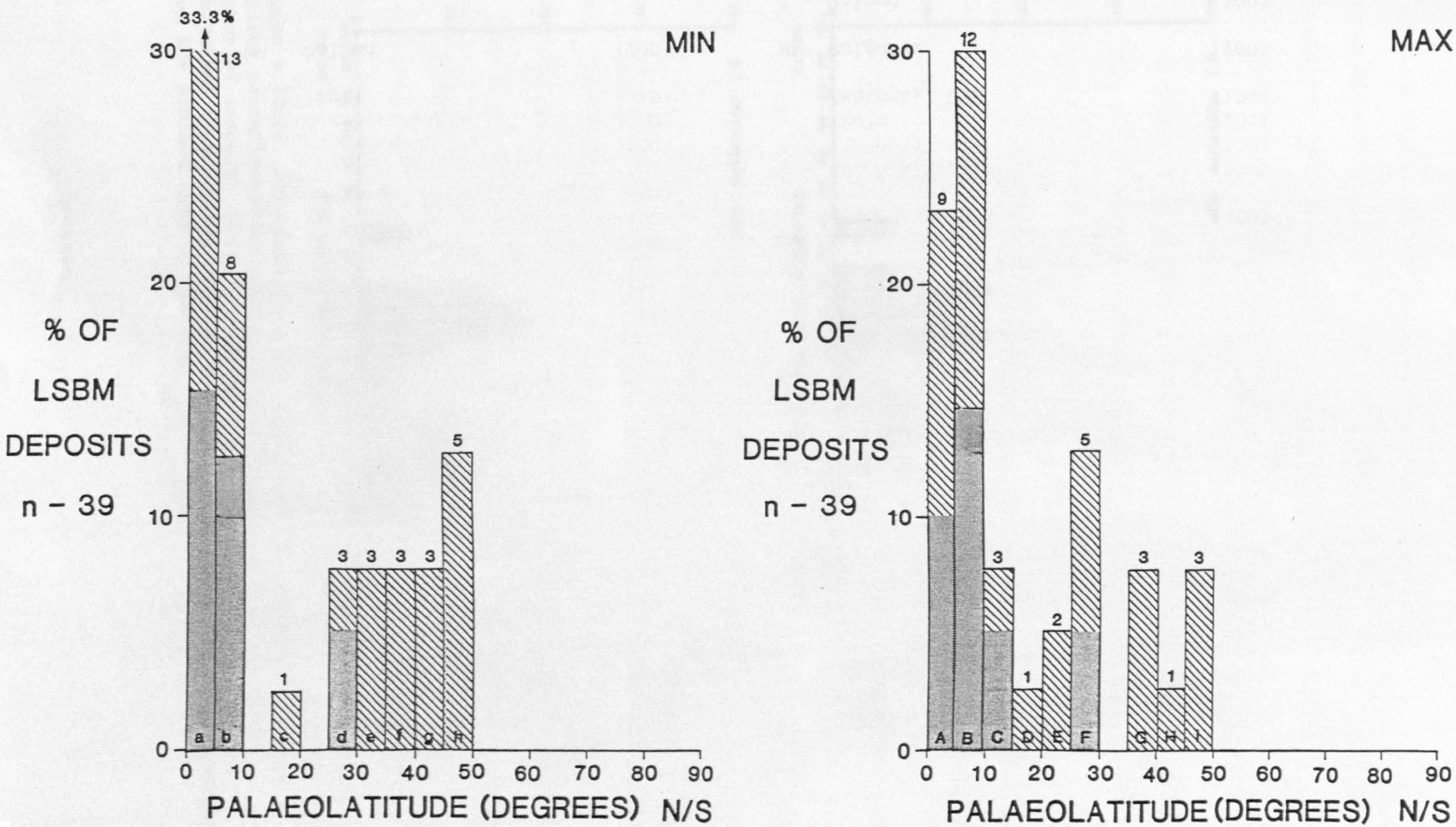


Figure 6.2(i). Palaeolatitude vs. Frequency for LSBM deposits from Tarling reconstructions. Shaded section= palaeolatitude in southern hemisphere, striped section=palaeolatitude in northern hemisphere. Figure at top of the columns represents number of examples in each latitude range. MIN=palaeolatitudes determined from minimum ages, MAX=palaeolatitudes determined from maximum ages of examples.

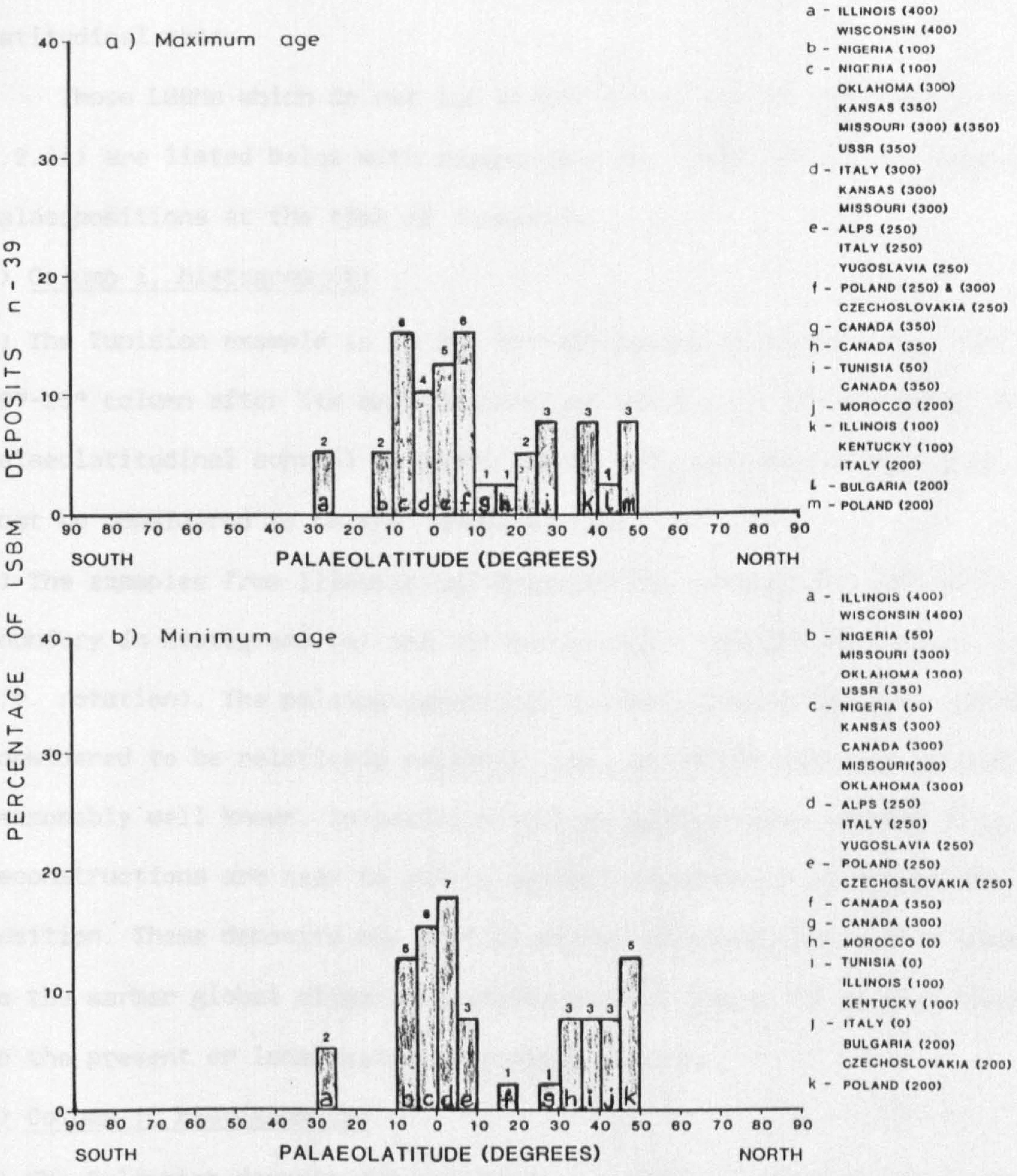


Figure 6.2(ii). Palaeolatitude vs. Frequency for LSBM deposits from Tarling reconstructions. Figure at top of columns represents the number of examples in each latitude range. a) palaeolatitudes determined from maximum ages of deposits. b) palaeolatitudes determined from the minimum ages of deposits.

control upon the formation of LSBM deposits is likely as the minimum histogram shows approximately 63% of deposits lie within 30° of the equator and almost 71% occur from 35° north to 35° south. The maximum histogram also shows this low latitude control as 82% of LSBMs lie in both the 30° and 35° north/south latitudinal zones.

Those LSBMs which do not lie within 30° to 35° of the equator (Figure 6.2.ii) are listed below with suggestions for their apparently anomalous palaeopositions at the time of formation.

1) Column i, histogram (b)

a) The Tunisian example is in the 35°-40° column at present but lies in the 20°-25° column after its maximum rotation (50 m.y.). If the 30° to 35° palaeolatitudinal control on LSBMs exists then the maximum age of the deposit must be considered to be the more 'correct'.

b) The examples from Illinois and Kentucky lie outside the 30° latitudinal boundary in histograms (a) and (b) occurring in the 35°-40° column (after a 100 m.y. rotation). The palaeogeographical reconstructions for this age are considered to be relatively reliable. The age of the deposits is also reasonably well known. In addition, the palaeolatitudes derived from the BP reconstructions are near to 40° so another explanation is needed for their position. These deposits may plot at higher palaeolatitudes than expected due to the warmer global climatic conditions which prevailed at that time relative to the present or local palaeogeographic effects.

2) Column j, histogram (b)

a) The Bulgarian deposit determined from a 200 m.y. rotation plots in the 40°-45° column. The age of this deposit is reasonably reliable but its position could be in some doubt because of differential plate motion in this region of Europe. However if the position of this area were incorrect then a number of other results would then be in doubt. The BP reconstruction produced a palaeolatitude of 35° which is more in keeping with the latitudinal limits of the majority of LSBMs. Perhaps the BP palaeoposition for this continent is more 'valid' than that of Tarling.

b) The Czechoslovakian deposits determined from a 200 m.y. rotation plot in the 40°-45° column. However the palaeolatitude determined from a 250 m.y. rotation occurs in the 5°-10° column in histogram (a). So the oldest age would be more acceptable in order for these LSBMs to comply with the 35° palaeolatitude limit suggested by the other examples.

3) Column k, histogram (b)

The Polish deposits plot in the 45°-50° column with a 200 m.y. rotation, but in the 5°-10° column after 250 m.y. rotation. However the age of these deposits is in some doubt as there may be three periods of mineralisation: the end of the Palaeozoic (250 m.y.), the Muschelkalk to the Mid Jurassic (243-180 m.y.) and around 5 m.y. ago (Sass-Gustkiewicz et al, 1982). So no satisfactory explanations can be offered for the palaeopositions of these deposits. But if there is a palaeolatitude control then the end of the Palaeozoic age seems the most likely.

6.2.2 Oolitic Ironstone Deposits (OOFE)

The OOFE palaeolatitudes extend from 50° north to 50° south (Figure 6.3) with two obvious sub-sets i.e. 0° to 15° and 30° to 50° north/south of the equator. There is a marked absence of deposits from 15° to 30° north and south which is also reflected in the BP results. The majority of OOFEs (about 70%) lie in the mid-latitude sub-set of 30° to 50° north and south. More palaeolatitudes plot in the northern than in the southern hemisphere.

It is only possible to draw tentative conclusions about this particular mineral deposit type as a small sample has been examined (n = 26). However there is a clear latitudinal control upon the formation of OOFEs with a similar range to that seen in the LSBM results. The correlation could be with

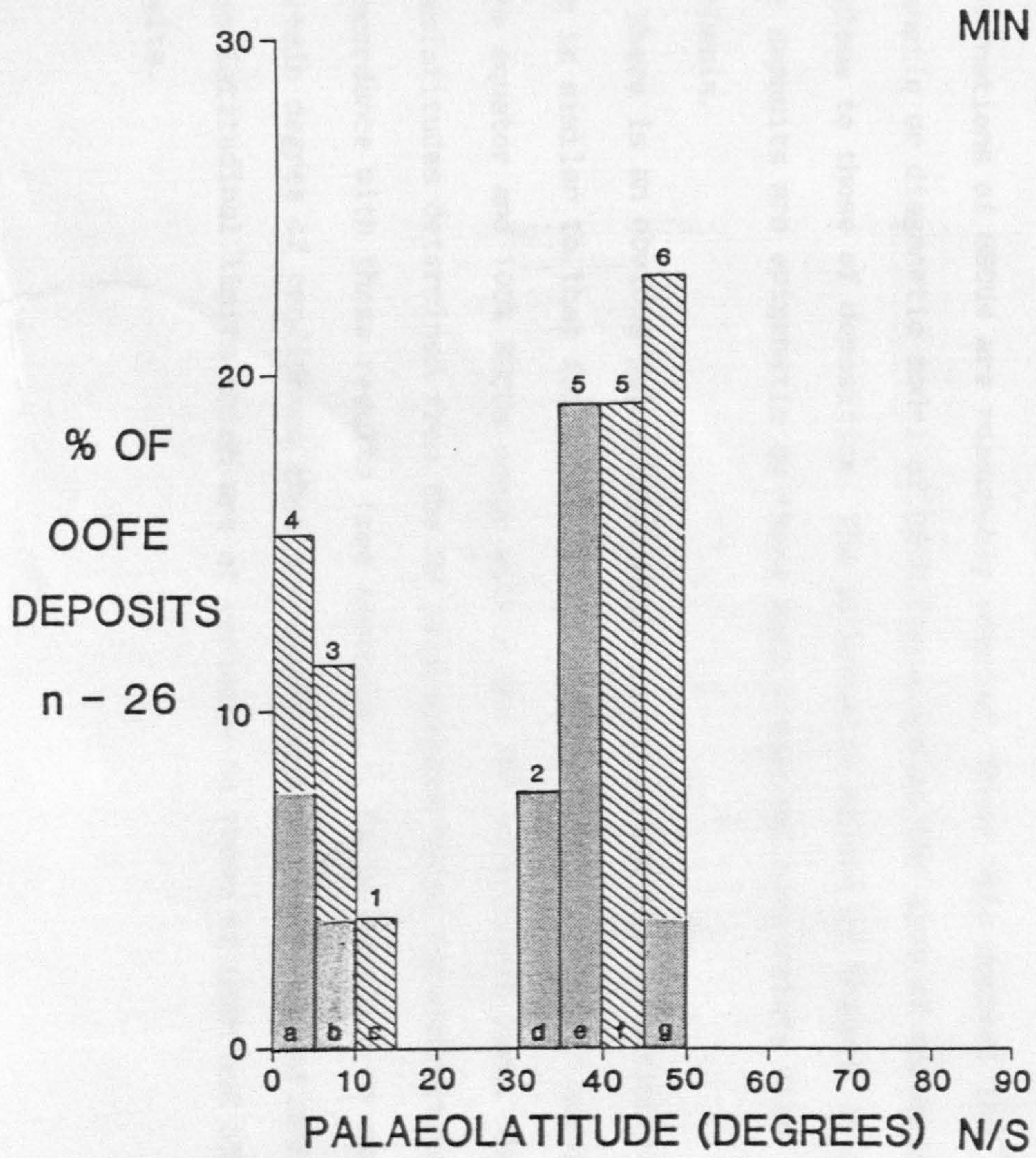
Figure 6.3: Oolitic Ironstone Deposits

Minimum Age Histogram

a -	Colombia	(50)
	Nigeria	(50)
	Saudi Arabia	(50)
b -	Colombia	(0,50)
	Nigeria	(50)
c -	Venezuela	(0)
d -	Alabama, USA	(400)
	Australia	(0)
e -	England	(100)
	Spain	(450)
f -	England	(200)
g -	England	(130,200)
	Newfoundland	(450)

Maximum Age Histogram

A -	Colombia	(50)
	Nigeria	(50)
	Saudi Arabia	(50)
B -	Colombia	(0,50)
	Nigeria	(50)
C -	Venezuela	(0)
D -	Alabama, USA	(400)
	Australia	(0)
E -	Spain	(450)
F -	England	(200)
G -	England	(130,200)
	Newfoundland	(450)



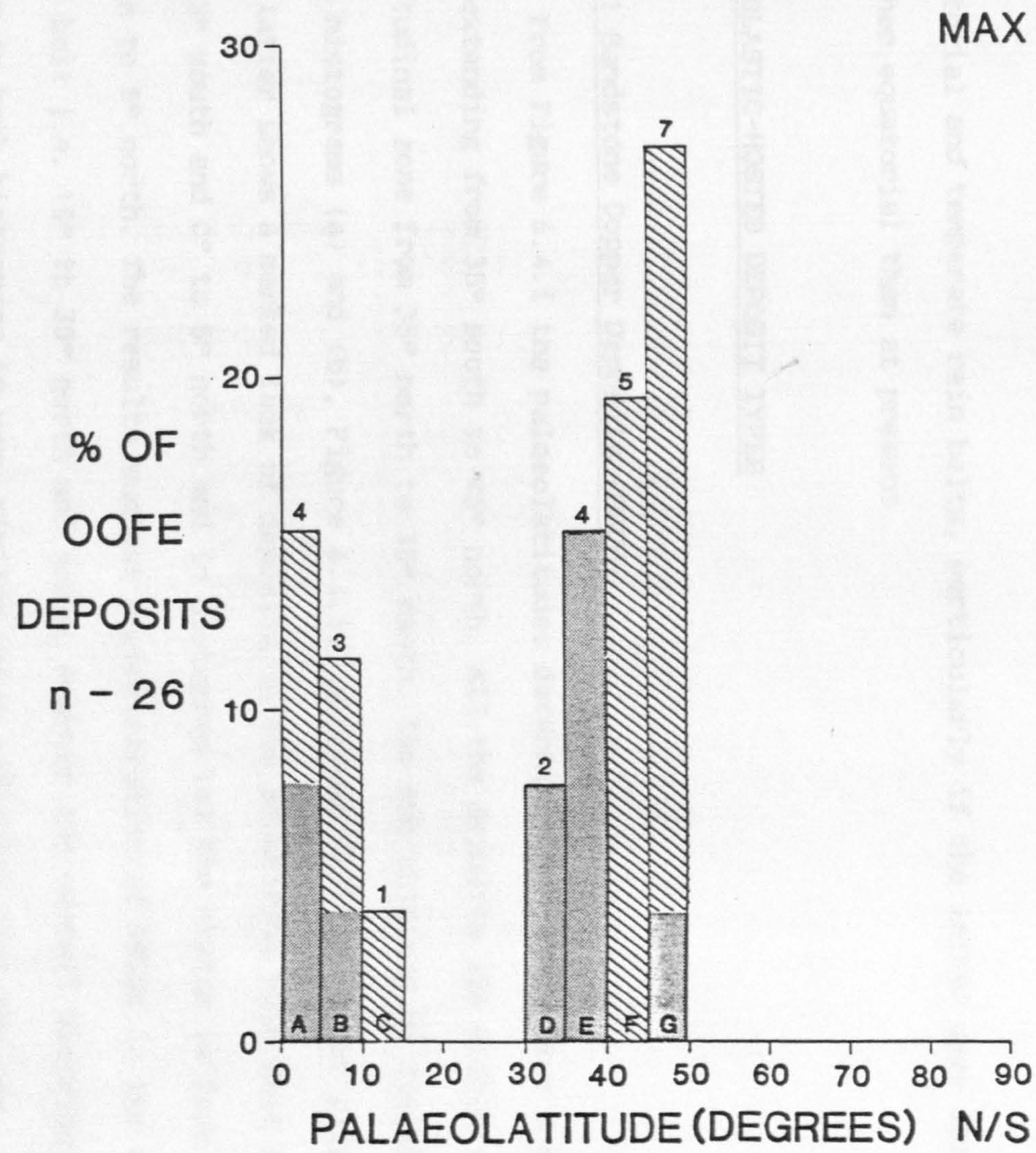


Figure 6.3. Palaeolatitude vs. Frequency for OOFE deposits from Tarling reconstructions.
For figure legend see Figure 6.2(i).

equatorial and temperate rain belts, particularly if the latter were somewhat further equatorial than at present.

6.3 CLASTIC-HOSTED DEPOSIT TYPES

6.3.1 Sandstone Copper Deposits (SSCU)

From Figure 6.4.i the palaeolatitudes determined for SSCUs have a range of 60° extending from 35° south to 25° north. All the deposits lie within the latitudinal zone from 35° north to 35° south. Two sub-sets can be distinguished from histograms (a) and (b), Figure 6.4.ii, although they do differ slightly. The latter shows a marked lack of deposits in the equatorial rain belt from 5° to 15° south and 0° to 5° north and in histogram (a) the hiatus is from 20° south to 5° north. The results suggest a concentration of SSCUs in the warm arid belt i.e. 15° to 35° north and south. However the overall distribution of SSCUs in both histograms is very similar which indicates that the age determinations of SSCUs are reasonably precise. These data support the syngenetic or diagenetic model of SSCU formation as the ages of mineralisation are close to those of deposition. The alternative school of thought is that these deposits are epigenetic so these data presented here refute this hypothesis.

There is an obvious palaeolatitudinal control upon the formation of SSCUs which is similar to that for LSBMs. There are 89% SSCUs within 30° north/south of the equator and 100% SSCUs occur within the 35° north/south zone. The palaeolatitudes determined from the BP palaeogeographical reconstructions are in accordance with these results (see Appendix I). It is possible to state with a certain degree of confidence that the occurrence of SSCUs is confined to strict latitudinal limits which are at variance to those of LSBM and OOFE deposits.

Figure 6.4 (i): Sandstone Copper Deposits

Minimum Age Histogram

a	Nova Scotia, Canada	(300)
b	New Brunswick, Canada	(300)
c	New Mexico, USA	(250)
d	Algeria	(100)
	Bolivia	(0)
	Morocco	(100)
e	Angola	(100)
	New Mexico, USA	(200)
f	Angola	(100)
g	Angola	(100)

Maximum Age Histogram

A	New Mexico, USA	(250)
B	New Mexico, USA	(250)
C	Algeria	(100)
	Morocco	(100)
D	Angola	(130)
	Bolivia	(50)
	New Brunswick, Canada	(350)
	Nova Scotia, Canada	(350)
E	Angola	(100)

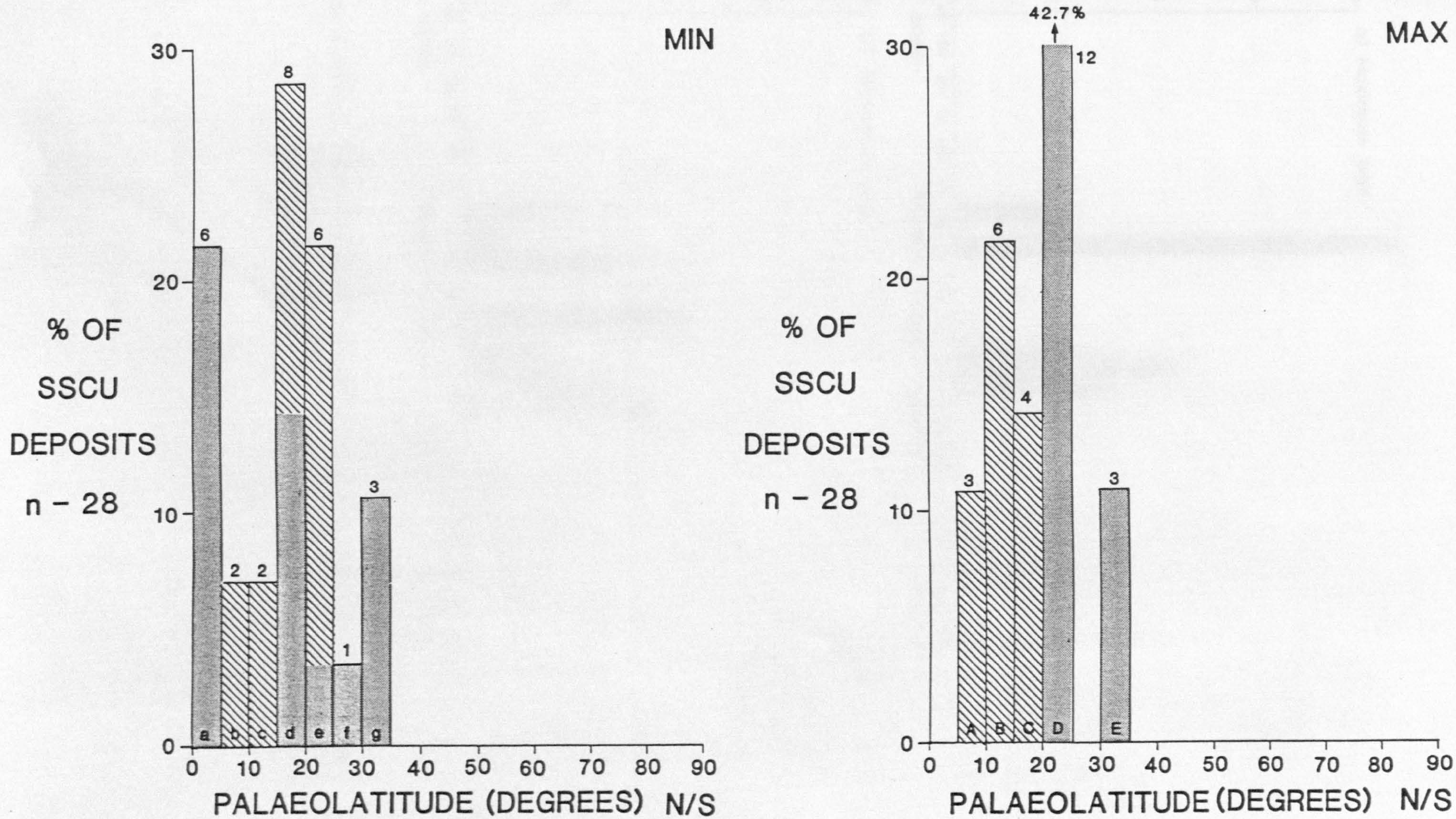


Figure 6.4(i). Palaeolatitude vs. Frequency for SSCU deposits from Tarling reconstructions.
For figure legend see Figure 6.2(i).

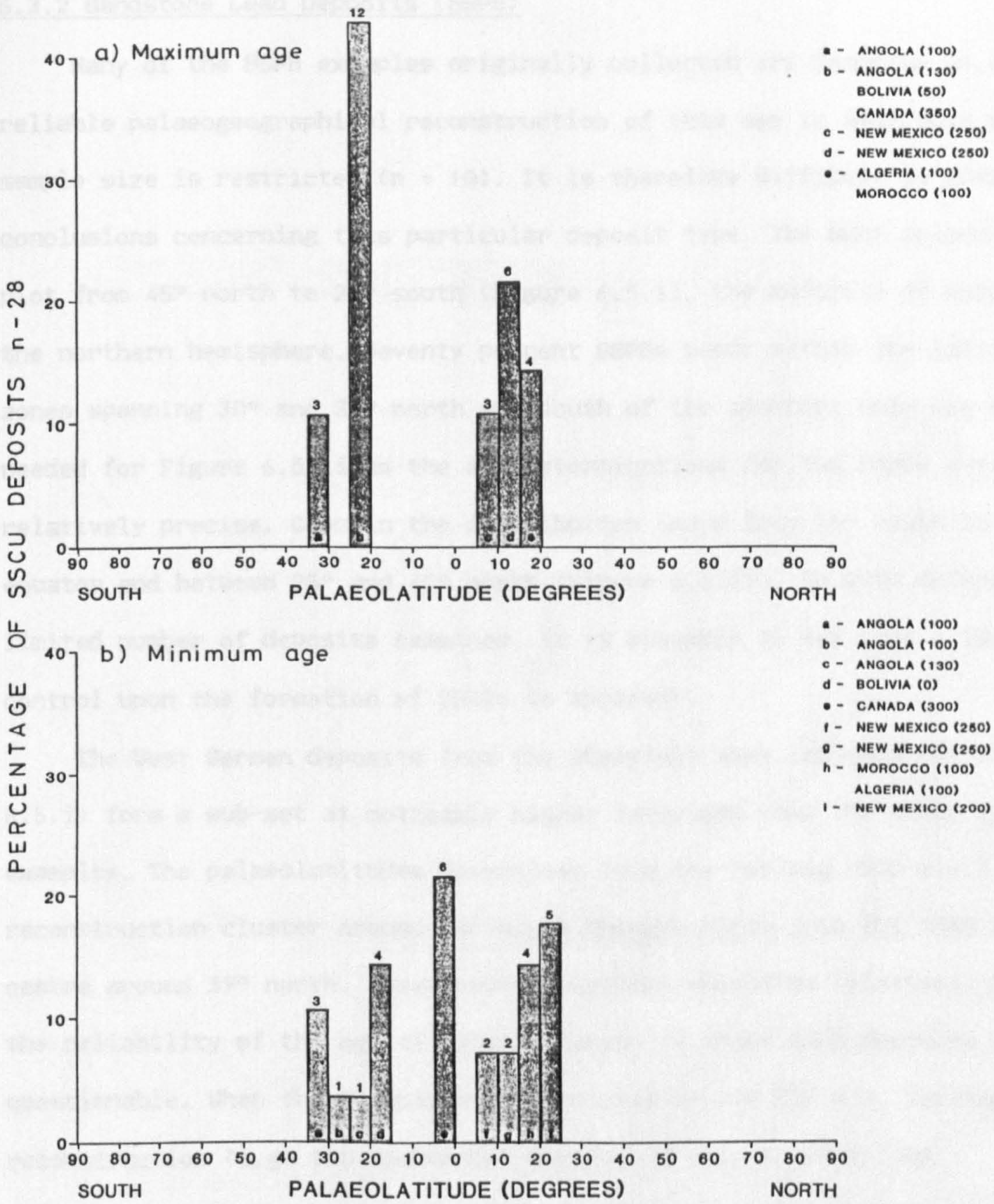


Figure 6.4(ii). Palaeolatitude Vs. Frequency for SSCU deposits from Tarling reconstructions.
For figure legend see Figure 6.2(ii).

6.3.2 Sandstone Lead Deposits (SSPB)

Many of the SSPB examples originally collected are Cambrian in age but no reliable palaeogeographical reconstruction of this age is available so the sample size is restricted (n = 10). It is therefore difficult to draw any firm conclusions concerning this particular deposit type. The SSPB palaeolatitudes plot from 45° north to 25° south (Figure 6.5.i), the majority of which lie in the northern hemisphere. Seventy percent SSPBs occur within the latitudinal zones spanning 30° and 35° north and south of the equator. Only one diagram was needed for Figure 6.5.ii as the age determinations for the SSPBs were relatively precise. Gaps in the distribution occur from 15° south to the equator and between 25° and 40° north (Figure 6.5.ii). So with deference to the limited number of deposits examined, it is possible to say that a latitudinal control upon the formation of SSPBs is apparent.

The West German deposits from the Oberpfalz area (columns e/E Figure 6.5.i) form a sub-set at noticably higher latitudes than the other SSPB examples. The palaeolatitudes determined from the Tarling (200 m.y.) reconstruction cluster around 44° north whereas those from BPs (200 m.y.) map centre around 39° north. These reconstructions should be relatively reliable so the reliability of the age of mineralisation of these SSPB deposits may be questionable. When these deposits are rotated by the 250 m.y. Tarling reconstruction (e.g. the Mechernich deposit of West Germany) the palaeolatitudes plot around 6° north. If there is a low latitude control upon their formation these deposits may be older than previously thought. Conversely the SSPB deposits may be formed within different latitude limits to those deposits previously described.

6.3.3 Shale Base-Metal Deposits (SHBM)

The palaeolatitudinal extent of SHBM deposit (Figure 6.6.i) is 45° north to 35° south, with the majority of deposits plotting in the northern

Figure 6.5 (i): Sandstone Lead Deposits

Minimum Age Histogram		Maximum Age Histogram	
a - L`Argentiere, France	(250)	A - L`Argentiere, France	(250)
b - High Rolls, USA	(250)	B - High Rolls, USA	(250)
W.Germany	(250)	W. Germany	(250)
c - Morocco	(100, 300)	C - Morocco	(100, 300)
d - Angola	(130)	D - Angola	(130)
Morocco	(100)	Morocco	(100)
e - W. Germany	(200)	E - W. Germany	(200)

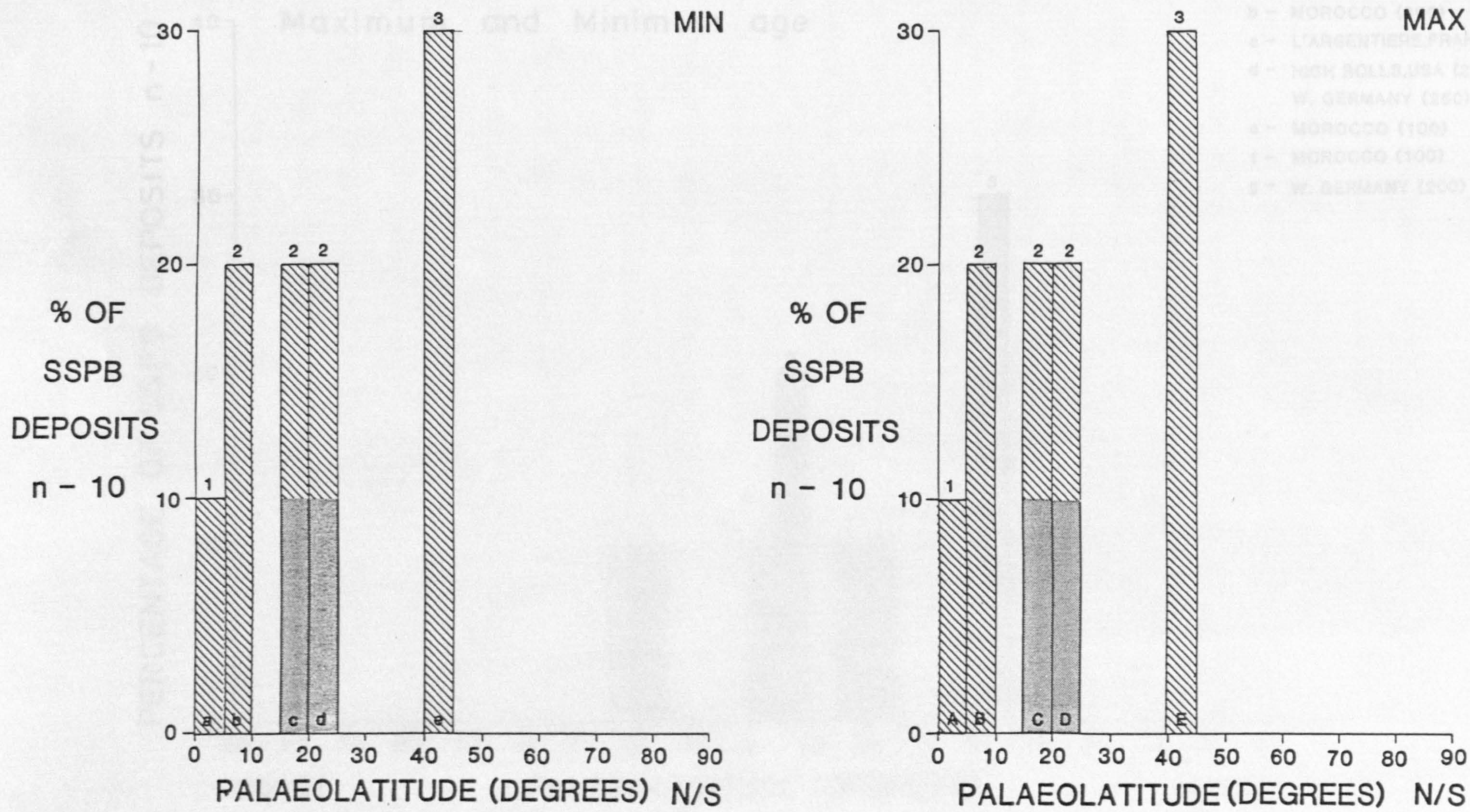


Figure 6.5(i). Palaeolatitude vs. Frequency for SSPB deposits from Tarling reconstructions. For figure legend see Figure 6.2(i).

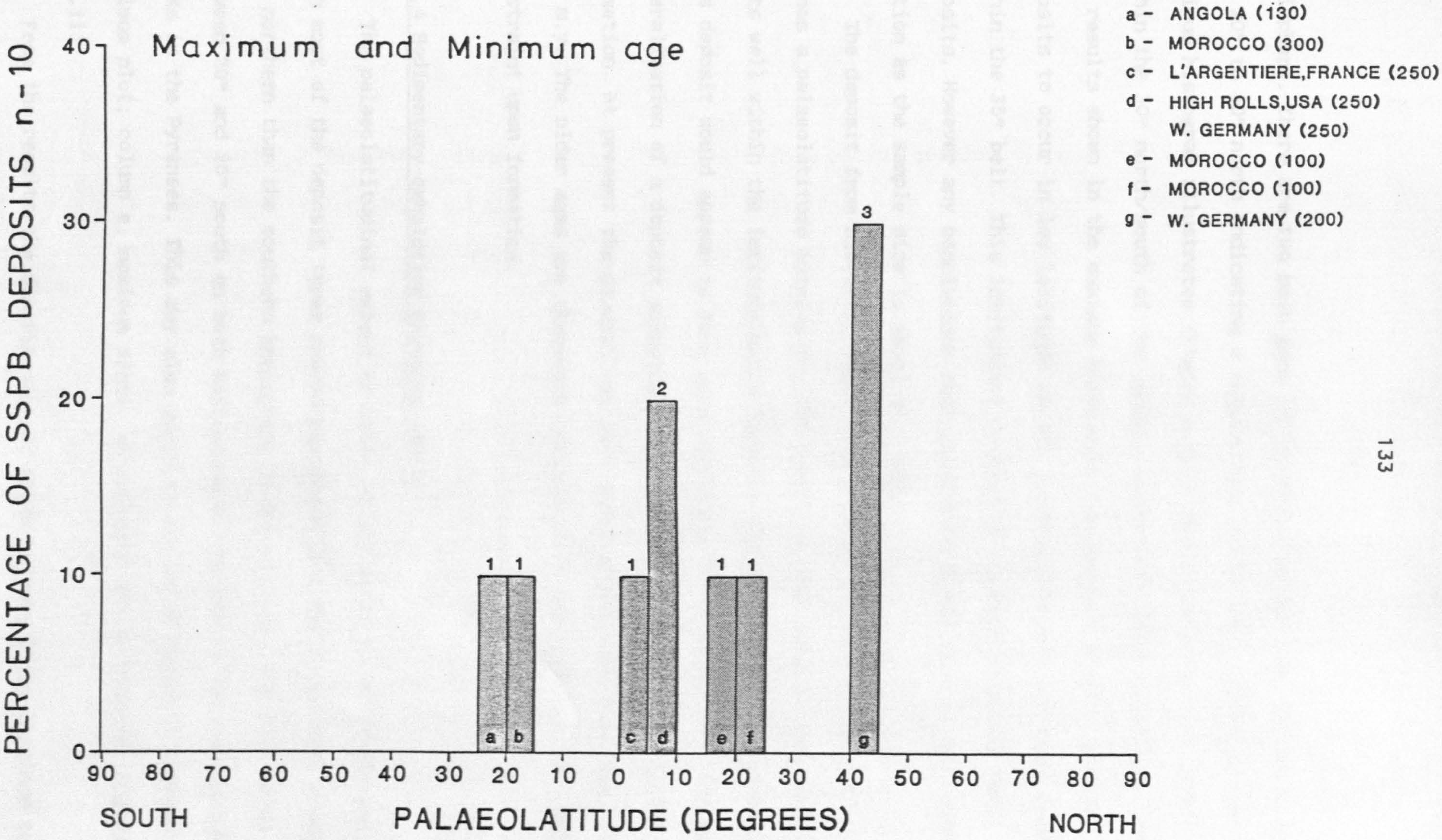


Figure 6.5(ii). Palaeolatitude vs. Frequency for SSPB deposits from Tarling reconstructions. Figure at top of columns represents the number of examples in each latitude range.

hemisphere. There are two main gaps in the distribution i.e. 30° south to 0° and 20° to 40° north indicating a suppression in the warm, arid regions. The minimum histogram illustrates (Figure 6.6.i) that 89% of the SHBM sample lie within the 30° north/south of the equator whilst 94% occur within the 35° belt. The results shown in the maximum histogram illustrate a greater tendency for deposits to occur in low latitudes as 94% occur within 30° north/south and 100% within the 35° belt. This latitudinal control is as obvious as for SSCU deposits. However any conclusions made concerning SHBMs must be made with some caution as the sample size is small (n = 18).

The deposit from Slovenia, Yugoslavia (Figure 6.6.i, minimum plot, column e) has a palaeolatitude outside the 35° limit, at 45°. However the same deposit plots well within the latitude belts from the 250 and 300 m.y. reconstructions. This deposit would appear to be a good candidate for a revision of the age of mineralisation of a deposit assuming a low latitude control upon deposit formation. At present the deposit has been given a wide age range i.e. 286 to 213 m.y. The older ages are therefore preferable to maintain the 35° latitude constraint upon formation.

6.3.4 Sedimentary-exhalative Deposits (SDEX)

The palaeolatitudinal extent of SDEXs is 30° north to 45° south and, as with most of the deposit types previously described, there are more examples in the northern than the southern hemisphere (Figure 6.7.i). The high value between 20° and 25° south on both histograms may be due to the concentration of SDEXs in the Pyrenees. This may also apply to the Irish deposits (column f, minimum plot; column e, maximum plot). No sub-sets can be defined (Figure 6.7.ii).

From the results the distribution of SDEXs appears to be confined to the low latitude region i.e. less than 30° north and south. The minimum histograms reveal that 76% of SDEXs lie between 30° north and south of the equator and 85%

Figure 6.6 (i): Shale Base-Metal Deposits

Minimum Age Histogram		Maximum Age Histogram	
a - Texas, USA	(250)	A - Slovenia, Yugoslavia	(300)
		Texas, USA	(250)
b - England	(250)		
E. Germany	(250)		
W. Germany	(250)		
Holland	(250)		
Poland	(250)		
Oklahoma, USA	(250)		
c - Bolivia	(300)	B - England	(250)
		E. Germany	(250)
		W. Germany	(250)
		Holland	(250)
		Poland	(250)
		Oklahoma, USA	(250)
d - Bolivia	(300)	C - Bolivia	(300)
e - Selwyn Basin, Canada	(450)	D - Bolivia	(300)
f - Slovenia, Yugoslavia	(200)	E - Selwyn Basin, Canada	(450)

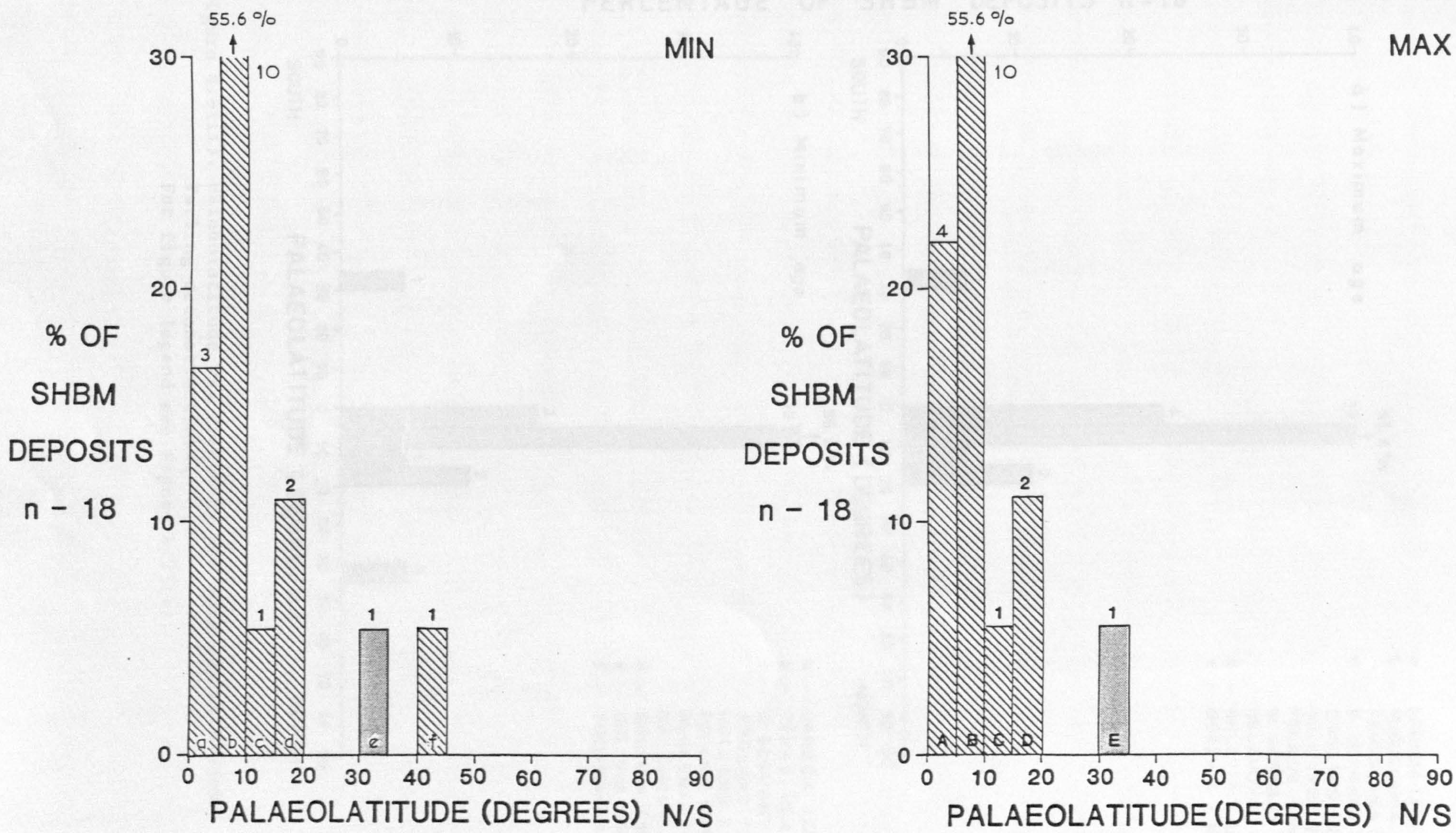


Figure 6.6(i). Palaeolatitude vs. Frequency for SHBM deposits from Tarling reconstructions.
For figure legend see Figure 6.2(i).

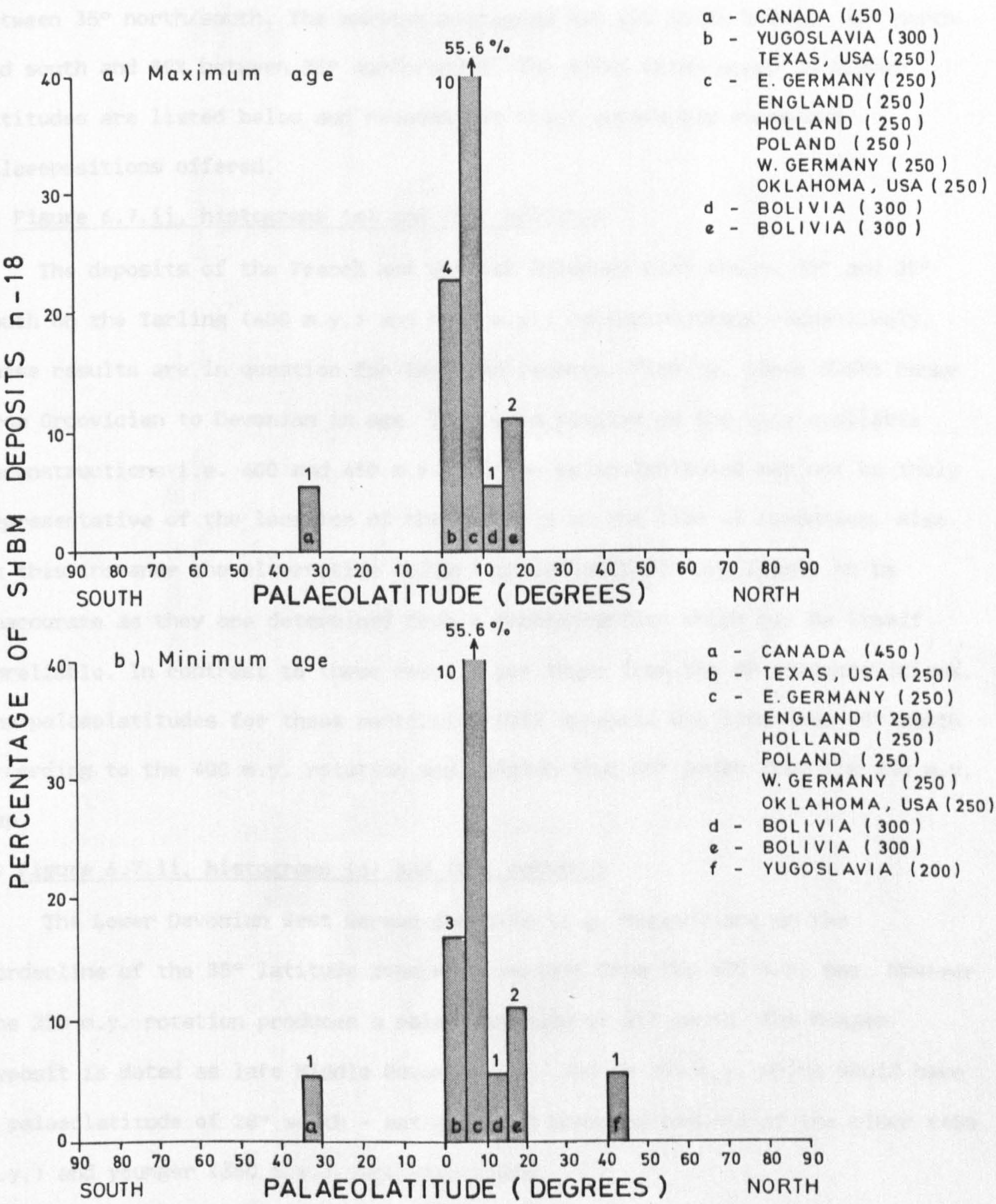


Figure 6.6(ii). Palaeolatitude vs. Frequency for SHBM deposits from Tarling reconstructions.
For figure legend see Figure 6.2(ii).

between 35° north/south. The maximum histogram has 80% SDEXs between 30° north and south and 89% between 35° north/south. The SDEXs which occur in higher latitudes are listed below and reasons for their apparently anomalous palaeopositions offered.

1) Figure 6.7.ii, histograms (a) and (b), column a

The deposits of the French and Spanish Pyrenees plot around 45° and 25° south on the Tarling (400 m.y.) and (450 m.y.) reconstructions respectively. These results are in question for two main reasons. Firstly, these SDEXs range from Ordovician to Devonian in age. They were plotted on the only available reconstructions i.e. 400 and 450 m.y., so the palaeolatitudes may not be truly representative of the location of the deposits at the time of formation. Also in this instance the alternative (older) palaeolatitudes are likely to be inaccurate as they are determined from a reconstruction which may be itself unreliable. In contrast to these results are those from the BP reconstructions. The palaeolatitudes for these particular SDEX deposits are less than 30° south according to the 400 m.y. rotation and greater than 30° south from the 450 m.y. map.

2) Figure 6.7.ii, histograms (a) and (b), column b

The Lower Devonian West German deposits (e.g. Meggen) are on the borderline of the 35° latitude zone when derived from the 400 m.y. map. However the 350 m.y. rotation produces a palaeolatitude of 21° south. The Meggen deposit is dated as late Middle Devonian i.e. 380 to 374 m.y. which would have a palaeolatitude of 28° south - extrapolated from the results of the older (400 m.y.) and younger (350 m.y.) reconstructions.

3) Figure 6.7.ii, histogram (b), column a

The Spanish SDEX deposit La Troya has a palaeolatitude greater than 40° south determined from the 400 m.y. map, whereas at 450 m.y. it plots just over 30° south. This is not an instance where more precise dating of a deposit is possible as an older reconstruction produced the more appropriate result. The

Figure 6.7 (i): Sedimentary Exhalative Deposits

Minimum Age Histogram

a - Arkansas, USA (300)
Australia (350)
Austria-Italy (300)
Bawdwin, Burma (450)
Ireland (300)
Kansas, USA (300)
Vulcan Prospect, Can (400)

b - Argentina (450)
Bo Noi, Thailand (450)
Logan Mts, Canada (450)
Rubiales, Spain (300)

c - Mae Sod, Thailand (130)
Thailand (400,450)

d - Alaska, USA (300)
Tom & Jason deps, Can (350)

e - Belgium (350)
Dzhezkazgan, USSR (350)
Italy (350)
Kazakhstan, USSR (350)
Meggen, W.Ger (350)
Pyrenees, France (450)

f - Kazakhstan, USSR (350)
Morocco (200)
Reocin, Spain (100)

g - Austria (400)
France (400)
Rammelsberg, W.Ger (400)
Spain (450)

h - Central Pyrenees (400)
La Troya, Spain (400)

Maximum Age Histogram

A - Australia (350)
Bawdwin, Burma (450)
Kansas, USA (300)
Vulcan Prospect, Can (400)

B - Argentina (450)
Bo Noi, Thailand (450)
Logan Mts, Canada (450)

C - Arkansas, USA (350)
Thailand (400,450)

D - Alaska, USA (300)
Ireland (350)
Tom & Jason deps, Can (350)

E - Austria (450)
Austria-Italy (350)
Belgium (350)
Dzhezkazgan, USSR (350)
Kazakhstan, USSR (400)

E - Mae Sod, Thailand (250)
Pyrenees, France (450)
Tynagh, Ireland (350)

F - Kazakhstan, USSR (350)
La Troya, Spain (450)
Morocco (200)
Reocin, Spain (100)
Rubiales, Spain (350)

G - Austria (400)
France (400)
Italy (400)
Rammelsberg, FRG (400)
Spain (450)

H - Meggen, W. Ger (400)

I - France (400)
Pyrenees (400)

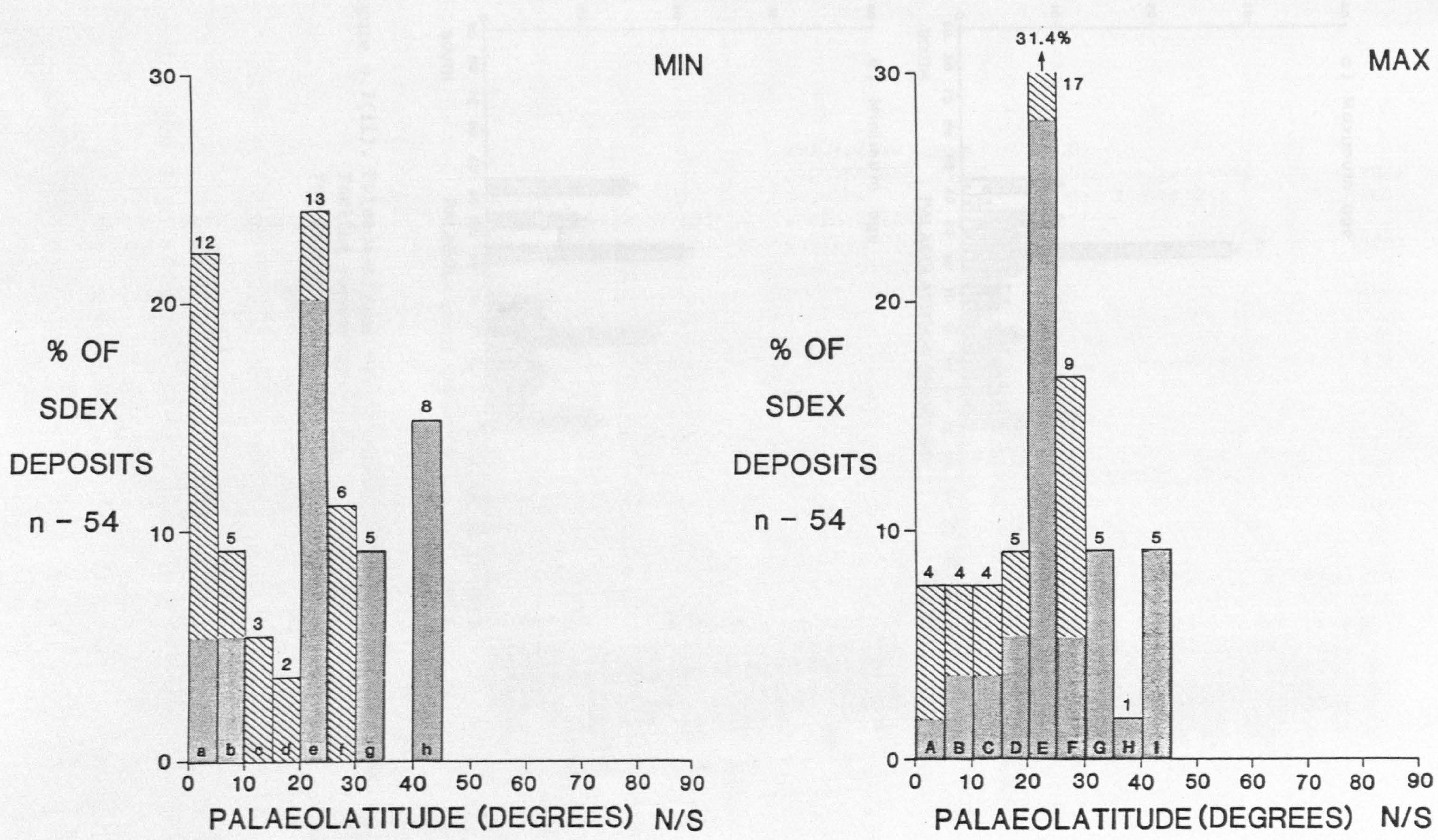


Figure 6.7(i). Palaeolatitude vs. Frequency for SDEX deposits from Tarling reconstructions.
For figure legend see Figure 6.2(i).

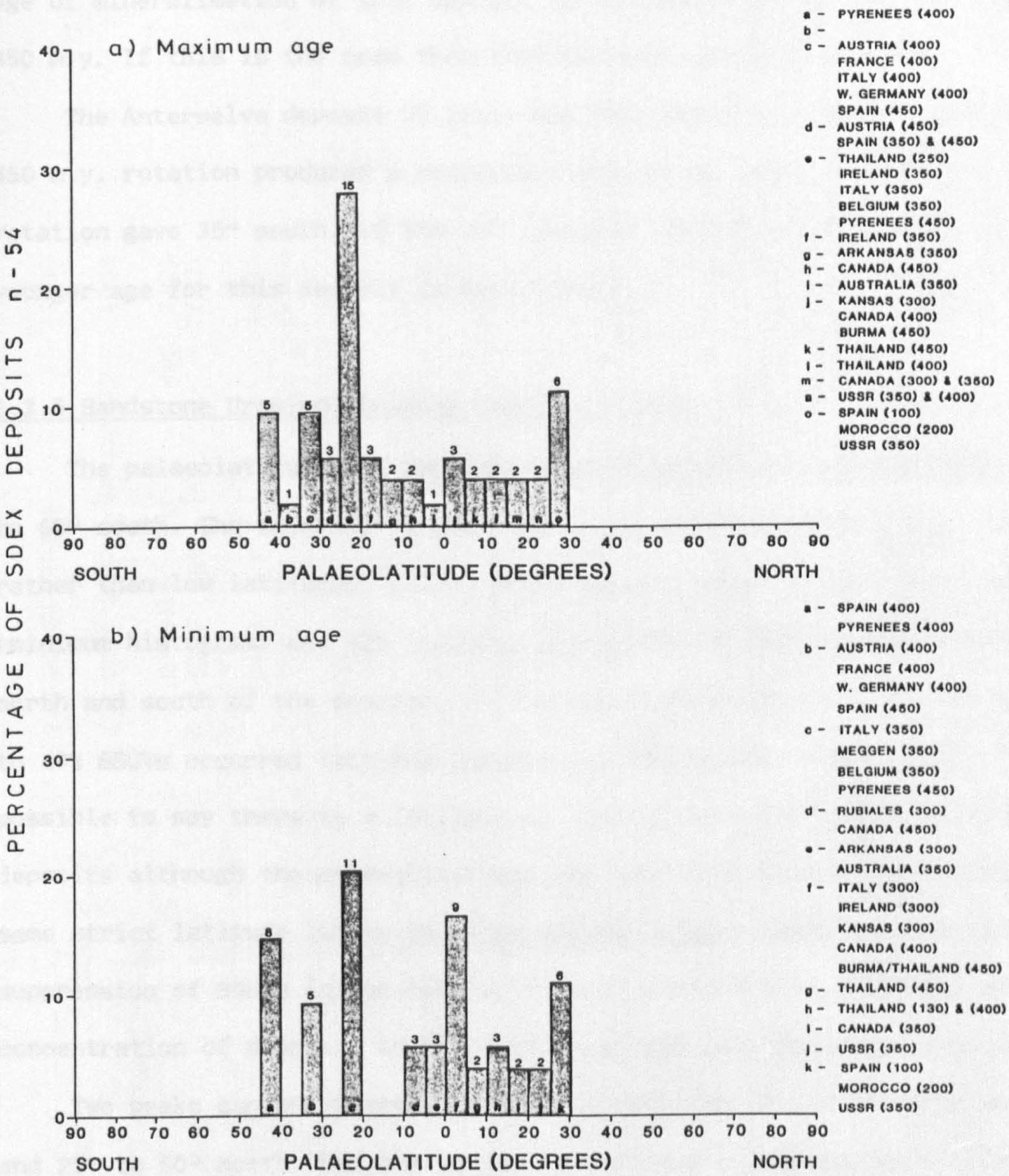


Figure 6.7(ii). Palaeolatitude vs. Frequency for SDEX deposits from Tarling reconstructions.
For figure legend see Figure 6.2(ii).

age of mineralisation of this deposit is in some doubt and may be older than 450 m.y. If this is the case then this palaeolatitude is invalid.

The Anterselva deposit of Italy has been dated as 408-360 m.y. However the 350 m.y. rotation produced a palaeolatitude of 24° south whereas the 400 m.y. rotation gave 35° south. If the 30° latitude constraint is applied then the younger age for this deposit is more likely.

6.3.5 Sandstone Uranium-Vanadium Deposits (SSUV)

The palaeolatitudes of SSUV deposits (Figure 6.8.i) extend from 50° north to 65° south. The majority of deposits occur in mid-latitudes i.e. 30° to 50° rather than low latitudes. Unlike other mineral deposit types only one third (minimum histogram) and 43% (maximum histogram) of SSUVs formed between 30° north and south of the equator. In the region from 35° south to 35° north, 46% to 49% SSUVs occurred (minimum and maximum histograms, repectively). It is possible to say there is a latitudinal control upon the formation of SSUV deposits although the palaeolatitudes are such that they do not conform to the same strict latitude limits as other deposit types. There appears to be a suppression of SSUVs in the hot, arid regions of the low latitudes and a concentration of deposits in the equatorial and warm temperate rain belts.

Two peaks can be observed in the distribution, 0° to 5° north and south and 25° to 50° north, but the latter is confined to the northern hemisphere (Figure 6.8.ii). This result could be due to the fact that more examples plot in the northern than the southern hemisphere so peaks in the distribution patterns are emphasised. In addition this asymmetry of distribution may reflect a sampling bias. The palaeolatitudes for SSUVs determined from the BP reconstructions follow a similar trend to that described for Tarling's results. Although many of the deposits plot in even higher palaeolatitudes with the BP programs.

Figure 6.8 (i): Sandstone Uranium-Vanadium Deposits

Minimum Age Histogram

a	Brazil	(250)
	France	(250)
	Hungary	(250)
	Italy	(250)
	Switzerland	(250)
	Yugoslavia	(250)
b	W. Germany	(250)
c	E. Germany	(300)
	N.T., Australia	(300)
	Thailand	(130)
d	India	(0, 250)
	Pakistan	(0)
e	Beverley, Australia	(0)
	India	(0)
	Pakistan	(0)
f	Japan	(0)
	New Mexico, USA	(130)
g	Arizona, USA	(50)
	Colorado, USA	(130)
	Tyrol, Austria	(200)
	Utah, USA	(50)
	Wyoming, USA	(0)
h	Lake Frome, Australia	(50)
i	South Africa	(250)

Maximum Age Histogram

A	Australia	(350)
	Brazil	(250)
	France	(250, 300)
	Hungary	(250)
	Italy	(250)
	Switzerland	(250)
	Tyrol, Austria	(250)
	Yugoslavia	(250)
B	W. Germany	(250)
C	Colorado, USA	(250)
	E. Germany	(300)
	Thailand	(130)
D	India	(250)
	Pakistan	(0)
E	Beverley, Australia	(0)
	Pakistan	(0)
F	Japan	(0)
	New Mexico, USA	(130)
G	Arizona, USA	(130)
	Utah, USA	(130)
H	Utah, USA	(130)
	Wyoming, USA	(50, 130)
I	Lake Frome, Australia	(50)
J	South Africa	(300)

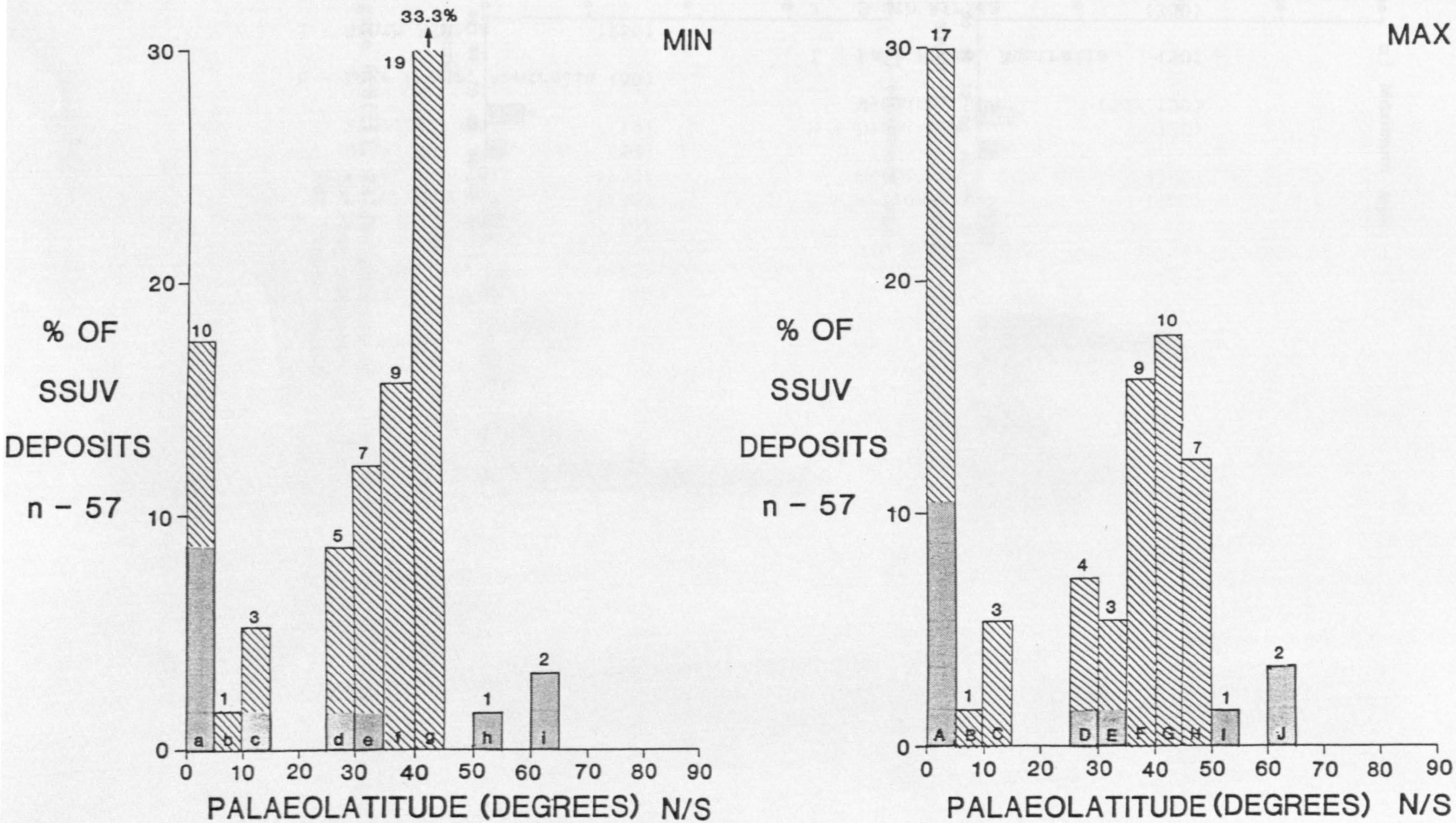


Figure 6.8(i). Palaeolatitude vs. Frequency for SSUV deposits from Tarling reconstructions. For figure legend see Figure 6.2(i).

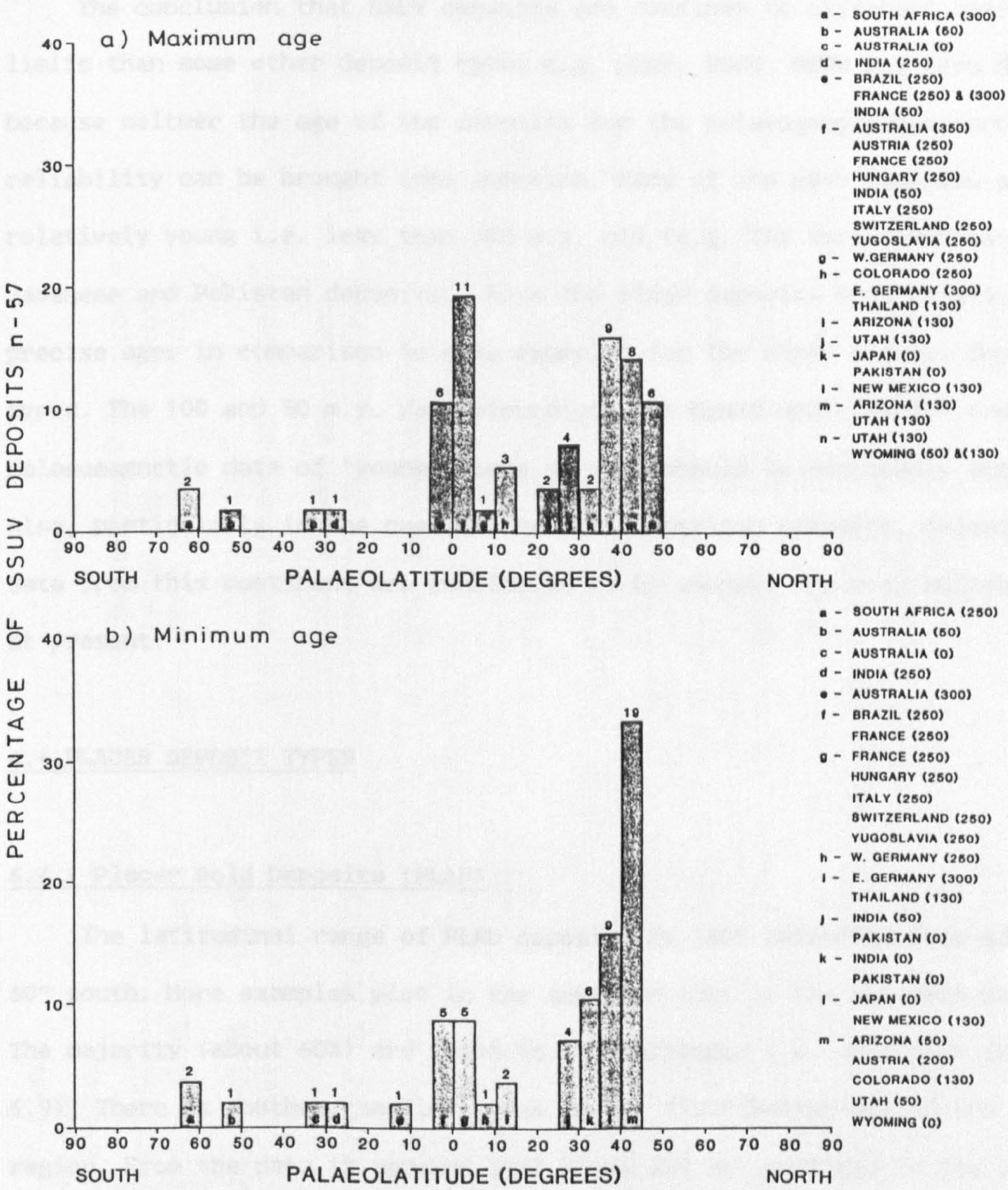


Figure 6.8(ii). Palaeolatitude vs. Frequency for SSUV deposits from Tarling reconstructions.
For figure legend see Figure 6.2(ii).

The conclusion that SSUV deposits are confined to different latitude limits than some other deposit types e.g. LSBM, SSCU, SSPB has been drawn because neither the age of the deposits nor the palaeogeographic reconstruction reliability can be brought into question. Many of the SSUV examples are relatively young i.e. less than 100 m.y. old (e.g. the North American, Indian, Japanese and Pakistan deposits). Also the older deposits have relatively precise ages in comparison to some examples for the other mineral deposits types. The 100 and 50 m.y. reconstructions are based upon results from palaeomagnetic data of 'young' rocks so they should be reasonably accurate. Also, particularly in the case of the North American deposits, palaeomagnetic data from this continent are considered to be amongst the most reliable in use at present.

6.4 PLACER DEPOSIT TYPES

6.4.1 Placer Gold Deposits (PLAU)

The latitudinal range of PLAU deposits is 125° extending from 65° north to 60° south. More examples plot in the southern than in the northern hemisphere. The majority (about 60%) are found in mid-latitudes i.e. 40 to 60° (Figure 6.9). There is another, smaller, peak in the distribution around the 60° to 65° region. From the data it appears that PLAUs are not confined to the low latitude region but are concentrated in the warm temperate rainfall belt. This is illustrated by both the maximum and minimum histograms (Figure 6.9) which show that only 18% of PLAU deposits lie between 30° north and south of the equator and 20% between 35° north/south.

These conclusions are drawn from present day latitudes rather than palaeolatitudes because of the very young age of most of the PLAU deposits so their reliability cannot be in question. For those deposits which must be rotated the BP reconstructions produce similar palaeolatitudes to those of

Figure 6.9: Placer Gold Deposits

Minimum Age Histogram			Maximum Age Histogram		
a -	Papua New Guinea	(0)	A -	Papua New Guinea	(0)
	Colombia	(0)		Colombia	(0)
b -	Guatemala	(0)	B -	Guatemala	(0)
	Honduras	(0)		Honduras	(0)
c -	Guatemala	(0)	C -	Guatemala	(0)
	Mexico	(0)		Mexico	(0)
d -	Mexico	(0)	D -	Mexico	(0)
e -	Mexico	(0)	E -	Mexico	(0)
f -	Mexico	(0)	F -	Australia	(0)
				California	(0)
				South Island, NZ	(0)
g -	Australia	(0)	G -	South Island, NZ	(0)
	California, USA	(0)			
	South Island, NZ	(0)			
h -	South Island, NZ	(0)	H -	South Island, NZ	(0,130)
i -	South Island, NZ	(0,50,130)	I -	Siberia, USSR	(0)
				South Island, NZ	(50)
j -	Siberia, USSR	(0)	J -	South Island, NZ	(100)
	South Island, NZ	(50)			
k -	South Island, NZ	(100)	K -	Yukon, Canada	(0)
				Alaska, USA	(0)
				Siberia, USSR	(0)
l -	Yukon, Canada	(0)			
	Alaska, USA	(0)			
	Siberia, USSR	(0)			

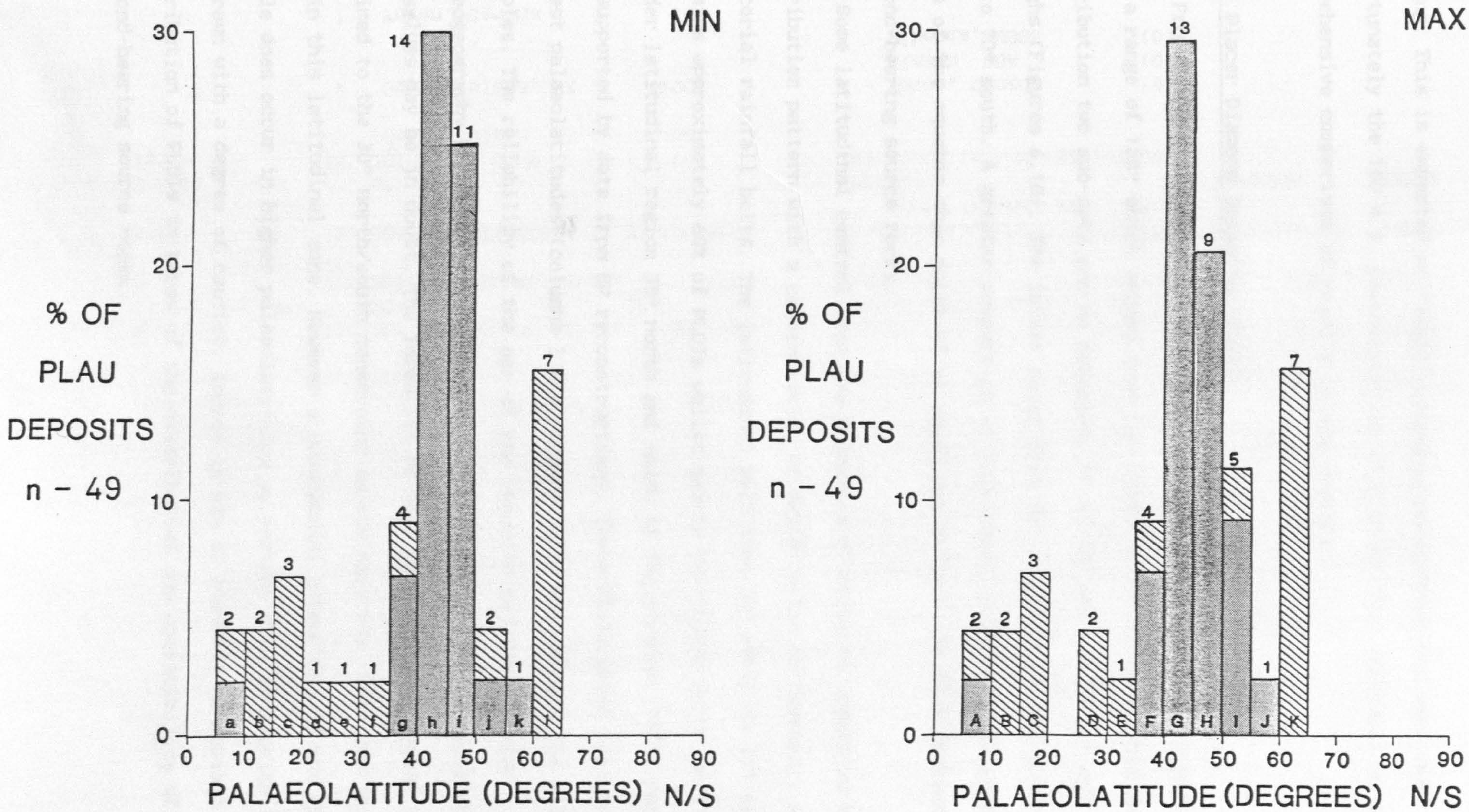


Figure 6.9. Palaeolatitude vs. Frequency for PLAU deposits from Tarling reconstructions.
For figure legend see Figure 6.2(i).

Tarling. This is expected as 'young' reliable reconstructions were used. Unfortunately the 100 m.y. reconstruction from BP was not available so a more comprehensive comparison of results is not possible.

6.4.2 Placer Diamond Deposits (PLDI)

From Tarling's reconstructions PLDIs have palaeolatitudes of formation with a range of 125° which extend from 50° north to 75° south. Within the distribution two sub-sets can be observed, 0° to 15° and 40° to 60° and two troughs (Figures 6.10). The latter occur from 35° to 40° north and south and 60° to 70° south. A greater proportion of PLDI examples have palaeolatitudes south of the equator than north of it which may reflect the distribution of diamond-bearing source rocks.

Some latitudinal control upon the formation of PLDIs is suggested by their distribution pattern with a concentration of deposits in the temperate and equatorial rainfall belts. The latitudinal belt from 30° north to 30° south contains approximately 60% of PLDIs whilst nearly two-thirds occur in the broader latitudinal region 35° north and south of the equator. These results are supported by data from BP reconstructions. Those PLDIs which plot in the highest palaeolatitudes (columns l & m, Figure 6.10) are amongst the oldest examples. The reliability of the age of the deposits and the proposed palaeogeography become questionable at these ages. Hence the palaeolatitudes themselves may be in doubt. The formation of PLDI deposits does appear to be confined to the 30° north/south constraint as the majority (60%) are found within this latitudinal zone. However a substantial proportion of the PLDI sample does occur in higher palaeolatitudes so perhaps this conclusion should be drawn with a degree of caution. Indeed it may be possible to explain the distribution of PLDIs in terms of the distribution and accessibility of diamond-bearing source rocks.

Figure 6.10: Placer Diamond Deposits

Minimum Age Histogram

a	Borneo	(130)
	Ghana	(130)
	Ivory Coast	(130)
	Liberia	(130)
	Venezuela	(100)
	West Africa	(130)
b	Brazil	(0,350)
	Central Af. Rep.	(50)
	Ivory Coast	(130)
	Liberia	(130)
c	Brazil	(0,350)
	Central Af. Rep.	(50)
	Mali	(130)
d	Brazil	(0,100)
e	Brazil	(0,100)
	Zaire	(50,130)
f	Brazil	(50,100)
g	USA	(130)
h	USA	(130)
i	South Africa	(130)
	USA	(0)
j	Australia	(450)
	Michigan, USA	(0)
k	South Africa	(50,100)
l	South Africa	(400)
m	Zimbabwe	(300)

Maximum Age Histogram

A	Borneo	(130)
	Ghana	(130)
	Ivory Coast	(130)
	Liberia	(130)
	West Africa	(130)
B	Brazil	(0,130,350)
	Central Af. Republic	(50)
	Ivory Coast	(130)
	Liberia	(130)
	Venezuela	(100)
C	Brazil	(0,350)
	Central Af. Republic	(50)
	Mali	(130)
D	Brazil	(0,130)
E	Brazil	(0)
	Zaire	(50,130)
F	Namibia	(0)
G	Brazil	(130)
	USA	(130)
H	USA	(130)
I	South Africa	(130)
	USA	(0)
J	Australia	(50)
	Michigan, USA	(0)
	South Africa	(100)
K	South Africa	(100)
L	Zimbabwe	(400)
M	Zimbabwe	(300)

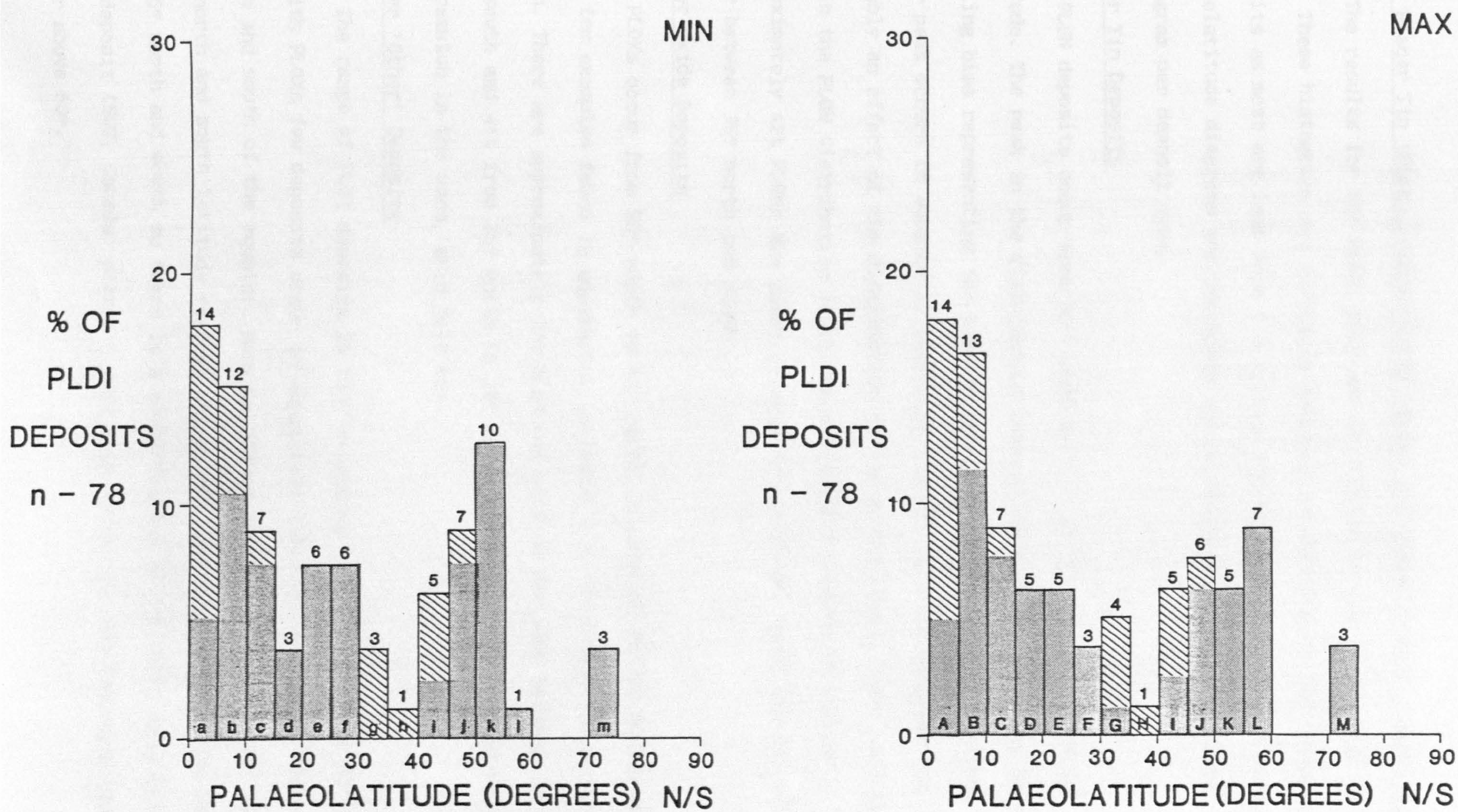


Figure 6.10. Palaeolatitude vs. Frequency for PLDI deposits from Tarling reconstructions.
For figure legend see Figure 6.2(i).

6.4.3 Placer Tin (PLSN), Placer Oxide (PLOX) and Placer 'Other' (PLOT) Deposits

The results for the PLSN, PLOX and PLOT deposits are illustrated in Figure 6.11. These histograms are virtually composed of the present latitudes of the deposits as most are less than 1 m.y. old. Maximum and minimum age-derived palaeolatitude diagrams are therefore not required so there is only one histogram per deposit type.

Placer Tin Deposits

PLSN deposits occur from 50° south to 30° north, a range of 80° of latitude. The peak in the distribution between 40° and 45° south may be a sampling bias representing the concentration of deposits in New Zealand. The other peak occurs in equatorial latitudes i.e. 0° to 5° north/south and is also probably an effect of the distribution of tin source rocks. There is a marked gap in the PLSN distribution in the warm arid belt between 5° and 25°. Approximately 42% PLSNs are found between 30° north and south and 46% PLSNs occur between 35° north and south.

Placer Oxide Deposits

PLOXs occur from 50° south to 40° north, a range of 90° of latitude with very few examples found in equatorial latitudes i.e. less than 15° north and south. There are approximately 30% PLOX deposits in the zone between 30° north and south and 41% from 35° north to 35° south of the equator suggesting a suppression in the warm, arid belt too.

Placer 'Other' Deposits

The range of PLOT deposits is 120° extending from 50° south to 70° north. As with PLOXs few deposits occur in equatorial latitudes i.e. less than 15° north and south of the equator. However 52% of the PLOT sample occurs in the 30° north and south latitude region, whilst 56% are found between the latitudes of 35° north and south so there is a concentration in the warm, arid belt. Only one deposit (NWT, Canada) plots in very high latitudes: the remainder do not occur above 50°.

Figure 6.11: Placer Tin Deposits

a - Bangka I., Indonesia (0)
    Billiton, Indonesia (0)
    Malaysia (0)

b - Belize, Mexico (0)
    Queensland, Australia (0)

c - Durango, Mexico (0)

d - NSW, Australia (0)

e - Victoria, Australia (0)

f - South Island, NZ (0)
    Tasmania (0)

g - Stewart Island, NZ (0)

Placer Oxide Deposits

a - Sri Lanka (0)

b - Colima, Mexico (0)
    Oaxaca, Mexico (0)
    Honduras (0)

c - Queensland, Australia (0)
    Natal, South Africa (0)
    Mexico (100)

d - NSW, Australia (0)
    W. Aust., Australia (0)

e - North Island, NZ (0)

f - South Island, NZ (0)
    New Jersey, USA (0)

g - South Island, NZ (0)
    Stewart Island, NZ (0)

Placer 'Other' Deposits

a - Colombia (0)
    Sri Lanka (0)

b - Mato Grosso, Brazil (0)
    Minas Gerais, Brazil (0)

c - Minas Gerais, Brazil (0)

d - Santa Catarina, Brazil (0)
    Sonora, Mexico (0)
    Natal, South Africa (0)

e - W. Aust., Australia (0)

f - North Island, NZ (0)

g - South Island, NZ (0)

h - South Island, NZ (0, 50)

i - NWT, Canada (0)

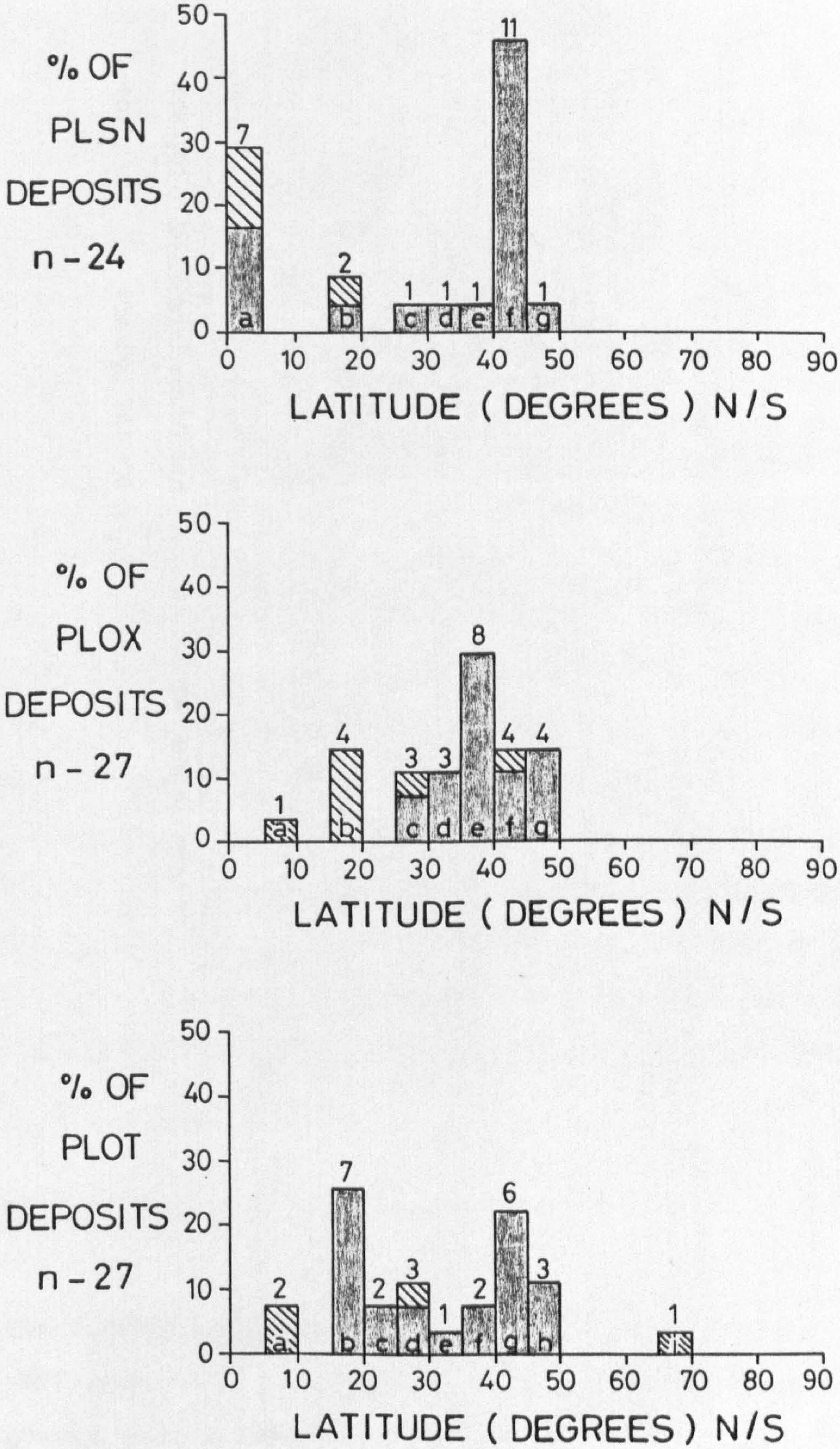


Figure 6.11. Palaeolatitude vs. Frequency for PLSN, PLOX and PLOT deposits from Tarling reconstructions. For figure legend see Figure 6.2(i).

To summarize, all three deposit types are confined to within 50° of the equator, except the one PLOT example in Canada previously noted. The largest proportion of each of the samples was collected from the southern hemisphere and the peaks in the distribution patterns may be due to the large collection of samples from New Zealand. However the distributions may also reflect the occurrence of source rocks. The results do not show a 30° latitude constraint upon the formation of these placer deposit types, although they do show a concentration in mid-latitudes i.e. 35 to 60°.

It is difficult to generalize about placer deposit distribution although they appear to be confined to low and mid-latitudes, few occurring above 60°. Placer deposits are uncommon in polar zones and the colder regions of the temperate rain belt. The number of PLAU, PLOT and PLOX deposits is suppressed in the equatorial rain belt i.e. 15° north and south of the equator, which may be a reflection of the greater chemical activity characteristic of this region. However the converse is true for the PLDI and PLSN deposits.

In conclusion the placer deposits do not obviously obey any specific latitudinal constraint upon their formation. The dominant factor in placer distribution may be the availability of the source rock which marks the influence of any other effect that may exist. For example the peak in the distributions around 40° is due to the concentration of these deposits in New Zealand.

6.5 'OTHER' DEPOSIT TYPES

6.5.1 Manganese Formations (MNFM)

The MNFM sample size is very small (n = 13) so the results must be interpreted with caution. The palaeolatitudes extend from 50° north to 50° south, a range of 100° (Figure 6.12). As with many other deposit types more examples plot north of the equator than south of it. The so-called 'peaks' in

Figure 6.12: Manganese Deposits

Minimum Age Histogram

a	China	(250)
b	Mexico	(0)
	Morocco	(100)
c	Mexico	(130, 200)
d	USSR	(50, 200)
e	Australia	(100)
	USSR	(50)

Maximum Age Histogram

A	Mexico	(250)
B	China	(300)
C	Mexico	(50, 200)
	Morocco	(130)
D	USSR	(50)
E	USSR	(50)
F	USSR	(200)
G	Australia	(100)

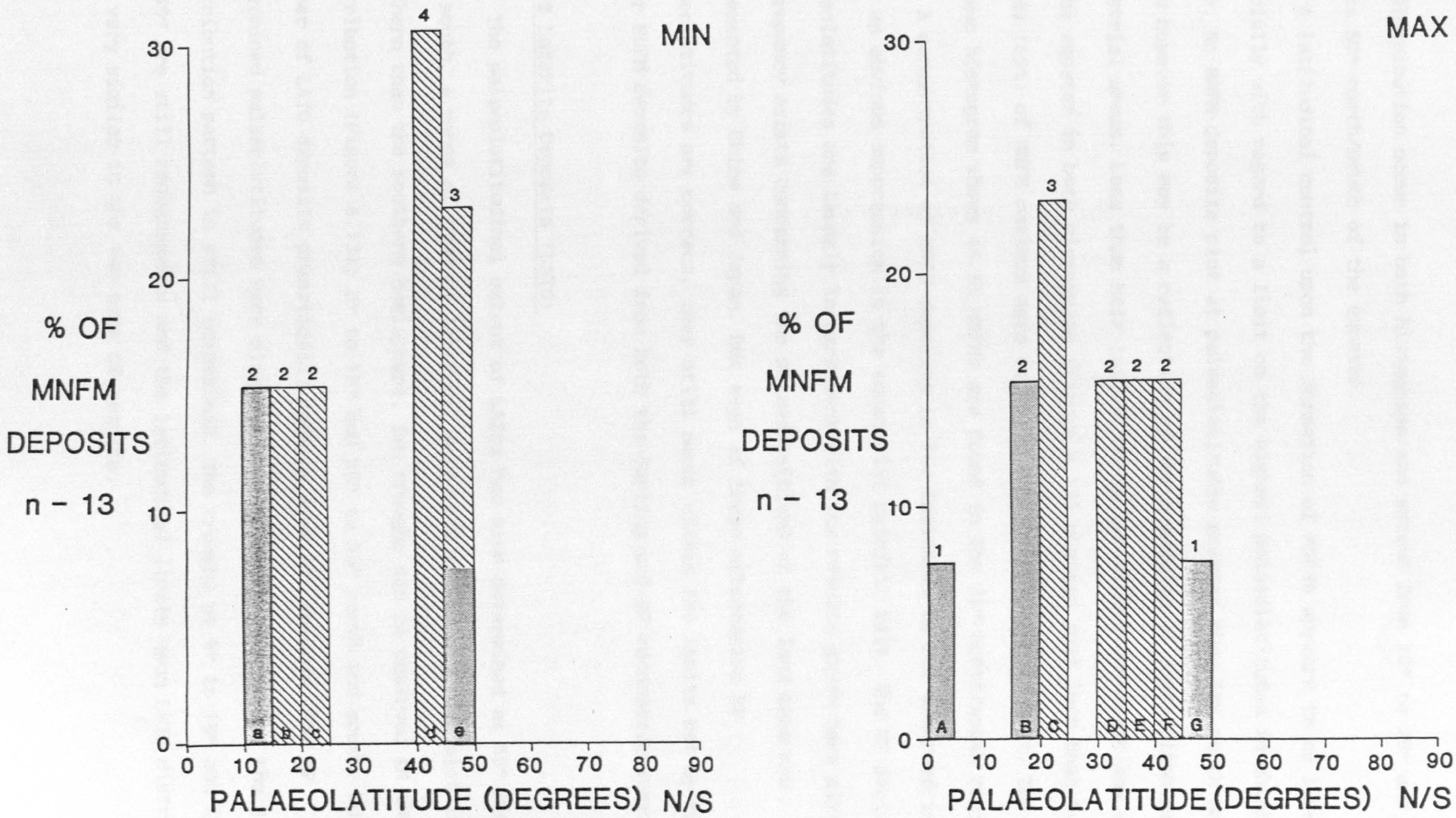


Figure 6.12. Palaeolatitude vs. Frequency for MNFM deposits from Tarling reconstructions.
For figure legend see Figure 6.2(i).

the distribution occur in both histograms and extend from 10° to 25° and from 30° to 50° north/south of the equator.

A latitudinal control upon the formation of MNFMs appears to be likely, especially with regard to a limit on the highest palaeolatitudes in which they occur. No MNFM deposits plot at palaeolatitudes greater than 50° north and south however this may be a reflection of the economics of exploration near industrial areas. Less than half (46.5%) MNFMs lie within 30° north and south of the equator in both histograms (Figure 6.12) whereas just less than two-thirds (62%) of MNFM maximum ages occur in the 35° north/south belt. The minimum histogram shows 46.5% MNFMs are found in the 35° north/south region.

A concentration of MNFM deposits in low latitudes is not observed in these with an obvious suppression in the equatorial rainfall belt. The BP derived palaeolatitudes are largely in agreement with the results given here although a discrepancy exists concerning the palaeopositions of the land mass now represented by China and Japan. But even if these alternative BP palaeolatitudes are correct, they still occur within the limits set by the other MNFM deposits derived from both the Tarling and BP reconstructions.

6.5.2 Laterite Deposits (LATO)

The palaeolatitudinal extent of LATOs has been determined as 55° north to 65° south, a range of 120° of latitude. More LATO palaeolatitudes plot in the northern than the southern hemisphere. Two troughs can be observed in the distribution (Figure 6.13); 5° to 15° and 25° to 35° north and south. The number of LATO deposits dramatically reduces from 45° onwards. The BP determined palaeolatitudes vary slightly from those of Tarling but the general distribution pattern is still maintained. The troughs at 5° to 15° and from 25° to 35° are still recognisable and the latitudinal limits upon LATO distribution are very similar in the two sets of results.

A palaeolatitude control on the distribution of LATOs is evident from the results. Approximately 73 to 76% LATOs occur within 35° north and south of the equator (minimum and maximum histograms respectively) and two-thirds of all LATOs lie within 30° of the equator irrespective of the age of the deposits from which the palaeolatitudes are derived. Many present day examples occur in higher-low to mid latitudes implying a wider latitudinal limit on their distribution than for other mineral deposit types. For example the deposits of Oregon, USA with a latitude of 50° at 50 m.y. and those of the Ukraine at 43° north 50 m.y. ago. However there is a reduction in the number of LATO deposits in the temperate rainfall belt.

Figure 6.13 (minimum histogram) shows one example which is at a much higher latitude than the others. The palaeolatitude for this Pakistan deposit has been determined from the rotation figures of India for 130 m.y. The BP rotation of the same age gave a palaeolatitude of 34° south which is more in accordance with the other observed results. Therefore the palaeoposition of the land mass on the Tarling reconstruction may be at fault.

The maximum histogram (Figure 6.13) also has an anomalous example - the deposit from Kandahar, Afghanistan (130 m.y. rotation). However when this deposit is rotated by 100 m.y. the palaeolatitude is 34° i.e. within low latitude limits. The deposit is dated as 144-97 m.y., so if the low latitude control is correct and is applied, the younger age for the deposit may be more appropriate. The BP determined palaeolatitude for this deposit at 130 m.y. is 13° north and for 100 m.y. is 18° north, both of which are vastly different from the latitudes determined from Tarling's reconstructions. This is one of the few occasions when a major discrepancy occurs in the palaeoposition of a region containing mineral deposit examples.

Figure .13: Laterite Deposits

Minimum Age Histogram

a - Aegean Islands (250)
Burundi (0)
India (50)
Indonesia (0)
Missouri, USA (300)
Pakistan (50)
Turkey (250)
Venezuela (0)

b - China (300)
Colombia (0)
Iran (200)
Ivory Coast (0)
Solomon Islands (0)

c - Phillipines (0)
Solomon Islands (0)

d - Brazil (0)
Cayman Islands (0)
China (300)
Dominion Republic (0)
Fiji (0)
Guatemala (0)
Haiti (0)
Jamaica (0)
Queensland, Aust (0)
Tonga Group (0)
Vietnam (250)

e - Brazil (0)
China (300)
Cuba (0)
Greece (100)
Iran (250)
Italy (100)

e - Kampuchea (250)
Loyalty Group (0)
Mexico (0)
New Caledonia (0)
Pakistan (250)
Sardinia (100)
Spain (100)
USSR (300, 450)
Vietnam (0)

f - Pyrenees (100)
USSR (100, 350)
Yugoslavia (100)

g - Afghanistan (100)
Albania (50)
Austria (100)
Greece (130)
USSR (100)

h - Albania (200)
France (130)
Greece (0, 200)
Pakistan (350)
Rumania (130)
Spain (50)
USSR (100, 130)
Yugoslavia (130)

i - Oregon, USA (0)
USSR (50, 100)
Yugoslavia (0)

j - USSR (0, 50, 100)

k - USSR (0, 100)

l - Pakistan (130)

Maximum Age Histogram

A - Aegean Islands (250)
Albanian Alps (250)
Colombia (50)
Greece (250)
India (50)
Ivory Coast (50)
Pakistan (50)
Turkey (250)
Venezuela (50)

B - China (350)
Solomon Islands (0)
USSR (350)

C - Indonesia (50)
Missouri (350)
Pakistan (400)
Phillipines (0)
Solomon Islands (0)

D - Burundi (50)
Cayman Islands (0)
China (250, 350)
Cuba (50)
Dominion Republic (0)
Fiji (0)
Guatemala (50)
Haiti (0)
Jamaica (0)
Tonga Group (0)
Vietnam (250)

E - Brazil (50)
Greece (100)
Iran (250)
Italy (100)
Kampuchea (250)
Loyalty Group (0)

E - Mexico (50)
Pakistan (250)
Sardinia (100)
Spain (100)
USSR (450)
Vietnam (0)

F - Brazil (50)
Greece (50)
Iran (250)
Mexico (50)
Pyrenees (100)
USSR (350)

G - Albania (50)
Austria (100)
Greece (130)
Pakistan (200)
USSR (100)
Yugoslavia (50)

H - France (200)
Greece (200)
Rumania (130)
Spain (50)
USSR (100, 130)
Yugoslavia (130)

I - Queensland, Aust (50)
USSR (50, 100)

J - USSR (50, 100)

K - New Caledonia (50)
Oregon, USA (50)
USSR (100)

L - Afghanistan (130)

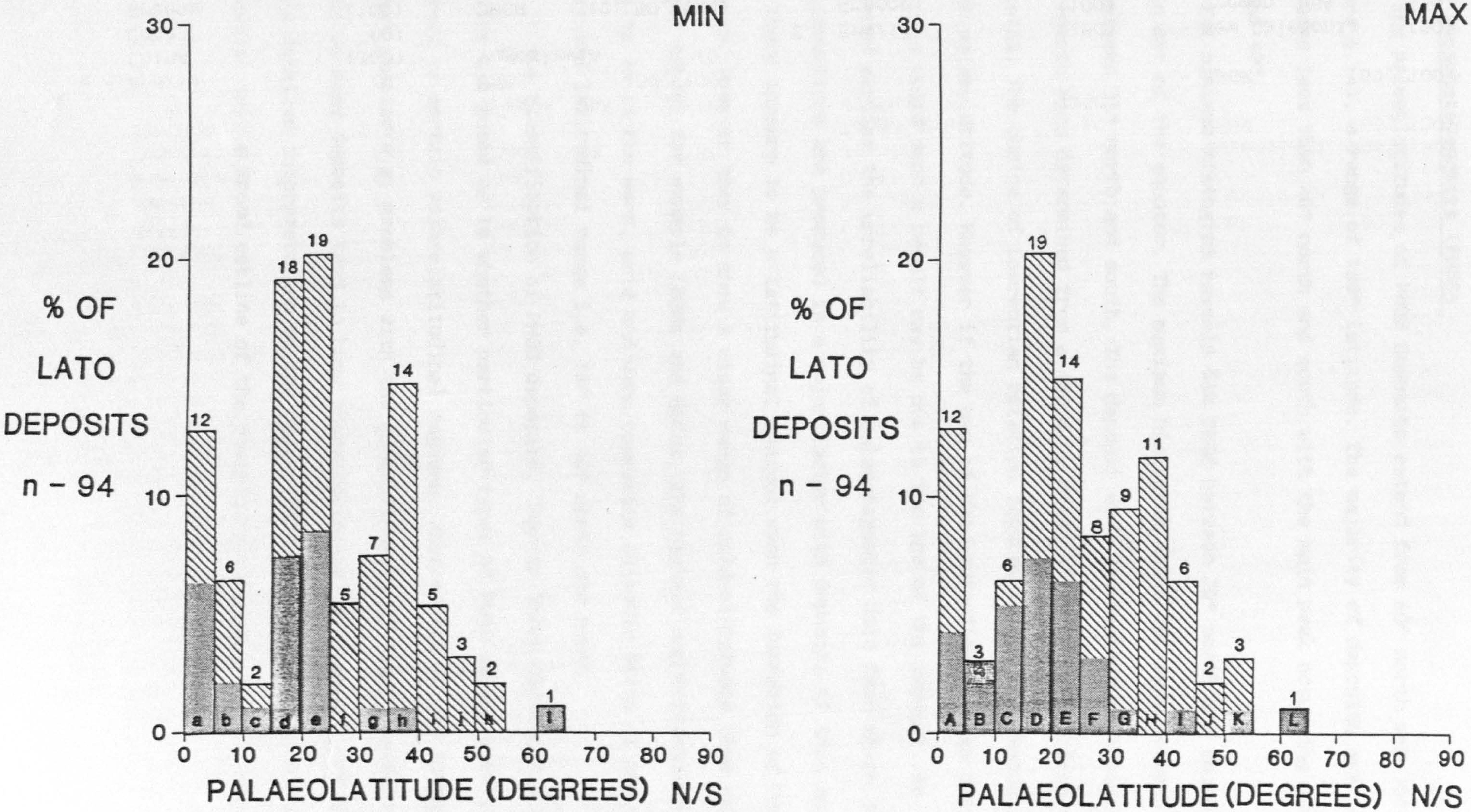


Figure 6.13. Palaeolatitude vs. Frequency for LATO deposits from Tarling reconstructions. For figure legend see Figure 6.2(i).

6.5.3 PhosphateDeposits (PHOS)

The palaeolatitudes of PHOS deposits extend from 60° north and 70° south (Figure 6.14), a range of 130° latitude. The majority of deposits are found in latitudes less than 40° north and south with the main peak occurring between 25° and 40°.

The minimum histogram reveals 56% PHOS between 30° north and south and 70% within 35° of the equator. The maximum histogram shows 58% PHOS between 30° and 74% between 35° north and south. The deposit with the highest palaeolatitude is from Puerto Rico determined from a 450 m.y. rotation with figures from Laurentia. The choice of Laurentian rotation figures may be the reason for such a high palaeolatitude. However if the use of the Laurentian figures for this region is valid such a result may be due to the age of the deposit. As mentioned earlier the unreliability of palaeomagnetic data from which the reconstructions are produced is a major factor with deposits of this age.

There appears to be a latitudinal control upon the formation of PHOS deposits. However they do show a wider range of palaeolatitudes than other deposit types, for example LSBMs and SSCUs. The highest concentration of PHOS deposits is in the warm, arid and warm temperate climatic belts of the higher-low to mid latitudinal range i.e. 15° to 40° north and south.

In the classification of PHOS deposits, Chapter Three section 3.2.5, the question was posed as to whether particular types of PHOS deposits were confined to certain palaeolatitudinal regions. However due to the reasons given in that chapter e.g. problems with the identification of these separate PHOS groups as many deposits tend to have characteristics of more than one group, such a detailed interpretation of the palaeolatitudinal distribution has proved difficult. Only a broad outline of the distribution has been given here.

Figure 6.14: Phosphate Deposits

Minimum Age Histogram

a - China (250, 450)
Norway (350)
USA (300)
USSR (350)

b - Australia (250)
Finland (450)
Peru (0)
USSR (400)

c - Curacao (0)
Iraq (50)
Pakistan (400)

d - Iraq (50)
Peru (0)
Sweden (450)
USA (250, 300, 350)
USSR (400, 450)

e - Australia (250)
Canada (300)
China (400)
England (450)
Mexico (0)
Pakistan (250)
USA (250)
USSR (450)
Vietnam (300)

f - Australia (450)
Florida, USA (0)
Kampuchea (300)
Mexico (0, 100)
Spain (350)
USSR (350, 450)
Wales (450)

g - Australia (450)
Florida, USA (0)
Georgia, USA (0)
Turkey (350)
S. Africa (0)
USSR (250, 400)

h - Australia (250)
Canada (250, 300)
Denmark (100)
England (100, 400)
Japan (0)
N. Carolina, USA (0)
Pakistan (350)
USA (400, 450)
USSR (400)

i - England (50)
USA (450)

j - Australia (450)
Canada (450)
Iran (350)
USA (450)

k - England (0)
Iran (350)

l - USSR (400)

m - USA (450)

Maximum Age Histogram

A - China (450)
Mexico (250)
USA (300)

B - Australia (250)
China (300)
Finland (450)
Peru (0)
USSR (400, 450)

C - China (450)
Curacao (0)
Iraq (50)
Norway (400)
Pakistan (400)
USA (350)

D - Canada (350)
Iraq (50)
Peru (0)
Sweden (450)
USA (250, 350)
USSR (450)

E - Australia (250)
Canada (300)
England (450)
Turkey (400)
USA (250)
USSR (450)
Vietnam (300)

F - Australia (300, 350)
Canada (300)
Florida, USA (0)
Kampuchea (300)
Mexico (0)
USSR (300,350,400,450)
Wales (450)

G - Australia (450)
Canada (300)
Florida, USA (0)
Georgia, USA (0)
Mexico (50)
Pakistan (300)
USSR (400, 450)

H - Denmark (100)
England (100)
Japan (0)
N. Carolina, USA (0)
Pakistan (350)
S.Africa (0)
USA (400, 450)
USSR (400)

I - Spain (400)
USA (450)

J - Australia (450)
Canada (450)
Iran (350)
USA (450)

K - England (0)
Iran (350)

L - USA (450)

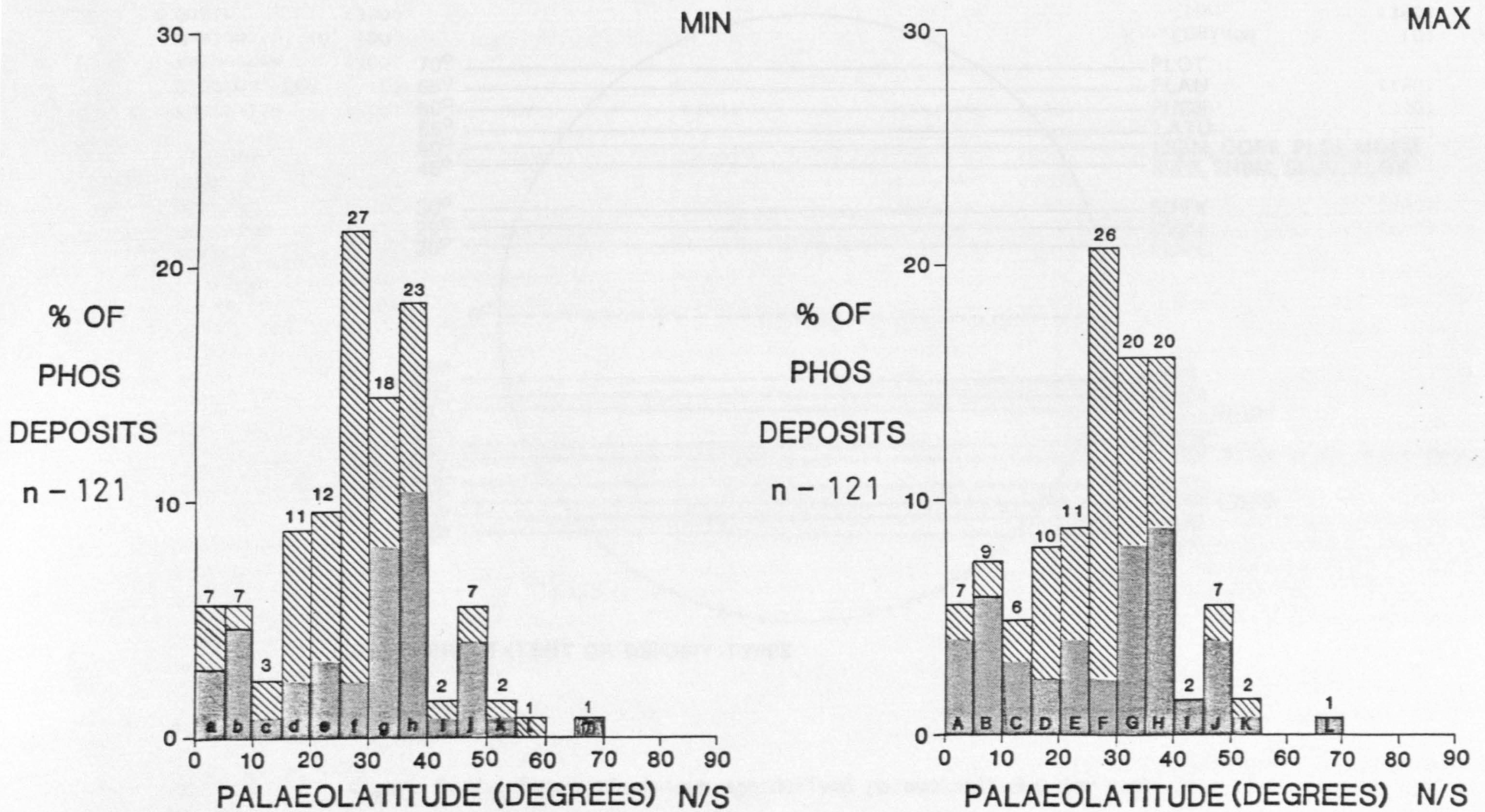


Figure 6.14. Palaeolatitude vs. Frequency for PHOS deposits from Tarling reconstructions.
For figure legend see Figure 6.2(i).

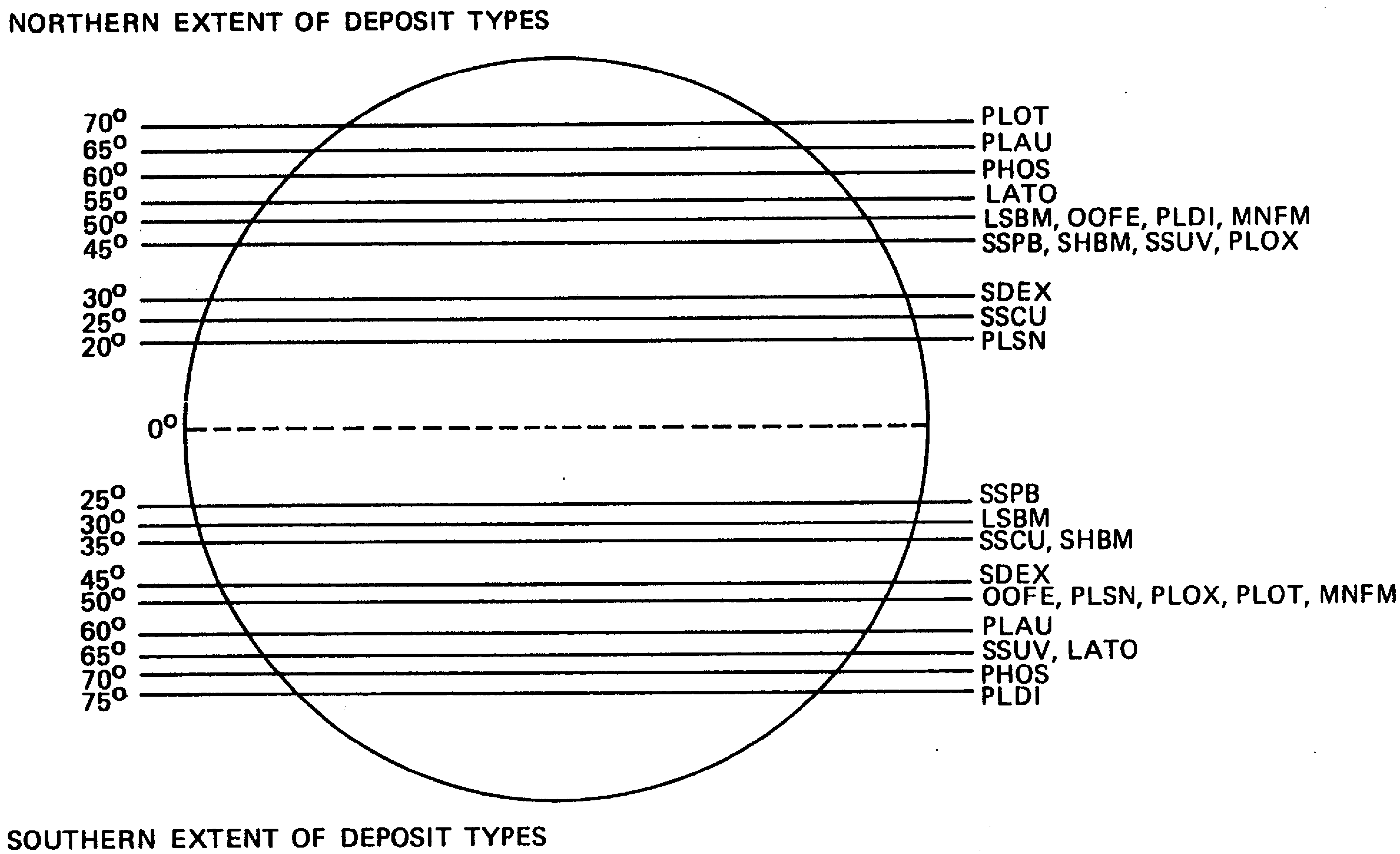


Figure 6.17a. Range of minimum age-derived palaeolatitudes for each Phanerozoic mineral deposit type.

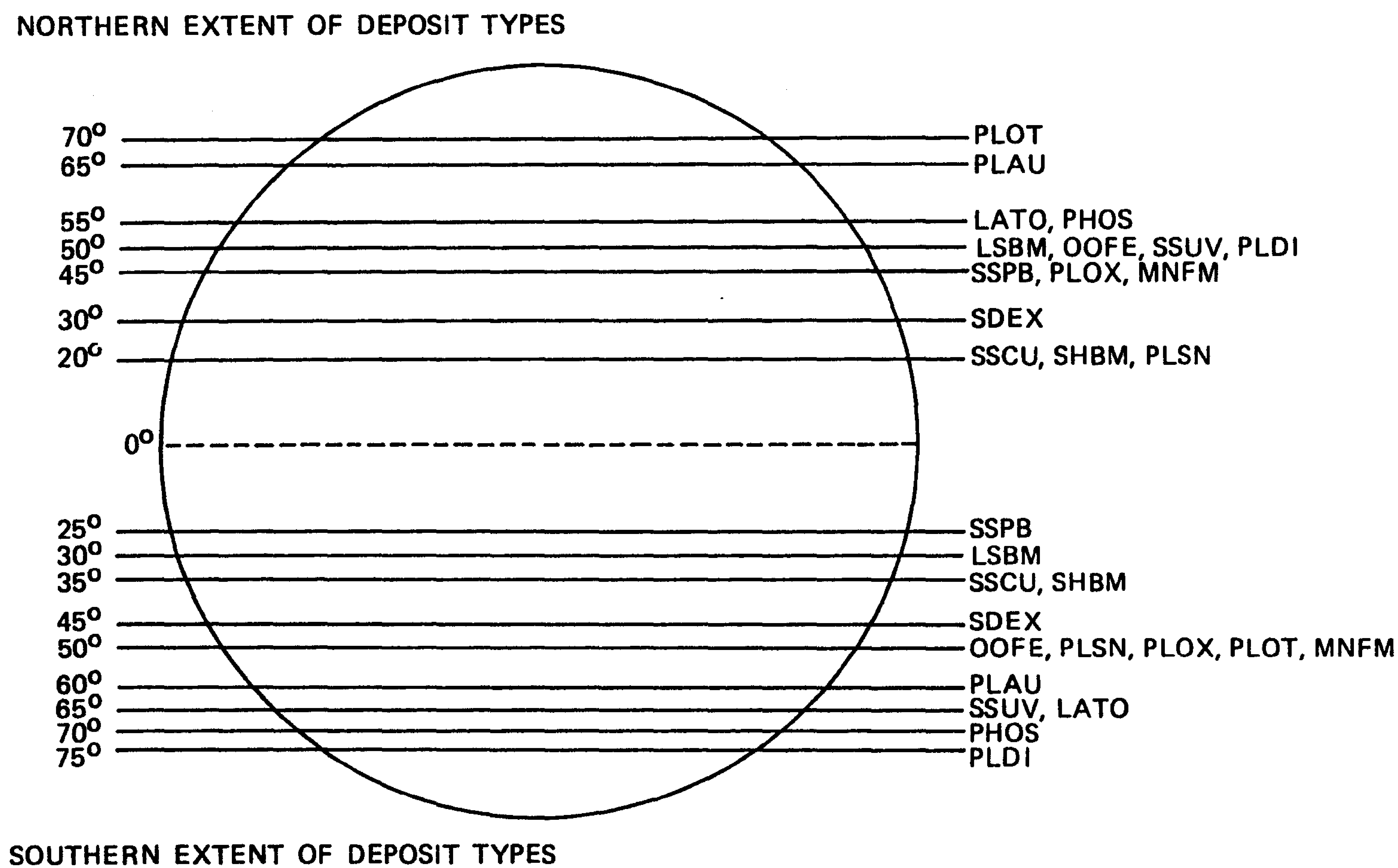


Figure 6.17b. Range of maximum age-derived palaeolatitudes for each Phanerozoic mineral deposit type.

6.6 STATISTICS

The data are too sparse for complicated frequency tests to be applied to individual deposit types. Hence the statistical tests have been kept to a minimum.

6.6.1 The Arithmetic Mean, Median and Mode.

The mean has been calculated for each mineral deposit type (Table 6.1) as the representative of a homogeneous group of which the members are recognisably similar. Equal weight was given to each latitude occurrence. The values for the mean vary from 23° south (PLSN) to 22° north (SSUV, minimum age). Those deposits with a mean value in the southern hemisphere include SSCU, SDEX and all the Placer deposits. The deposit types whose mean value is found north of the equator are LSBM, OOFE, SSPB, SHBM, SSUV, MNFM, LATO and PHOS. Two mineral deposit types have a mean value very near to the equator i.e. SSCU and PHOS suggesting an even number of deposits in each hemisphere after rotation.

Table 6.1 also shows the modal values for each deposit type. A mode is defined as the most commonly occurring value and is thought to be "typical" of the distribution in question. This is a method of determining a measure of the central tendency which is not upset by extreme values in the distribution. The modal values vary from the equatorial region (0-5°) for LSBM (min), SSCU (min), SSUV (max) and PLDI (min and max) to upper middle latitudes (50-55°) for PLDI (min) deposits. The modes for each deposit type are similar to their median and mean values as would be expected. Some deposit types e.g. OOFE (min) have three modal values reflecting their concentration in both equatorial and temperate rainfall belts. The PLDI deposits have two modal values which are very different (i.e. 0-5° and 50-55°) which may be due to their requirement for specific diamond source rocks. Both the arithmetic mean and modal values are valid. However the mode would form a very poor basis for any calculations of an arithmetical nature (e.g. standard deviation) as it has excluded arithmetical precision in the interests of presenting a typical result. In contrast the arithmetic mean may be untypical of the distributions in order to be numerically accurate (Moroney, 1951).

Many distributions are such that very great differences exist between the largest and smallest members. They also exhibit a marked lack of symmetry, the samples tending to cluster nearer to one extreme than the other. To calculate the averages for distributions of this type using the arithmetic mean would be misleading. For example, in this case with the data arranged north and south if the proportion of deposits in higher palaeolatitudes was high enough to affect the statistics, then the average palaeolatitude would increase appreciably. Such an average could not be taken as truly representative of the population in general. For examples such as the one described it is evident that a measure of central tendency is needed which is unaffected by the relatively few extreme values in the 'tail' of the distribution. This measure is known as the median and is the value in the central position of the distribution (Table 6.1). The median values for all the mineral deposit types used in this thesis vary from 41° south (PLAU) to 35° north (SSUV, minimum ages). Those deposit types with a median value in the southern hemisphere are SSCU, SDEX and all the Placer deposits. Those with a central position of the distribution in the northern hemisphere are LSBM, OOFE, SSPB, SSUV, MNFM, LATO and PHOS. This division of the results into the northern and southern hemispheres is the same as that for the mean values. Hence the palaeolatitude distribution of the mineral deposit types must not be concentrated in one extreme or the other i.e. towards high or very low latitudes.

The median is the latitude at which the land area to the north and south of it are equal only if the rate of mineral deposition is proportional to the surface area. If these factors are proportional then the symmetry of distribution of deposits is only important if the distribution of continental mass is even in both hemispheres. Hence the equator need not necessarily be the median of the distribution.

6.6.2 Skewness in the Distributions

When plotted on a histogram some distributions are symmetrical about their central value while other distributions are skewed i.e. show marked asymmetry. Such distributions are divided into two groups, as described below with particular reference to this research.

a) If the 'tail' of the distribution reaches out into the higher values of the variate (i.e. palaeolatitude) the distribution is said to show positive skewness e.g. LSBM (min and max), SHBM (min and max) and PLOX.

b) If the `tail' extends towards the lower values (i.e. lower palaeolatitudes) it is said to be negatively skewed e.g. OOFE (min and max), SSCU (min and max), SSPB, SDEX (min), PLAU, PLOT, LATO and PHOS.

The mineral deposit types whose distributions cannot be classed as either positively or negatively skewed include; SSCU (max), SDEX (max), SSUV (min and max), PLSN, PLDI (min and max) and MNFM (min and max).

Skewness can only be observed in those histograms which have been compiled from a combination of results from both hemispheres.

6.6.3 Standard Deviation and the Coefficient of Variation.

The Standard Deviation (s) has been calculated for each mineral deposit type (Table 6.1). However the results may be misleading as this test is most suitable for distributions which are unimodal i.e. those with one peak and a specific symmetry in the distribution. The formula for the standard deviation is;

$$s = \frac{\sqrt{\Sigma (x - \bar{x})^2}}{n}$$

where s is the standard deviation, n is the number of samples, $\bar{x}$ is the arithmetic mean and x are the individual values. The standard deviation values calculated for the majority of deposit types were very similar i.e. for LSBM, SSPB, SDEX, SSUV, PLDI, PLSN, PLOX, PLOT, MNFM, LATO and PHOS (max). The SSCU and SHBM deposits have 's' values lower than that of the group above - indeed SHBMs have the lowest 's' value of all the deposit types. OOFEs and PHOS (max) deposits have an 's' value of similar size, which is above that of the general group. The PLAUs have the highest variability in terms of distribution of palaeolatitudes.

Table 6.1: Mean, Median, Mode And Standard Deviation for each Deposit Type.

MIN DEPOSIT TYPE		NO. OF EXAMPLES	MEAN (degrees)	MEDIAN (degrees)	MODE (degrees)	STANDARD DEVIATION (degrees)	COEFFICIENT OF VARIATION %
LSBM	MIN	39	14.3	4.3	0-5	21.6	20.7
	MAX	39	9.5	5.4	+/-5-10	19.5	19.6
OOFE	MIN	26	9.3	7.7	-35-40 & 40-50	33.7	33.9
	MAX	26	9.8	7.7	45-50	34.1	34.2
SSCU	MIN	28	-0.6	-3.6	-0-5	18.6	20.8
	MAX	28	-6.9	-21.0	-20-25	18.2	21.9
SSPB	MIN & MAX	10	14.4	13.0	40-45	23.6	22.6
SHBM	MIN	18	9.2	8.8	5-10	11.2	11.3
	MAX	18	6.9	8.5	5-10	10.0	10.3
SDEX	MIN	54	-8.9	-5.2	-20-25	23.4	23.7
	MAX	54	-11.0	-21.0	-20-25	22.2	22.0
SSUV	MIN	57	22.2	35.0	40-45	27.7	24.7
	MAX	57	17.9	29.9	0-5	27.3	25.3
PLAU	MIN	49	-12.6	-41.2	-40-45	42.2	54.4
	MAX	49	-13.2	-41.2	-40-45	42.9	55.9
PLDI	MIN	78	-16.4	-11.5	0-5 & -50-55	28.6	38.9
	MAX	78	-16.6	-11.0	0-5	28.8	39.2
PLSN	MIN & MAX	24	-23.3	-39.0	-40-45	23.2	34.8
PLOX	MIN & MAX	27	-22.8	-37.2	-35-40	27.0	40.2
PLOT	MIN & MAX	27	-22.8	-28.5	-15-20	26.1	38.8
MNFM	MIN	13	21.2	23.5	40-45	28.1	25.3
	MAX	13	16.5	24.1	20-25	27.2	25.5
LATO	MIN	94	13.1	18.8	35-40	25.1	24.4
	MAX	94	10.4	18.1	15-20	26.6	26.5
PHOS	MIN	121	3.2	15.8	25-30	30.6	32.8
	MAX	121	1.5	9.4	25-30	29.9	32.7

N.B. A negative sign denotes a latitude in the southern hemisphere. All results are based on palaeolatitudes arranged both north and south of the equator i.e. 90°N to 90°S.

To test whether the distribution of palaeolatitudes of one mineral deposit type is relatively more variable than that of another, Pearson's Coefficient of Variation (v) has been calculated for each deposit category (Table 6.1). The test is defined as;

$$v = \frac{100 \, . \, s}{\bar{x}}$$

where v is the coefficient of variation and the other symbols are as for the previous equation. Expressed as a percentage of the mean palaeolatitudes, the PLAU and PLOX deposits show the greatest variability. The PLDI, PLSN, PLOT, OOFE and PHOS groups have similar coefficients of variation which are below that of the other two. Those deposit groups with the lowest variability include LSBM, SSCU, SSPB, SHBM, SDEX, SSUV, MNFM and LATO. As with the standard deviation values the PLAU deposits are the most variable. However, in general, the carbonate- and clastic-hosted mineral deposit groups are the least variable in terms of the percentage of mean palaeolatitudinal occurrence.

6.7 Relative Distributions of Land Mass and Mineral Deposits

The results for each deposit type concerning the percentages of deposits in certain latitudes at specific periods in the Earth's history can only be interpreted correctly if the proportion of land mass in those latitudes at that time is known. If the percentage of mineral deposit occurrence is the same as the percentage of land mass it is not possible to demonstrate whether any palaeolatitude control upon the distribution of the deposits exists e.g. the occurrence of 60% land mass and 60% mineral deposits within 30° north and south of the equator for a specific time precludes the determination of any palaeolatitude influence that may exist. Table 6.2 was produced in an effort to solve this dilemma. Briefly the results of this investigation using Tarling's reconstructions are as follows.

Table 6.2: % Continental Mass: % Mineral Deposits for each Rotation

		NORTHERN HEMISPHERE		SOUTHERN HEMISPHERE	
PERIOD NAME	ROTATION (M.Y.)	% CONT AREA	% MIN DEPS	% CONT AREA	% MIN DEPS
PRESENT	0	65.7		34.3	
	MIN AGES		47.2		52.8
	MAX AGES		41.9		58.1
EOCENE	50	60.4		39.6	
	MIN AGES		53.1		46.9
	MAX AGES		57.3		42.7
CRET	100	53.6		46.4	
	MIN AGES		68.3		31.7
	MAX AGES		69.8		30.2
U.JUR	130	61.4		38.6	
	MIN AGES		70.8		29.2
	MAX AGES		68.2		31.8
L.JUR	200	58.9		41.1	
	MIN AGES		100.0		0.0
	MAX AGES		96.4		3.6
PERMO - TRIASS	250	43.8		56.2	
	MIN AGES		66.2		33.8
	MAX AGES		74.3		25.7
U.CARB	300	34.6		65.4	
	MIN AGES		41.7		58.3
	MAX AGES		37.1		62.9
L.CARB	350	22.0		78.0	
	MIN AGES		34.4		65.6
	MAX AGES		21.6		78.4
L.DEV	400	16.7		83.3	
	MIN AGES		25.9		74.1
	MAX AGES		18.2		81.8
SIL & ORD	450	18.1		81.9	
	MIN AGES		38.3		61.7
	MAX AGES		37.5		62.5

a) <u>Table 6.2, Northern Hemisphere, 0 - 450 m.y.</u>

i) Minimum Age-Derived Palaeolatitudes. These show the percentage of mineral deposits is greater than the percentage of land mass in the northern hemisphere from the Cretaceous (100 m.y.) to the Ordovician/Silurian period (450 m.y.). The exception to this trend is at present when the percentage of land mass is greater than that of mineral deposits. This phenomenon may be caused because at present only syngenetic deposits are seen. The secondary (epigenetic) deposits that will form in rocks of the present age have yet to do so and therefore cannot be considered. Hence the apparent lack of deposits relative to land mass. The proportions of both these criteria for the 50 m.y. and 130 m.y. periods are very similar.

ii) Maximum Age-Derived Palaeolatitudes. The percentages of continental mass and mineral deposit occurrence are very similar for the Eocene (50 m.y.), Upper Jurassic (130 m.y.) and Upper Carboniferous to Lower Devonian (300-400 m.y.) periods. There is a greater proportion of land in the northern hemisphere than the percentage of mineral deposits at the present time but the converse is the case for the Cretaceous, Lower Jurassic (200 m.y.), Permo-Triassic (250 m.y.) and Ordovician/Silurian periods.

b) <u>Table 6.2, Southern Hemisphere, 0 - 450 m.y.</u>

i) Minimum Age-Derived Palaeolatitudes. The proportion of surface area is greater than the rate of mineral deposit formation for the Cretaceous, Upper Jurassic, Permo-Triassic, Lower Carboniferous (350 m.y.) and the Ordovician/Silurian periods. The reverse is true from the present to 50 m.y. ago and very similar proportions of surface area and mineral deposit occurrence result for the Upper Carboniferous (300 m.y.) and Lower Devonian (400 m.y.). No palaeolatitudes were determined in the southern hemisphere for mineral deposits of Lower Jurassic age.

ii) Maximum Age-Derived Palaeolatitudes. The only period when the percentage of mineral deposits is greater than the continental mass in the southern hemisphere is at present. Very similar percentages resulted for the Eocene and the Upper Carboniferous to Lower Devonian periods. From the Cretaceous to the Permo-Triassic and during the Ordovician/Silurian periods the proportion of land mass is greater than that for mineral deposits.

To sum up, when the Earth is divided into the northern and southern hemispheres the percentages of mineral deposits and of continental mass are not identical. In general they do follow similar trends i.e. if there is a greater surface area in the southern hemisphere then the majority of mineral deposits occur in the same hemisphere e.g. during the Upper Carboniferous. The exceptions are at the present and during the Permo-Triassic period.

Up until now this research has been centred upon the division of the Earth into three zones, each of 60° of latitude. To be consistent the relative proportions of continental mass and mineral deposits have been calculated for these latitude zones so a more detailed comparison of their relative distributions is possible (as shown in Table 6.3) than from Table 6.2.

a) Table 6.3, >30°N, Section A, 0-450 m.y.

i) Minimum Age-Derived Palaeolatitudes. There are three periods when the percentage of land mass is greater than mineral deposit occurrences (the Present, Permo-Triassic, Upper Carboniferous) and two periods (Lower Jurassic and Lower Devonian) when the reverse is true. The surface area and rate of mineral deposit formation are proportional from the Eocene to the Upper Jurassic period (50-130 m.y.). No deposits of Lower Carboniferous age plot in this area and no continents occur in this region on the Ordovician/Silurian reconstruction.

ii) Maximum Age-Derived Palaeolatitudes. There are three periods when the percentage of land mass is greater than the percentage of mineral deposit occurrences (the Present, Eocene and Upper Carboniferous) and two when the reverse is true (Upper and Lower Jurassic). An agreement between the surface

Table 6.3: % Continental Mass: % Mineral Deposits for each Rotation

		NCLE DATA >30° N (SECTION A)		NCLE DATA 30°-30° (SECTION B)		NCLE DATA >30° S (SECTION C)	
PERIOD NAME	ROTATION (M.Y.)	% CONT AREA	% MIN DEPOSITS	% CONT AREA	% MIN DEPOSITS	% CONT AREA	% MIN DEPOSITS
PRESENT	0	53.6		19.9		26.4	
	MIN AGES		21.6		45.6		32.8
	MAX AGES		16.2		44.9		38.9
EOCENE	50	39.4		40.4		20.2	
	MIN AGES		34.7		44.9		20.4
	MAX AGES		25.3		61.3		13.3
CRET	100	31.5		43.5		25.0	
	MIN AGES		30.2		52.4		17.5
	MAX AGES		32.1		43.4		24.5
U.JUR	130	35.5		39.5		25.0	
	MIN AGES		41.7		45.8		12.5
	MAX AGES		51.5		36.4		12.1
L.JUR	200	27.0		47.2		25.8	
	MIN AGES		66.7		33.3		0.0
	MAX AGES		67.9		28.6		3.6
PERMO -	250	18.6		45.0		36.4	
TRIASS	MIN AGES		3.1		92.2		4.6
	MAX AGES		0.0		100.0		0.0
U.CARB	300	10.8		36.4		52.8	
	MIN AGES		2.1		91.7		6.3
	MAX AGES		2.9		80.0		17.1
L.CARB	350	6.3		39.4		54.3	
	MIN AGES		0.0		71.9		28.1
	MAX AGES		0.0		86.3		13.7
L.DEV	400	6.4		45.0		48.6	
	MIN AGES		11.1		29.6		59.3
	MAX AGES		6.0		33.3		60.6
SIL &	450	0.0		52.1		47.9	
ORD	MIN AGES		6.4		59.6		34.0
	MAX AGES		7.1		64.3		28.6

area and rate of mineral deposit formation exists for the Cretaceous and Lower Devonian periods. No deposits of Permo-Triass or Lower Carboniferous age plot in this area and no land occurs on the Ordovician/Silurian reconstruction above 30° north.

b) Table 6.3, 30°N - 30°S, Section B, 0-450 m.y.

i) Minimum Age-Derived Palaeolatitudes. During the Lower Jurassic and Lower Devonian periods the percentage of land mass is greater than that of mineral deposits. However at present, in the Cretaceous, Permo-Triassic, Lower and Upper Carboniferous periods the converse is the case. The two factors are virtually proportional for the Eocene, Upper Jurassic and Ordovician/Silurian periods.

ii) Maximum Age-Derived Palaeolatitudes. These results are similar to those for the minimum age-derived palaeolatitudes. But the proportions of land mass and mineral deposits are only similar for the Cretaceous and Upper Jurassic periods. Another difference is that there is a greater percentage of mineral deposits than land mass for the Ordovician/Silurian period. Also during the Permo-Triassic all the mineral deposits were formed in this area.

c) Table 6.3, >30°S, Section C, 0-450 m.y.

i) Minimum Age-Derived Palaeolatitudes. The percentage of land mass is greater than that of mineral deposits from the Cretaceous to the Lower Carboniferous and in the Ordovician/Silurian period. There were no deposits of Lower Jurassic age which plotted in this area. The percentage of mineral deposits is greater than that of land mass only in the Lower Devonian period and at present. The relative proportions of land mass and mineral deposits are remarkably similar for the Eocene period.

ii) Maximum Age-Derived Palaeolatitudes. The percentage of land mass is greater in the Eocene, from the Upper Jurassic to the Lower Carboniferous and during the Ordovician/Silurian period. However the reverse is true for the Lower Devonian and at present. No mineral deposit examples of Permo-Triassic age plot in this area and the distribution of land and rate of mineral deposit formation are proportional for the Cretaceous period.

It has been mentioned that the errors in the determination of the palaeolatitudes probably give an average error bar of 5° upon the values. So the relative distributions of land and mineral deposits have been calculated for the broader latitude zone from 35° north to 35° south (Table 6.4) to see if this division influences the results shown in Table 8.

a) Table 6.4, >35°N, Section A, 0-450 m.y.

i) Minimum Age-Derived Palaeolatitudes. The percentage of land mass is greater than that of mineral deposits at present, in the Cretaceous, Permo-Triassic and Upper Carboniferous periods. The percentage of mineral deposits is about twice that of the land mass for the Lower Jurassic and Lower Devonian periods. The relative distributions of land mass and mineral deposits are virtually the same for the Eocene and Upper Jurassic. No deposits of Lower Carboniferous age plot in the area above 35° north and the Ordovician/Silurian map has no land mass in this region.

ii) Maximum Age-Derived Palaeolatitudes. The general trend shown is that the percentage of land mass is greater than that of mineral deposits e.g. for the present to the Cretaceous and in the Lower Devonian. The reverse is true for the Upper and Lower Jurassic periods. No deposits of Permo-Triassic, Lower or Upper Carboniferous age plot in this area and no land occurs in this region on the Ordovician/Silurian reconstruction.

b) Table 6.4, 35°N - 35°S, Section B, 0-450 m.y.

i) Minimum Age-Derived Palaeolatitudes. The overall pattern shown by deposits in this area is that the percentage of mineral depsits is greater than the percentage of land mass. The exception is the Lower Jurassic period. There is a strong correlation between the relative distributions of land and mineral deposits in the Eocene, Lower Devonian and Ordovician/Silurian.

Table 6.4: % Continental Mass: % Mineral Deposits for each Rotation.

PERIOD NAME	ROTATION (M.Y.)	NCLE DATA >35° N (SECTION A) % CONT AREA	NCLE DATA >35° N (SECTION A) % MIN DEPOSITS	NCLE DATA 35°-35° (SECTION B) % CONT AREA	NCLE DATA 35°-35° (SECTION B) % MIN DEPOSITS	NCLE DATA >35° S (SECTION C) % CONT AREA	NCLE DATA >35° S (SECTION C) % MIN DEPOSITS
PRESENT	0	50.0		24.4		25.7	
	MIN AGES		15.6		58.0		26.4
	MAX AGES		12.1		57.0		30.8
EOCENE	50	35.0		46.5		18.6	
	MIN AGES		32.0		48.0		20.0
	MAX AGES		16.2		70.3		13.5
CRET	100	31.8		48.0		20.2	
	MIN AGES		23.0		59.0		18.0
	MAX AGES		24.1		55.6		20.4
U.JUR	130	32.4		43.7		24.0	
	MIN AGES		31.3		56.3		12.5
	MAX AGES		25.8		46.2		10.8
L.JUR	200	24.9		51.0		24.1	
	MIN AGES		63.9		33.3		2.8
	MAX AGES		67.9		32.1		0.0
PERMO - TRIASS	250	15.4		49.1		35.1	
	MIN AGES		1.5		93.9		4.5
	MAX AGES		0.0		100.0		0.0
U.CARB	300	10.0		42.7		47.4	
	MIN AGES		2.0		89.8		8.2
	MAX AGES		0.0		86.1		13.9
L.CARB	350	5.2		45.3		49.6	
	MIN AGES		0.0		80.6		19.4
	MAX AGES		0.0		88.2		11.8
L.DEV	400	4.8		51.6		43.7	
	MIN AGES		7.4		51.9		40.7
	MAX AGES		3.0		51.5		45.5
SIL & ORD	450	0.0		59.4		40.6	
	MIN AGES		4.3		66.0		29.8
	MAX AGES		3.6		71.4		25.0

ii) Maximum Age-Derived Palaeolatitudes. The periods which show correlation between percentages of land and mineral deposits are the Cretaceous, Upper Jurassic and Lower Devonian. Otherwise the results are the same as those for the minimum age-derived palaeolatitudes. However during the Permo-Triassic period all the mineral deposits occurred in this equatorial region.

c) Table 6.4, >35°S, Section C, 0-450 m.y.

i) Minimum Age-Derived Palaeolatitudes. The percentage of land mass is greater than that of mineral deposits for the Lower and Upper Jurassic, Lower and Upper Carboniferous and in the Ordovician/Silurian period. The converse is true for the Permo-Triassic. There is some correlation of the relative distributions of land mass and mineral deposits in the Lower Devonian and from the Cretaceous to the present.

ii) Maximum Age-Derived Palaeolatitudes. There is a higher proportion of land mass than mineral deposits for the majority of periods i.e. Eocene, Upper Jurassic, Lower and Upper Carboniferous and the Ordovician/Silurian period. The only exception is at present. Some correlation occurs in the Cretaceous and Lower Devonian between the two proportions and no deposits of Lower Jurassic or Permo-Triassic age plot in this area.

Previously all results determined from Tarling's reconstructions have been compared with the data derived from BP palaeogeographic reconstructions. So the relative distributions of land mass and mineral deposits according to the BP maps are given in Table 6.5.

a) Table 6.5, >30°N, Section A, 50-200 m.y.

i) Minimum Age-Derived Palaeolatitudes. The percentage of land mass is greater than the percentage of mineral deposits in this area from the Eocene to the Lower Jurassic period.

ii) Maximum Age-Derived Palaeolatitudes. The relative distribution of continental mass and mineral deposits are very similar for the Cretaceous and

Table 6.5: % Continental Mass: % Mineral Deposits for each rotation.

		BP DATA >30°N (SECTION A)		BP DATA 30°N-30°S (SECTION B)		BP DATA >30°S (SECTION C)	
PERIOD NAME	ROTATION (M.Y.)	% CONT AREA	% MIN DEPOSITS	% CONT AREA	% MIN DEPOSITS	% CONT AREA	% MIN DEPOSITS
EOCENE	50	52.5		16.5		31.0	
	MIN AGES		40.0		50.0		10.0
	MAX AGES		28.4		59.5		12.2
CRET	100	42.5		21.3		36.2	
	MIN AGES		35.6		47.5		16.9
	MAX AGES		38.5		42.3		19.2
U.JUR	130	39.7		21.3		39.0	
	MIN AGES		30.8		53.8		15.4
	MAX AGES		38.6		52.9		8.6
L.JUR	200	50.8		24.2		25.0	
	MIN AGES		58.3		41.7		0.0
	MAX AGES		60.7		39.3		0.0

Upper Jurassic periods. The proportion of land mass is much larger than the percentage of mineral deposits in this area during the Eocene. The reverse is true for the Lower Jurassic period.

b) Table 6.5, 30°N - 30°S, Section B, 50-200 m.y.

i) Minimum Age-Derived Palaeolatitudes. The results for this area i.e. 30° north to 30° south are notable as the percentage of mineral deposits is greater than the percentage of land mass for every period from 50 to 200 m.y. ago.

ii) Maximum Age-Derived Paleolatitudes. These produce exactly the same results as those described for minimum age-derived palaeolatitudes.

c) Table 6.5, >30°S, Section C, 50-200 m.y.

i) Minimum Age-Derived Palaeolatitudes. The Eocene to Upper Jurassic periods show that the percentage of land mass is much greater than that of mineral deposits for latitudes above 30° south. No mineral deposit examples of Lower Jurassic age plot in this area.

ii) Maximum Age-Derived Palaeolatitudes. These palaeolatitudes give exactly the same results as those described above for the minimum age-derived palaeolatitudes.

6.8 The Results in Tabular Form

The palaeolatitudes are listed in tabular form in Appendix I. A table was compiled for each mineral deposit type comprising all the examples used. Data on each example includes the mine/deposit name, its location, present co-ordinates (computer coded as described in Chapter 5, section 5.3.1), age of mineralisation in millions of years and the palaeolatitudes determined from the Tarling and BP reconstructions. The column headed `ROTATION' requires more in the way of an explanation. The number on the left hand side of the column i.e. 50, 100...450 indicates the age of the reconstruction from which the palaeolatitude was determined. The abbreviated continental names on the right hand side of the column represent the continental mass upon which the deposit was thought to have been situated at the time of formation. This procedure has been described in Chapter 5, section 5.3.2. The accession numbers (A.No.) were used for the correlation of data during the research and are of no relevance to the conclusions.

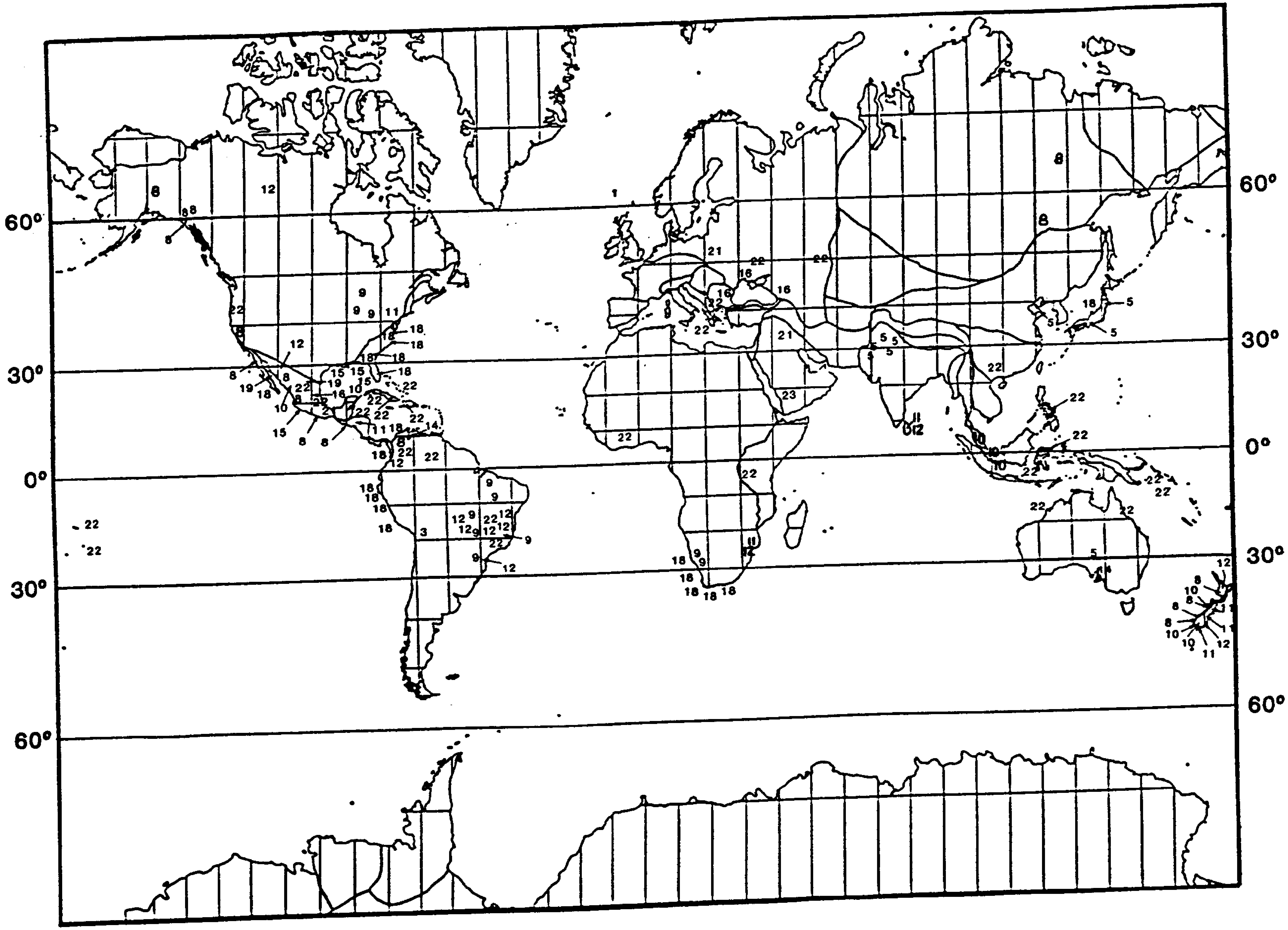


Figure 6.15a. Distribution of sediment-hosted mineral deposits being formed at present.
Numbers represent mineral deposit types - key as in text (page 98).

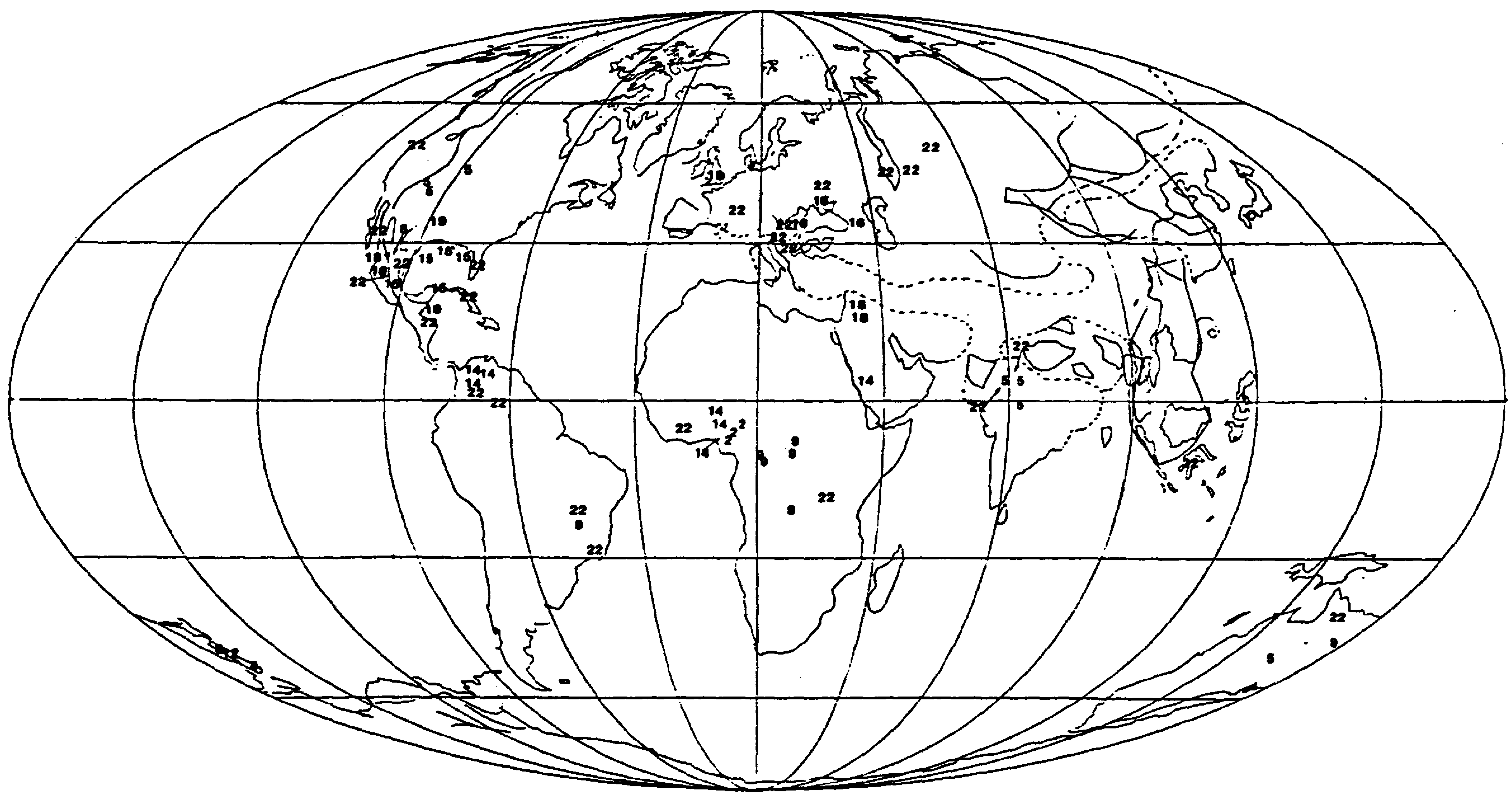


Figure 6.15b. Molleweide Projection of 'Eocene' (50m.y.) Palaeogeography based on palaeomagnetic data of Tarling (1983) as given in Chapter Five. Latitude and longitude grids at 30° intervals. Numbers represent mineral deposit types - key as in text (page 98). Palaeolatitudes determined from minimum ages of mineralisation.

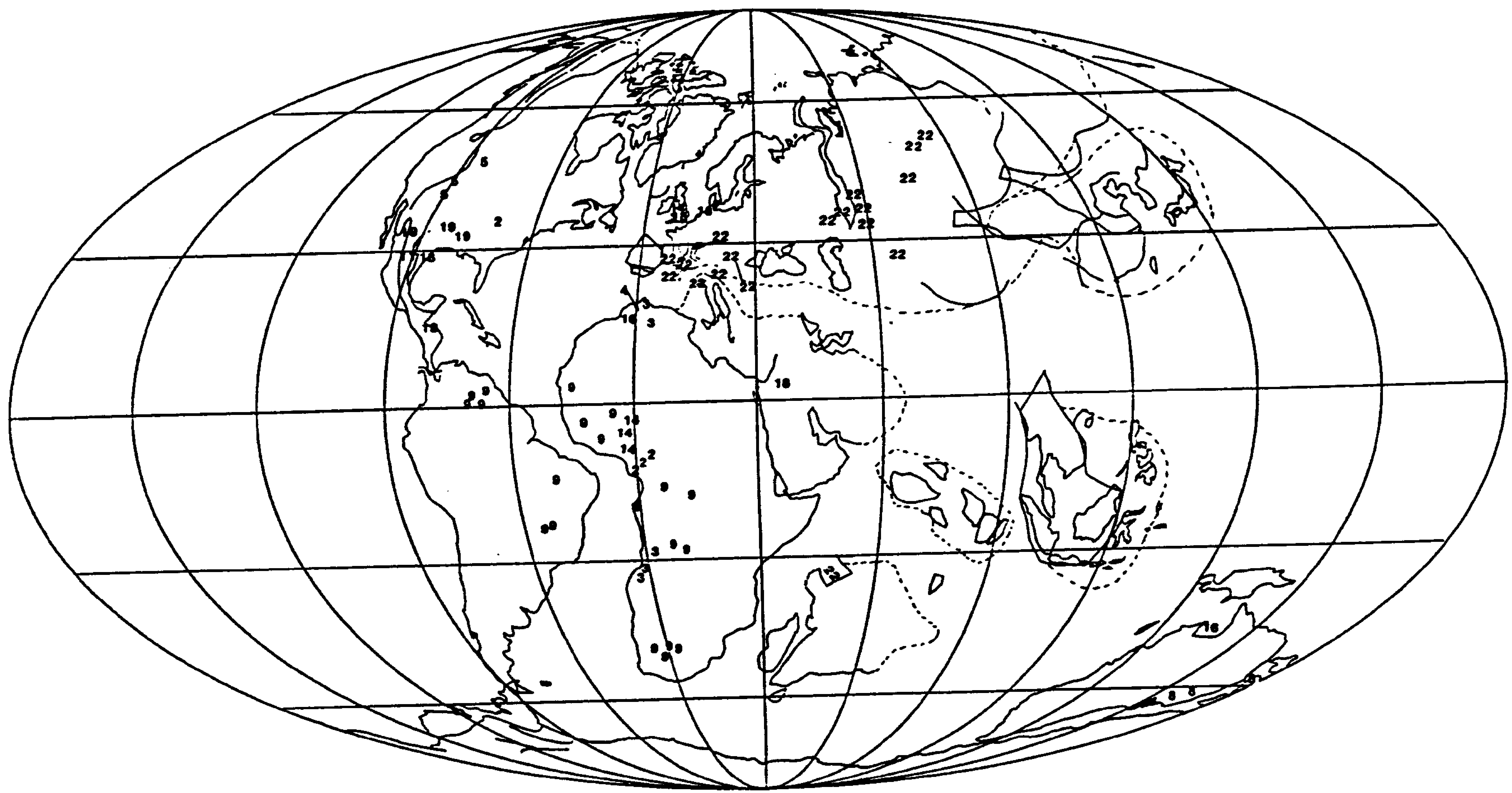


Figure 6.15c. Molleweide Projection of 'Cretaceous' (100m.y.) Palaeogeography. Key as in text (page 98).

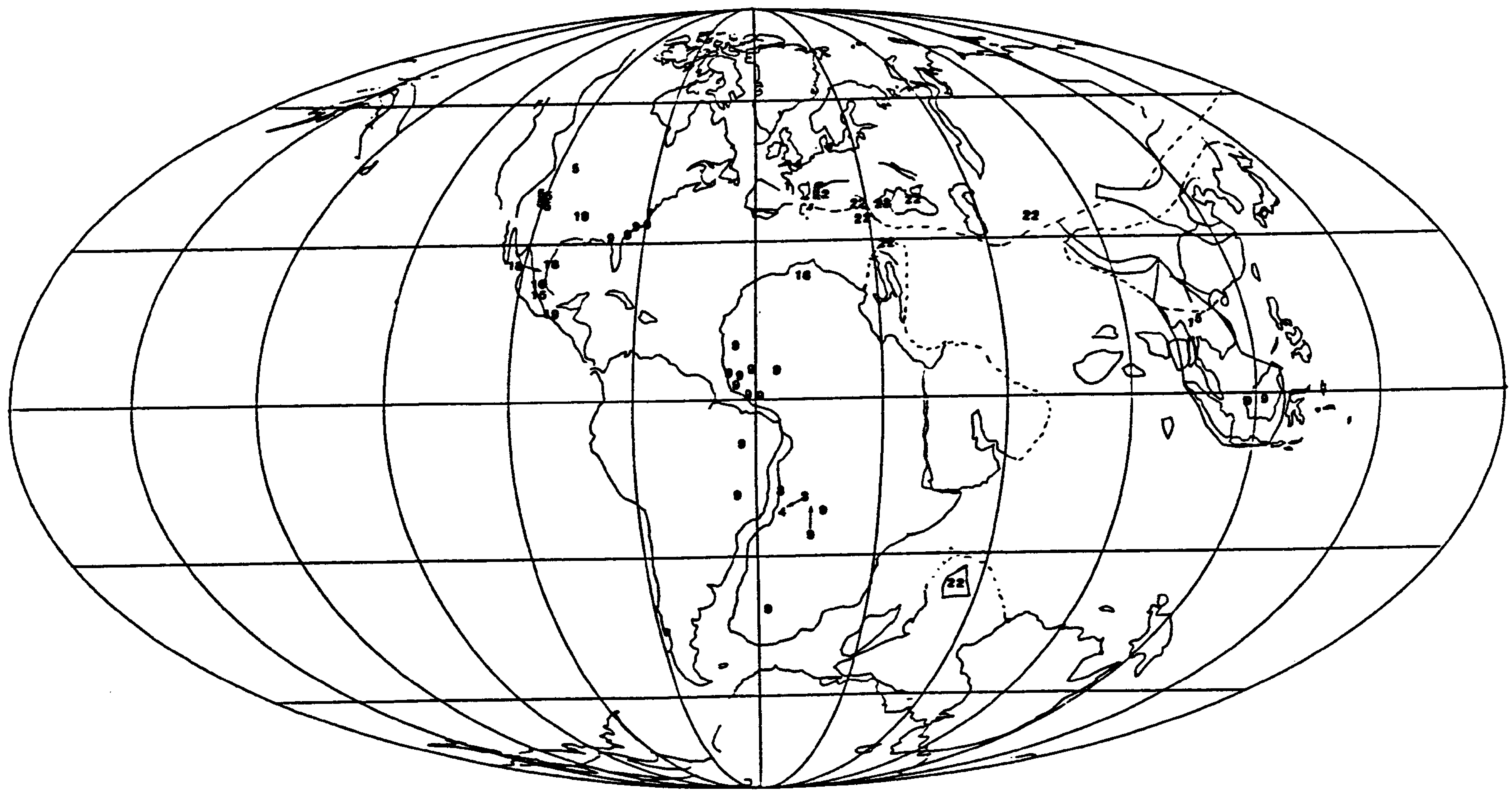


Figure 6.15d. Molleweide Projection of 'Upper Jurassic' (130m.y.) Palaeogeography. Key as in text (page 98).

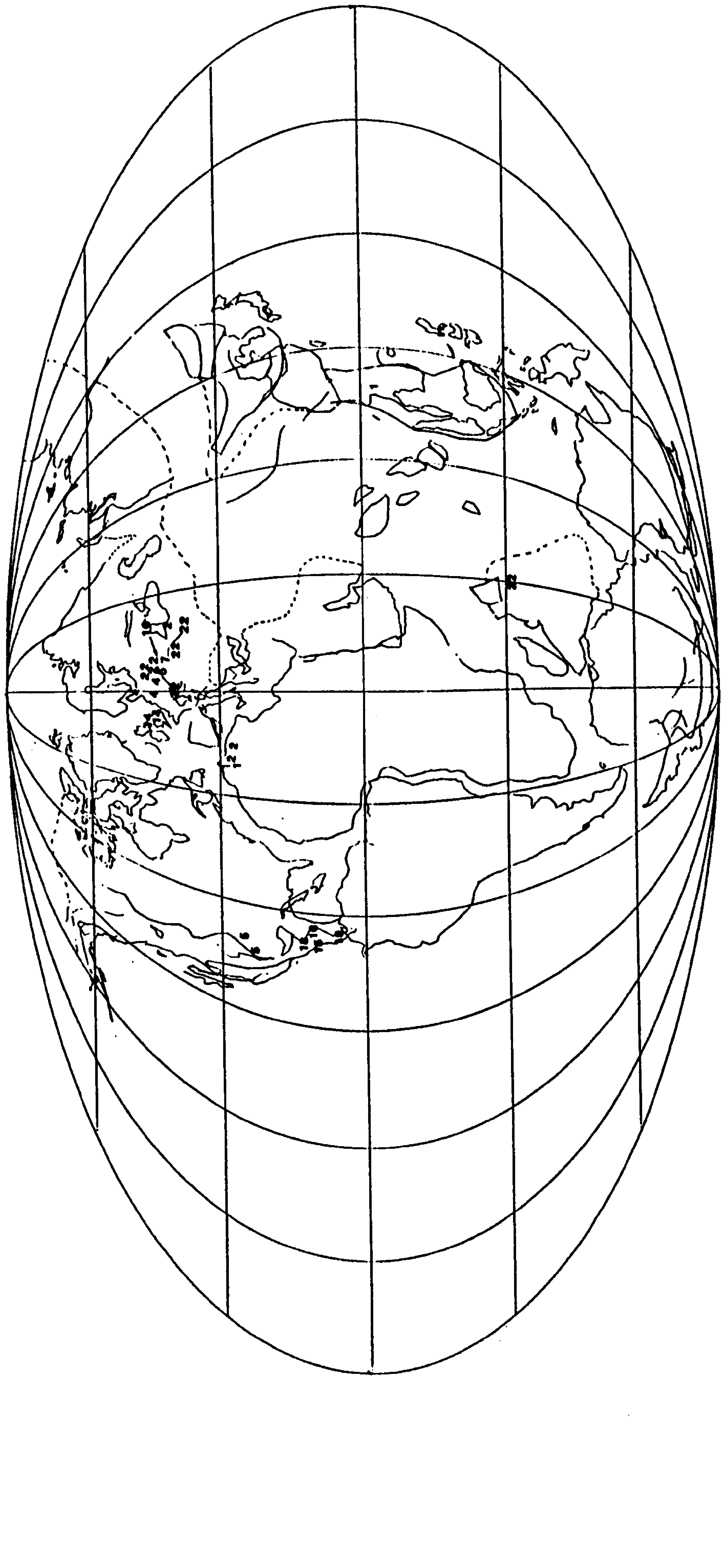


Figure 6.15e. Molleweide Projection of 'Lower Jurassic' (200m.y.) Palaeogeography. Key as in text (page 98).

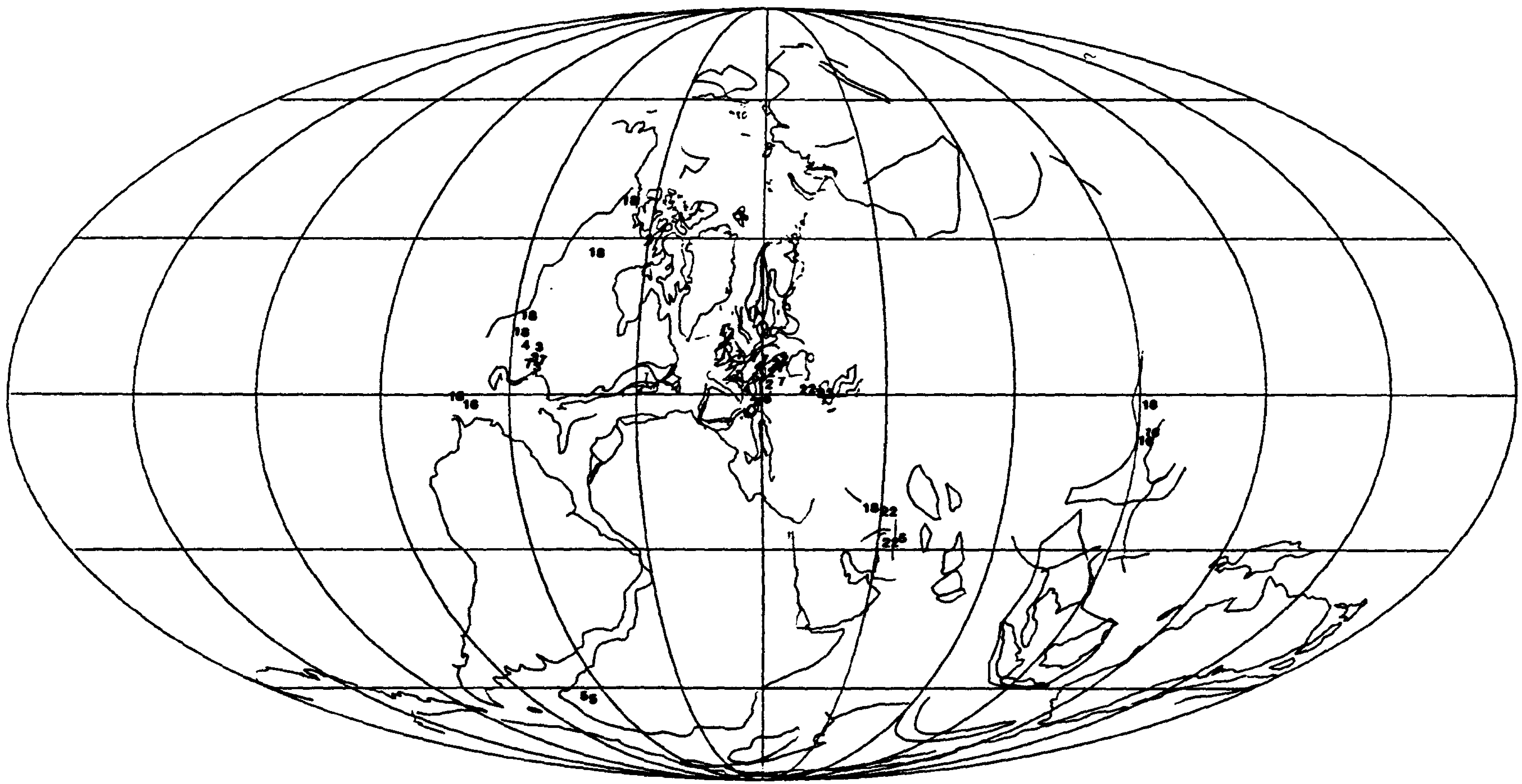


Figure 6.15f. Molleweide Projection of 'Upper Permian' (250m.y.) Palaeogeography. Key as in text (page 98).

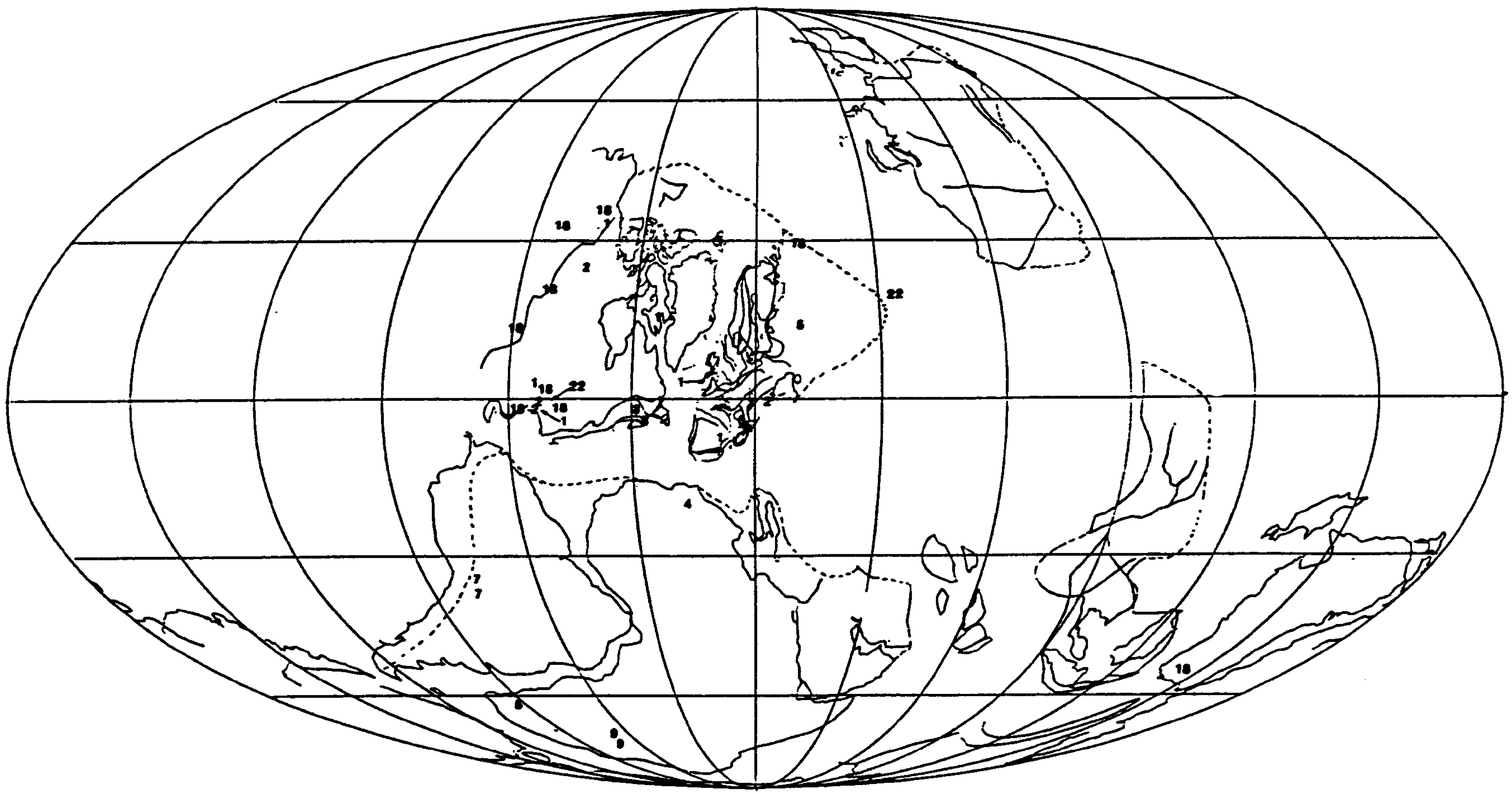


Figure 6.15g. Molleweide Projection of 'Upper Carboniferous' (300m.y.) Palaeogeography. Key as in text (page 98).

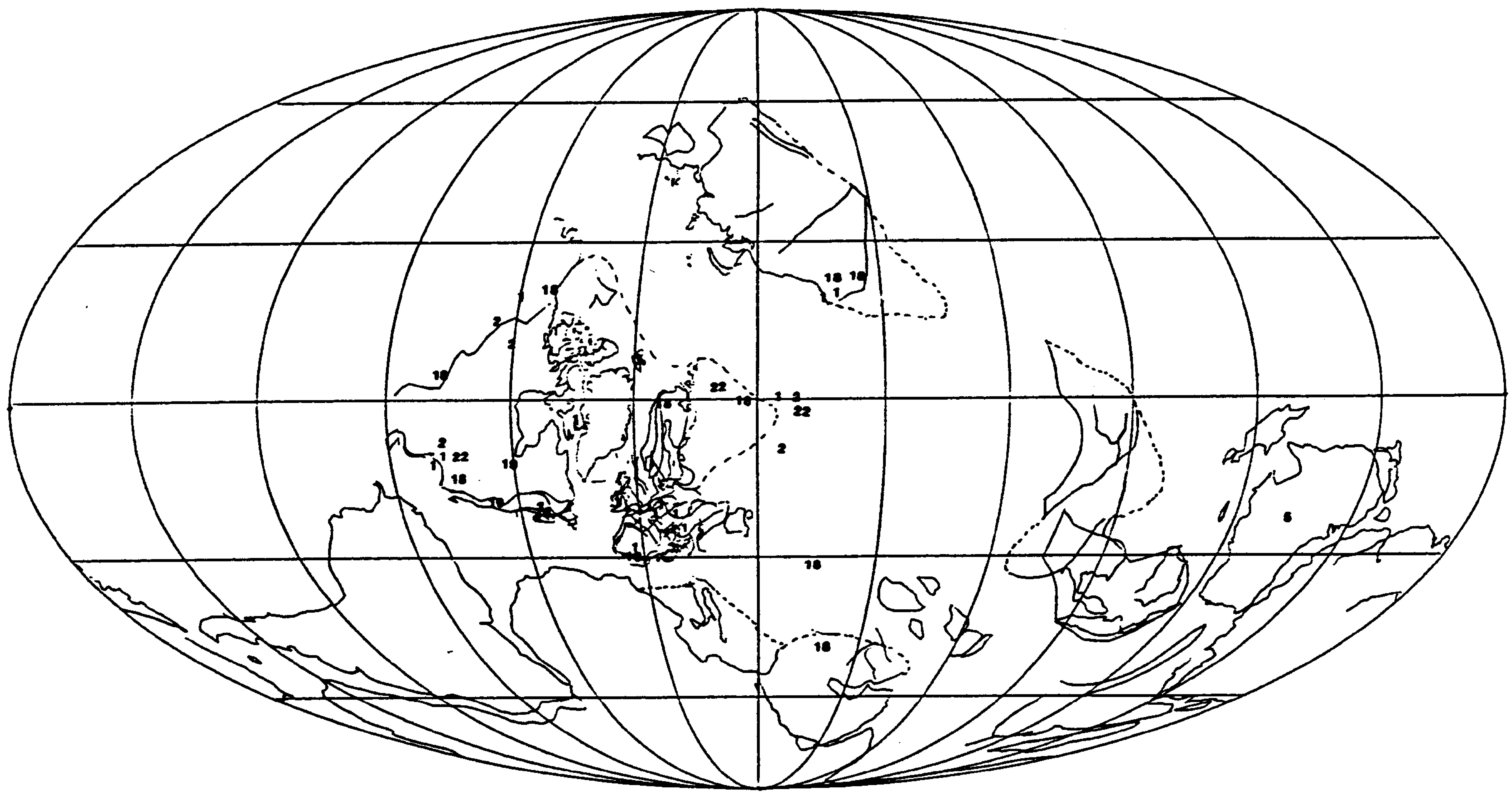


Figure 6.15h. Molleweide Projection of 'Lower Carboniferous' (350m.y.) Palaeogeography. Key as in text (page 98).

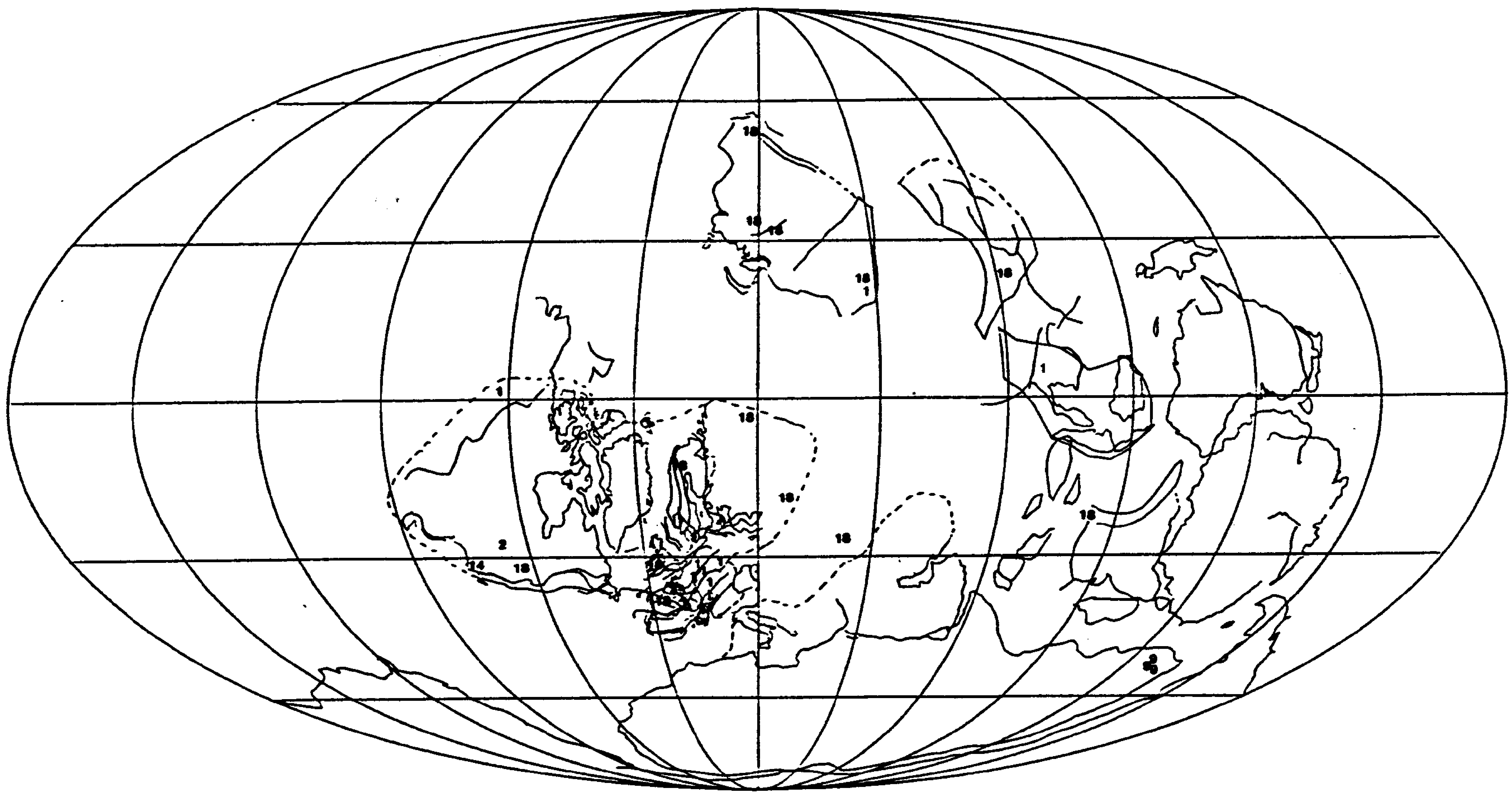


Figure 6.15i. Molleweide Projection of 'Lower Devonian' (400m.y.) Palaeogeography. Key as in text (page 98).

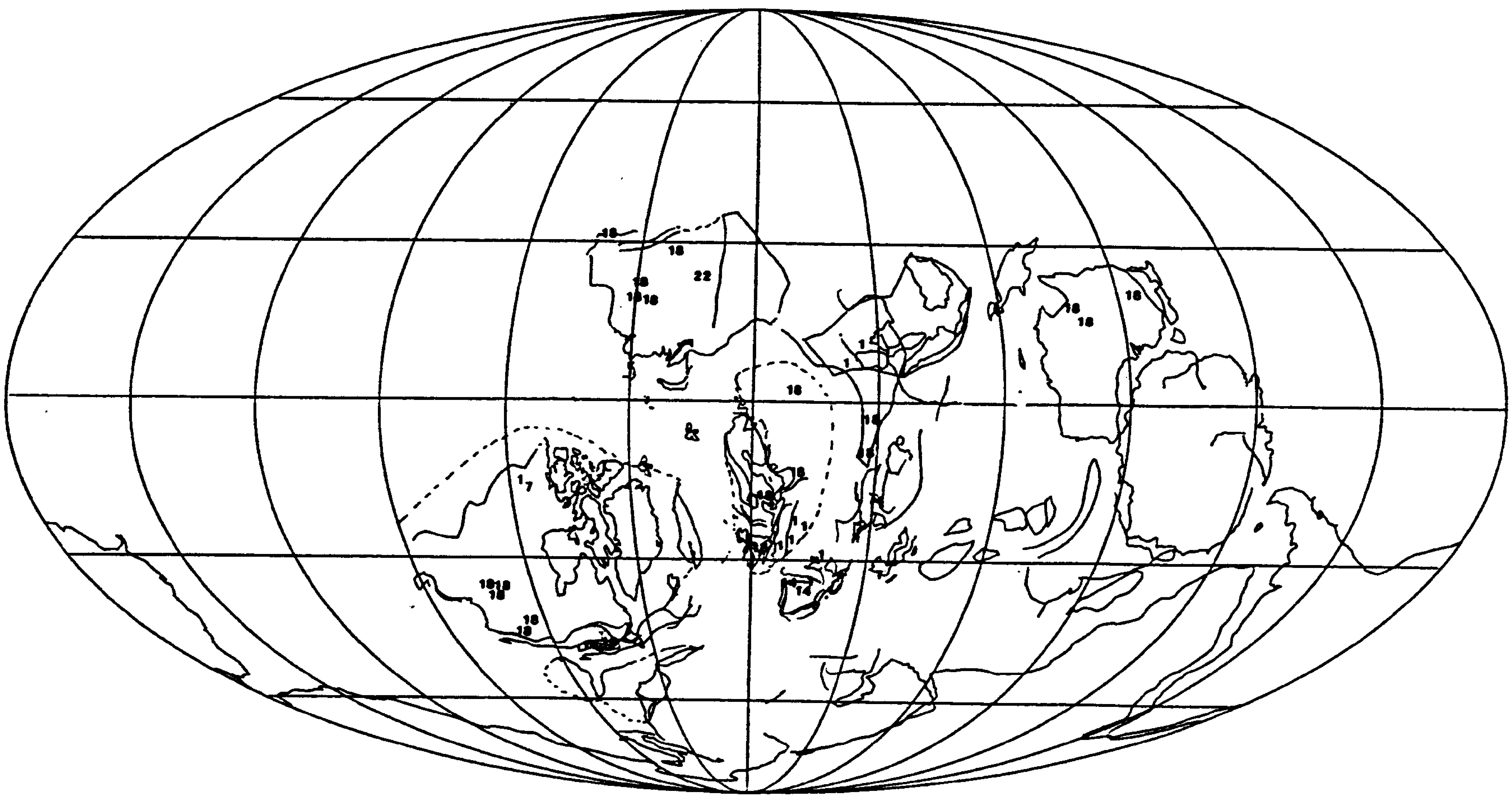


Figure 6.15j. Molleweide Projection of 'Ordovician/Silurian' (450m.y.) Palaeogeography. Key as in text (page 98).

6.9 The Results in Diagrammatic Form.

The palaeogeographic reconstructions of Tarling labelled Figures 6.15 a-j represent 50 million year time slice maps from 0 to 450 m.y. ago. The palaeolatitudes plotted are those determined from the minimum ages given for the deposits. These maps are not fully described here. Instead the results for individual deposit types as shown in the previous histograms have been discussed in detail.

6.10 Conclusion

Considering the northern and southern hemispheres together SSCUs have the narrowest latitudinal range of occurrence i.e. 35° from the equator. SSPB, SHBM and SDEX deposits have a range of 45°, LSBMs, OOFEs and MNFMs occur within 50° of the equator. PLAUs are found within 60°, SSUV and LATOs 65°, PHOS 70° and PLDI deposits occur up to 75° of latitude from the equator. From the standard deviation and coefficient of variation the variability of the distributions of most of the sediment-hosted deposit types discussed in this thesis is very similar.

The deposits which show the greatest variabilities are the placers, OOFE and PHOS groups which are clearly syngenetic in origin. However within those of least variability the MNFM and LATO groups are the only types which are generally considered to be syngenetic. The remainder are thought to be epigenetic to varying degrees i.e. LSBMs are considered to be epigenetic whereas SSCU could be early diagenetic. It is difficult to assess the significance of these data. The sample size of many of the mineral deposit groups is very small (i.e. < 30) so the reliability of the statistics may be questionable. Secondly, many of the placer and LATO deposits are 1 m.y. old or younger whereas the majority of deposits with least variability are older than

50 m.y. so the statistics have been calculated on palaeolatitudes derived from palaeogeographic reconstructions rather than from actual latitudes. However it may be that syngenetic deposits encounter more numerous occasions where all the conditions of formation are met and so the frequency and latitudinal extent of deposit occurrences is greater. But the conditions of the formation of epigenetic deposits are more limited and more strictly confined hence there is less variability in their palaeolatitudes of distribution.

Table 6.2 shows the percentages of land mass and mineral deposits in the northern and southern hemispheres for each geological period. The general trend is that there is a similar proportion of land mass to mineral deposit occurrence in each hemisphere for a given period. However latitudinal zones of 60° produce a different trend. It is evident from Tables 6.3 and 6.4 that the rate of mineral deposit formation within certain latitudes is not directly related to the percentage of continental mass in the same latitude zone. This is also true for the data derived from the BP maps shown in Table 6.5. These results indicate that the distribution of mineral deposits in each 60° latitude zone is uneven otherwise the trend shown in Table 6.2 for the northern and hemispheres would be repeated in the narrower latitudinal belts. The findings clearly substantiate the case for a palaeolatitude control upon the formation of some mineral deposit types.

A number of Ordovician/Silurian deposits are anomalous. This is probably a reflection of the poor quality of the data available for deposits of this age - in terms of mineralisation and deposit age estimates and also the palaeomagnetic data from which palaeogeographic reconstructions were derived.

In general the discrepancy between the two sets of palaeolatitudes (i.e. those of Tarling and BP) is very small. There is only one instance when a palaeolatitude plots in both the northern hemisphere and the southern hemisphere i.e. from the 130 m.y. reconstruction with regard to the region now represented by Afghanistan and Pakistan. However these results are questionable

Table 6.6: Summary table showing ranges of palaeolatitudes derived from minimum and maximum ages of deposits.

DEPOSIT TYPE	MINIMUM AGE NORTH	MINIMUM AGE SOUTH	MAXIMUM AGE NORTH	MAXIMUM AGE SOUTH
LSBM	50°	30°	50°	30°
OOFE	50°	50°	50°	50°
SSCU	25°	35°	20°	35°
SSPB	45°	25°	45°	25°
SHBM	45°	35°	20°	35°
SDEX	30°	45°	30°	45°
SSUV	45°	65°	50°	65°
PLAU	65°	60°	65°	60°
PLDI	50°	75°	50°	75°
PLSN	20°	50°	20°	50°
PLOX	45°	50°	45°	50°
PLOT	70°	50°	70°	50°
MNFM	50°	50°	45°	50°
LATO	55°	65°	55°	65°
PHOS	60°	70°	55°	70°

as there was uncertainty as to whether this region should be rotated with the main African continent or the Indian subcontinent. Generally the greatest discrepancy in palaeolatitude sets was for the Indian results, probably due to inadequate palaeomagnetic observations. The other area from which differing results are produced is Central America. Again this is to be expected because of uncertainty of which continent to rotate this region with. The other area composed of micro-continents is Sundaland (i.e. Indonesia, Thailand etc.) which may be a potential source of error shows very few discrepancies between the results from Tarling and BP. Otherwise the two sets of palaeolatitudes are +/- 10° of each other.

An examination of the palaeolatitudes for specific regions reveals there are no particular regions which have anomalous palaeolatitudes on a series of contiguous time slices. However the area known as Europe today at 200 m.y. is notable as many mineral deposit types which are generally in low latitudes plot at higher latitudes than would be expected. These include LSBM 45°N; OOFE 45°N; SHBM 42°N; SSPB 44°N; MNFM 43°N. Generally the BP palaeolatitudes determined for the same area at that time are still outside the expected low latitude limits for these deposit types despite being about 5° lower than the Tarling determined palaeolatitudes. This may be due to the prevailing climatic conditions of the Earth at that time (see Chapter Eight, section 8.2.2.3).

In conclusion there is a strong indication that there is a palaeolatitude control upon the formation of some types of Phanerozoic sediment-hosted mineral deposits. However the palaeolatitudinal range of different deposit types is seen to vary considerably as summarized below.

Some deposits are concentrated (at least in part) in the equatorial rainfall belt and they include the LSBM, OOFE, SHBM, SDEX, SSUV, PLDI and PLSN deposits. Conversely the SSCU, PLAU, PLOT, PLOX, MNFM and PHOS groups are suppressed in numbers in this region. The warm arid climatic belt from 15° to

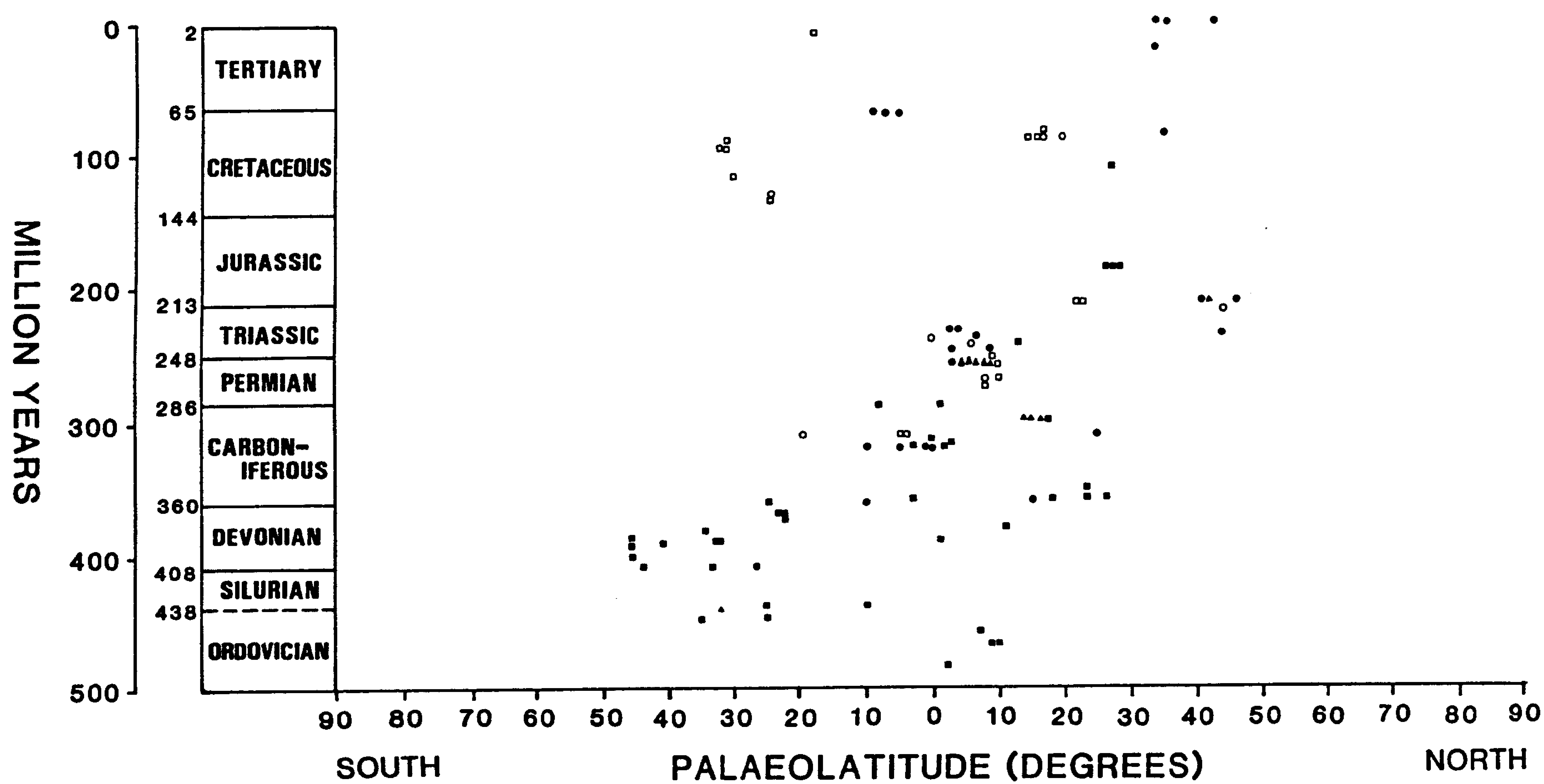


Figure 6.16a. Mineral deposit types showing a marked low latitude control upon their formation. Palaeolatitudes derived from minimum ages of deposits.
● LSBM, □ SSCU, ○ SSPB, ▲ SHBM, ■ SDEX.

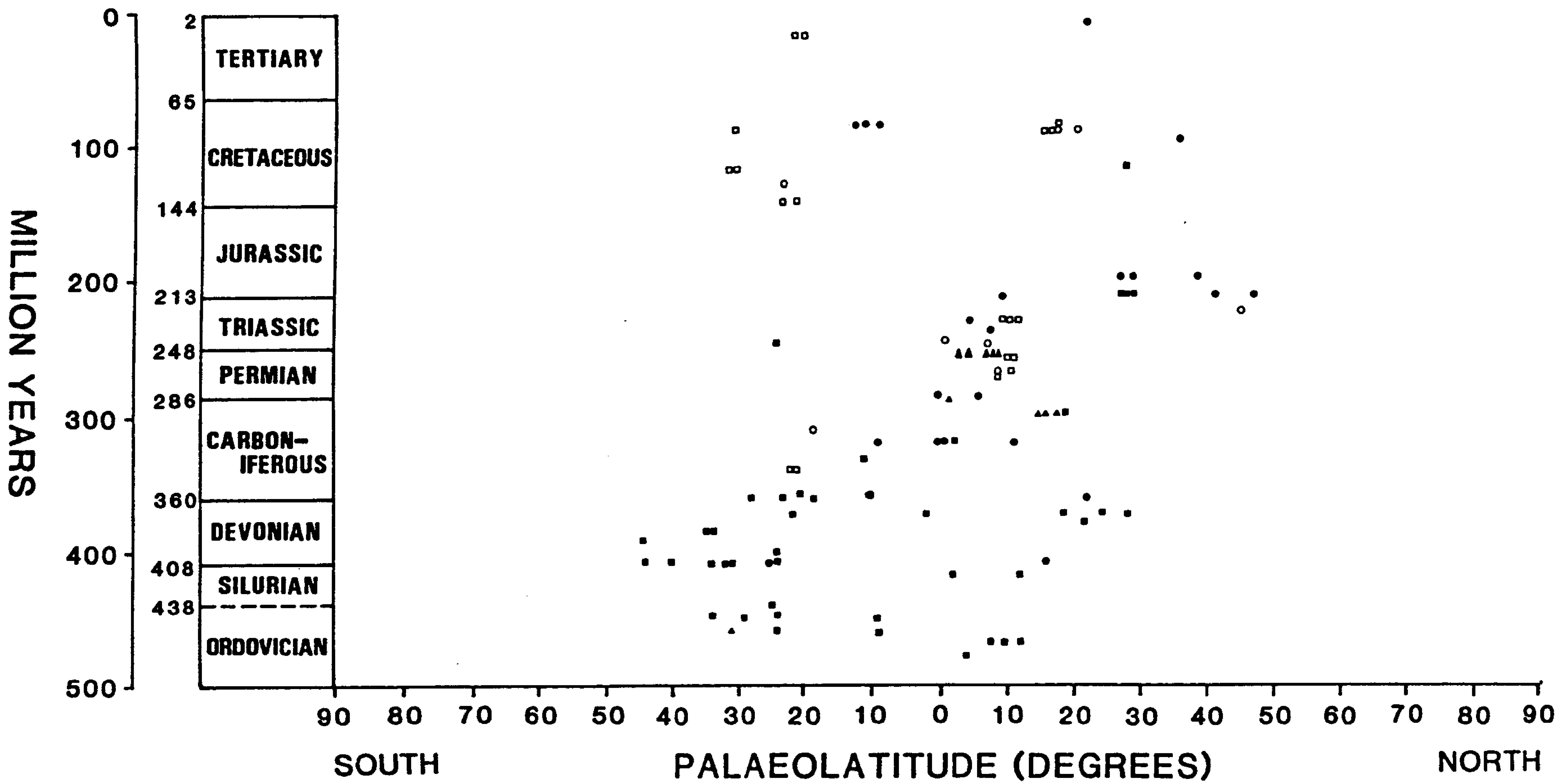


Figure 6.16b. Mineral deposit types showing a marked low latitude control upon their formation. Palaeolatitudes derived from maximum ages of deposits.
● LSBM, □ SSCU, ○ SSPB, ▲ SHBM, ■ SDEX.

40° shows a concentration of SSCU, PLOT, MNFM and PHOS deposits but very few examples of the SSPB, SHBM, SSUV, PLSN and PLOX groups occur in this latitude zone. Finally OOFE, SSUV, PLAU, PLDI, MNFM and PHOS deposits all show peaks in their distribution in the warm temperate rainfall belt (about 40° - 55°).

In the introduction it was mentioned that a low latitude control upon the formation of mineral deposits was particularly important for the reasons outlined. So observations from the data on this matter are detailed below. A 30° latitude limit upon the distribution of deposits at their time of formation does appear to exist for a number of deposit types examined here. Many of the examples which do not occur within this area (e.g. some LSBM deposits) are found within the error bars which may be in the order of 5° of latitude. One group of deposits which do not show the expected latitude control upon their formation are the SSUV deposits. The mineral deposit types which particularly illustrate the low latitude control upon their formation are the LSBM, SSCU, SSPB, SHBM and SDEX groups (Figures 6.16 a and b). A remarkably large proportion of these deposits plot within the zone 35° north and south of the equator - a trend which is observed in both the minimum and maximum age-derived palaeolatitude plots. Reasons for the anomalous positions of some of the examples have been given previously. In Chapter Three, section 3.2.7, a question was posed as to whether a continuum of stratabound base-metal (Cu, Pb, Zn) deposit types existed. These results seem to indicate that there is at least one genetic control upon some groups of mineral deposits which may be applicable to all types.

CHAPTER SEVEN

DISCUSSION: PART ONE: PALAEOMAGNETISM

The palaeomagnetic techniques used, the selection of data and the assumptions which have been made during the course of the research must now be more fully evaluated in view of the results which they have produced.

7.1 Errors

The total effect of the errors involved in the thesis is not only unknown but is also very difficult to determine. The different types of error occur in many aspects of this research and are given below.

7.1.1 Errors in the Palaeomagnetic Methods

In view of the difficulties associated with the selection of palaeomagnetic data as detailed in Chapter Two, it must be emphasised that much subjective evaluation is involved in such an analysis and selection procedure. These assessments should be most likely to agree when the data have also been objectively analyzed. However such objective information is not always available, particularly for earlier palaeoreconstruction studies (Tarling, 1985b).

It is obvious that any successful application of palaeomagnetism must depend upon being able to show that the assumptions which are made are fully justified. With regard to palaeomagnetism these concern:

i) the long term nature of the geomagnetic field,

ii) the origin of the permanent magnetisations of rocks,

iii) the magnetic stability of rocks.

The above have been described and discussed in both Chapter Two and Chapter Four, section 4.2 and it is not necessary to examine these assumptions further. With the acceptance of the above, the sources of error within palaeomagnetic techniques must now be outlined.

7.1.1 (a) Errors in Laboratory Techniques.

In laboratory determinations the following criteria are particularly important if a standard, relatively reliable procedure is to be achieved and maintained.

i) Preferably a minimum of five separate sites (N = 5) must have been sampled, with approximately five samples (n = 5) collected from each site thus tending to reduce the effect of secular variations and other geomagnetic fluctuations.

ii) These samples have been subjected to either alternating field (AF) or thermal (TH) demagnetisation procedures to establish their magnetic stability and number of magnetic components.

iii) α 95° for each site must not exceed 10°.

iv) The remanence of each individual rock specimen must have been accurately measured i.e. the direction of each vector must be within 5°, and its intensity within 5%. Such a requirement is conventionally met if the intensity of magnetisation is at least ten times that of the noise level of the magnetometer being used. The requirement is also satisfied if the specimen is magnetically homogeneous. However in some circumstances repeat measurements allow the vector to be defined despite a lower intensity and detailed sampling may also allow small-scale inhomogeneities to be averaged out (Tarling, 1985b).

Such criteria are inadequate as inhomogeneous and anisotropic data may have been included, but such properties are frequently either not measured or not reported (Tarling, 1983). Also these criteria are not themselves beyond question e.g. the minimum number of observations is much too small to ensure adequate averaging of secular variations. In addition to the inadequacies in these criteria it is also impossible to apply them rigidly e.g. much of the data from the USSR are not readily available in sufficient detail to make full evaluation. In these and many other cases the construction of the polar wander path has necessarily included the large element of subjective assessment mentioned in the introduction. This is discussed more fully in section 7.1.1.d.

7.1.1 (b) Errors in the determination of Primary Magnetisation

While magnetic stability is a necessary prerequisite if the primary (original) magnetisation is to be retrieved from a sample, secondary magnetisations could readily have arisen that may have a higher stability than the primary remanence. The original magnetisation may be unstable to later physical (e.g. temperature or pressure) or chemical processes and so be eliminated or modified over time. If the original magnetisation survives it may coexist with later, secondary magnetisations and be difficult to isolate and identify (Park, 1983). Determining whether the natural remanent magnetisation or residual permanent magnetisation of a sample is primary or not is often difficult and greatly influences the inferred palaeolatitude. Therefore the two basic problems with the interpretation of remanences are firstly whether any primary magnetisation remains in the sample to be detected and secondly is it possible to isolate this primary remanence from all secondary components?

In particular, any chemical changes that result in the creation of new haematite crystals are likely to be associated with a chemical remanence that may have an extremely high coercivity and high blocking temperature. If the primary remanence is carried by magnetite (of lower coercivity and lower blocking temperature) the secondary component will then have the highest stability. It is therefore essential to establish, if possible, the actual ages of the different components of remanence (Tarling, 1983).

To determine whether the magnetic minerals of a rock are of primary or secondary origin the textures of the minerals should be examined. Those minerals formed at high temperatures commonly have different textures from those that have formed by low-temperature oxidation (Haggerty, 1976). In addition the textural relationship of the minerals in a sedimentary rock will frequently allow an assessment of the sequence of diagenetic and post-diagenetic changes that have occurred. Identification of the minerals carrying the remanence can then be related to the petrological information and realistic evaluations can then be made of the probable age of any stable magnetisations associated with formation of specific minerals. However this is reliant upon new minerals being formed at the same time as the stable magnetisation. Such a method is invalid if the magnetisation alone is affected by later processes and no minerals are formed during these processes.

Chemical demagnetisation techniques, such as passing acids through permeable sediments to preferentially remove some components of the rock, may also aid in the determination of a magnetisation associated with detrital grains (Collinson, 1967). In the case of sediments carbonate cements can be readily removed together with the magnetisation associated with the primary grains (Henshaw and Merrill, 1980). The main difficulty with this technique is the necessity for high permeability e.g. in low permeability sediments the magnetisation associated with surface detrital grains may be lost by chemical action whilst the interior cement of the sample may be largely unaffected. Also the technique does not discriminate between detrital grains unchanged since their deposition and those which have been replaced or oxidized in situ but are still detrital grains. Petrological studies remain essential for such distinctions to be made.

A major test to establish which components are primary is the fold/tilt test which allows dating of the remanence relative to the time at which the tectonic movements took place. In most cases a remanence that has survived tectonic disturbance is likely to have been acquired very early in the rock's history but it could still have been acquired as a result of chemical changes after the rock had originally formed. Tarling (1983) considered that the largest source of error in orientation and sample collection may be in the assessment of the tectonic correction by which the observed directions of remanences are converted to their original orientation when their primary remance was acquired. In the selection of palaeomagnetic data Tarling (1983) noted that frequently it was assumed that observed magnetisations of samples were primary if no field or tilt tests were available to determine otherwise. On the occasions this assumption was made an element of unquantifiable unreliability was introduced into the results.

One of the major uncertainties in most palaeomagnetic studies is therefore distinguishing whether the magnetic ages of the rocks sampled are the same as the ages of the rocks themselves. If this can be established then the next major source of error is usually in the determination of the correction required for tectonic disturbances since the magnetisation was acquired.

7.1.1 (c) The Importance of Autochthonous and Allochthonous Tectonic Regimes.

Fundamental to any conclusions drawn using palaeomagnetic data to determine the palaeolatitude of a particular region is the reliability of the polar wander path for that particular tectonic block. It is vital to establish that the rocks under consideration do actually belong to the same tectonic unit for which the curve has been derived. Such a dilemma is faced when areas of considerable crustal mobility (e.g. within orogenic zones) are examined as small elements within the area may have moved as discrete separate units.

Some mineral deposit types are commonly found in rift zones but individual fault blocks may have been rotated separately as in the Lebanon where the blocks are on a scale of 50 x 200 km (Freund and Tarling, 1979). While in other areas, such as Alaska, tectonic units may have been translated over several hundreds or thousands of kilometres (Tarling, 1983). Even in a localised case considerable errors would be induced if the behaviour of any one block was thought to reflect the total regional pattern of fault motion. In folded terrain the supracrustal layers may be highly mobile so can move separately from the basement (allochthonous) or they may be rigidly attached to the basement (autochthonous). For autochthonous tectonic blocks, samples from any one locality will provide information on the behaviour of the entire block, but this will obviously not be the case in an allochthonous region and the degree to which such areas are truly autochthonous, when compared with the continental craton, is not usually known.

In conclusion it is imperative to ensure that sites for rock sampling are all located in the same autochthonous tectonic block so that they are unlikely to have moved relatively to each other since they were originally magnetised. The definition of the areal extent over which rocks of identical age yield an identical palaeomagnetic pole should reduce errors caused by this type of inaccuracy but such studies have not yet been attempted for most orogenic areas of the world. However it is interesting to note that the recognition of allochthonous terrains may also be important. For example such recognition in the U.S. cordilleras was consequent upon discrepancies within palaeomagnetic data. These misfits thus led to the development of the microplate and accreted terrain hypothesis.

7.1.1 (d) Evaluation of the Pole Positions used for the Palaeogeographic Reconstructions.

The selection of the palaeomagnetic data used to construct the Phanerozoic polar wander curves for most continents fulfilled the criteria given in Chapter Two as far as possible and they are described more fully in Tarling (1983, 1985a, 1985b). The pole positions used to create the palaeoreconstructions from which the palaeolatitudes were taken are also given in Tarling (1983). The pole positions and confidence limits were calculated on total data whenever possible i.e. using all the pole determinations for sites located on that cratonic block of the continent and reputed to be of that specific age. This method was considered practicable for poles corresponding to ages of less than 300 m.y.

A greater element of subjectivity was introduced into the interpretation of results when the poles for any particular time had an oval distribution. This could represent either a progressive movement of the continent relative to the pole during that time interval or the presence of more than one magnetic component. If the oval distribution tended to 'string' to younger parts of the polar wandering curve it was interpreted as being indicative of the presence of later components of magnetisation. Hence a mean pole position was estimated to lie within the oval, but at a point furthest from the younger part of the curve. Alternatively when the poles were found to be widely dispersed, then either the mean direction was used or the pole was assessed from a consideration of the distribution pattern relative to the known younger parts of the polar wandering path.

Other authors have simply averaged the pole positions. With regard to the choice of poles for Silurian and Devonian continents Livermore et al (1985) found that the quality of available poles for these continents differed between continental fragments. As a consequence they did not attempt to apply vigorous selection criteria to every pole but endeavoured to find the most reliable data available for each continental fragment. Ziegler et al (1982) selected poles for Mesozoic and Cenozoic reconstructions from summaries of Irving (1960-1965) and McElhinny (1968-1978) palaeomagnetic data but they assessed the pole positions using the same criteria as Tarling i.e. those data that were obtained from samples that were insufficiently demagnetised, those that were from unstable tectonic areas or those which did not exhibit a primary remanence were rejected.

Tarling (1983) considered that another subjective element in the assessment of pole positions, and hence a potential source of error, is the assumption that continental block movements only showed considerable change if the continent had collided with another continental block.(Most workers do not make this assumption). A sudden change in the direction of the motion along a polar wandering curve was normally considered to be associated with the time of an orogenic event in that continental block. e.g. the end Cretaceous Laramide orogeny could be expected to be related to a change in the North American polar wandering curve.

The polar wandering curves used in this project to construct the palaeoreconstructions are therefore subjective evaluations, especially for geological periods older than 300 m.y. However each curve has been assessed individually i.e. without consideration of implications for continental reconstructions. The exception is Antarctica which is of no consequence to this research as no mineral deposit examples were taken for that region. Virtually no pre-Permian palaeomagnetic data can be considered to meet fully the basic criteria required to establish that they represent a true average geomagnetic field direction for specific, known times. On this basis, any reconstructions based upon such observations must be strongly qualified by the need for further palaeomagnetic study. Many pole positions must still be interpolated between more reliable data e.g. those for the Lower Carboniferous are determined from Devonian and Upper Carboniferous or Permian data, and thus require to be evaluated against any other available controls (see section 7.1.2). Detailed reviews of the available data have been given by many authors (e.g. Morel and Irving, 1978 and 1981; Scotese et al, 1979; Ziegler et al, 1979; Van der Voo, 1982; Tarling, 1983) and so are not repeated here.

7.1.2 Errors involved in the production of Palaeoreconstructions.

The main source of error which affects the reliability of the reconstructions and hence the palaeolatitudes determined from them is the reconstruction of the reassembly. The projection of that reassembly as a map can give a misleading appearance of data although no real errors are involved in the construction. The Mollweide Projection was chosen in an effort to minimize the projection errors, being a type of equal-area projection in which the entire surface of the earth is depicted in one map. All parallels of latitude and the central (Greenwich) meridian are straight lines. Other meridians are curved and curvature increases towards the marginal meridians. The area delimited by two adjacent parallels and meridians is equal to any other area similarly enclosed. Therefore there is no distortion of areas (as occurs in the Mercator Projection) and the Mollweide-type is considered more suitable for showing global distributions of phenomena such as mineral deposits.

It has been shown (see section 7.1.1) that there are numerous potential sources of error in the selection and evaluation of palaeomagnetic data and that ultimately the reconstructions produced are of largely unknown reliability. Palaeomagnetic data alone are not yet sufficient to produce accurate palaeocontinental reconstructions for all periods. Indeed a number of different palaeogeographic reconstructions have been proposed for the Palaeozoic by various authors. Each set of reconstructions has been approached in a different manner e.g. Smith et al (1973 and 1981) and Morel and Irving (1978) use palaeomagnetic evidence alone; Scotese et al (1979) and Bambach et al (1980) gave palaeomagnetic evidence first priority, applying palaeomagnetic rather than palaeoclimatic data when discrepancies arose. All these palaeomagnetic-based reconstructions basically rely on the same data and yet are significantly different from one another indicating that the data provide fewer constraints on interpretation than may previously have been supposed (Boucot and Gray, 1983). As palaeomagnetism does not yet provide an infallible reconstruction other criteria can be used to aid such problems particularly in terms of the relative longitudinal positions between continents. Among the earliest reconstructions applying palaeoclimatic and tectonic informatiom together with palaeomagnetic data were those presented by Ziegler et al (1977a) in a study of the Silurian. The relative longitudinal positions of the continents can often be deciphered by studying the distributions of flora and fauna (e.g. McKerrow and Cocks, 1976; Ziegler et al, 1981b). Similarly the closure of ocean basins and the timing of continental-continental collisions place important constraints on the relative position of these palaeocontinents and must be taken into account in any palaeoreconstruction (Scotese et al, 1985).

Any of the disciplines mentioned would introduce additional potential sources of error other than those of palaeomagnetic data if they were used to assist in the development of palaeogeographical reconstructions. The only real test of the reliability of palaeomagnetic data is the geological "sense" (i.e. palaeoclimatic and biogeographical) which they make when applied to the production of palaeogeographic reconstructions. However if these geological aspects are used in the interpretation of palaeomagnetic data and the subsequent palaeogeographical reconstructions then the test becomes invalid for the argument is circular. However in this case the main aim of the research was not to establish the absolute reliability of the palaeomagnetic data but whether the level of known reliability at present allows the testing of the different models for the origin of mineral deposits from their palaeolatitudinal distributions. It is apparent from the discussion in the following chapter concerning climate and mineral deposit formation that the present reliability is sufficient for the purposes of this research. Each discipline is discussed in the following sections with an evaluation of the possible discrepancies which their application would introduce.

7.1.2 (a). The Palaeolongitude Problem.

The derivation of unique palaeoreconstructions would remain difficult even if well-dated, reliable palaeomagnetic records were available for more geographical regions and more finely divided time intervals. This is because of the longitudinal indeterminacy of palaeomagnetic data. It has been shown (see Chapter Two, section 2.3) that it is possible to put reconstructions into their former latitudes from the inclination and declination of the magnetic field preserved in continental rocks of the appropriate age but it is not so for the absolute longitudinal positions of fragments.

The problem of estimating the relative longitude separation of the major continents for the Late Mesozoic and Cenozoic (i.e. 180 m.y.) is assisted by the use of ocean floor data. However there are even greater difficulties with continental position during the Palaeozoic (i.e. prior to 245 m.y.) because virtually all the Palaeozoic ocean floor has been destroyed in subduction zones.

There are a number of solutions to the longitude problem which are briefly discussed here. Using the criterion of minimum motion between continents in time (e.g. Morel and Irving, 1978) both Mesozoic and Palaeozoic reconstructions have been made and the resultant maps were not too dissimilar to those based on ocean floor spreading (Irving, 1977).

An alternative to this method has been given in Chapter Two. When two continents are fixed with respect to each other their relative positions can be found by superimposing their apparent polar wander paths. Theoretically these should be identical, which is a rare occurrence. The problem in using this method for an arbitrary period is the evaluation of how much of the difference between two similar apparent polar wander paths is due to errors of measurement and how much to relative continental motion (Smith, 1985). Livermore et al (1986) considered that most of the contrasts between competing models for the Pangaean configurations were due to differences in palaeolongitude of the component fragments, so they used a version of the apparent polar wander path method to re-evaluate the palaeomagnetic data available for these configurations. They decided to examine the apparent polar wander paths of the component continental blocks for a longer period of time than was usual in an attempt to discount some of the models. A time window of 20 m.y. for data grouping was chosen as the minimum presently permitted by the number and quality of the available data. One conclusion that was drawn from this approach is that comparison of apparent polar wander path segments was preferable to the simple averaging of supposedly contemporaneous poles - an approach occasionally used by Tarling as described in section 7.1.1 (d).

Zonenshain et al (1985) wanted to estimate the true width of the Palaeozoic oceans (which can be estimated using kinematic data for Mesozoic and Cenozoic oceans) and to determine true plate motions. It was considered that the only good criterion for the determination of absolute motions is hot spot traces produced when plates pass over mantle plumes. Smith (1985) noted that provided hot spot traces were available for all continents, they could be placed in the same frame independently of any other data so solving the longitude problem. Livermore et al (1984) tested this approach for the past 90 m.y. and showed that the differences in latitude inferred from the palaeomagnetic and the hotspot reference frames are less than 7°. This discrepancy progressively increases to nearly 20° for the past 90-180 m.y.

but this is possibly due to the fact that the Jurassic and Cretaceous hotspot frames are not well known. Also hot-spot frames are not static e.g. those of the mid-Atlantic and mid-Indian Oceans must move in response to plate tectonic motions if they are to maintain a mid-oceanic position. However Zonenshain et al (1985) admitted that the accuracy of the reconstructions produced by this approach is hardly very high i.e. approximately 20% for the amount and direction of motion in the Late Palaeozoic and Early Mesozoic times. The accuracy decreases to no less than 30% in older reconstructions i.e. Lower Palaeozoic times. However the first reconstructions of absolute plate motions using hot spot trajectories in Cenozoic and Late Mesozoic times showed good coincidence with plate kinematic and palaeomagnetic data. This is to be expected as the argument is largely circular in that these three aspects are not mutually exclusive. It has been suggested (Zonenshain et al, 1985) that the Euler pole and hot spot methods could be combined in cases of incomplete data to find the best solution.

It is accepted that the longitude problem is not critical to this research the results of which are only concerned with the palaeolatitudes derived from the reconstructions. However it is advisable to be aware of any potential errors or discrepancies which may affect the positioning of continental fragments in palaeogeographic reassemblies.

7.1.2 (b) The Influence of Palaeoclimatic Evidence on Reconstructions.

The use of palaeoclimatic indicators to assist in the development of palaeoreconstructions has become increasingly important e.g. Robinson (1973); Drewry et al (1974) and Ziegler et al (1977). Robinson (1973) highlighted the effect of precipitation on sedimentary rocks and contrasted climatic gradients derived from these observations with temperature - this also has some influence on the deposition of certain sedimentary rocks. The occurrence of thick clastic sequences, coal swamps and glacial tillites in the rock record are thought to represent the effects of high precipitation rates, while dry climates are indicated by the occurrence of evaporites and desert sands. There is a negative association of precipitation and the formation of carbonates: water temperature seems to be the overwhelming factor in carbonate formation (Ziegler et al, 1979).

The notion that there is a direct relationship between palaeoclimatic conditions and certain lithologies has already has been mentioned and discussed in Chapter Four (see section 4.6). The discussion is not repeated here other than to emphasize the assumptions inherent in accepting that such a relationship exists i.e. that present precipitation patterns, circulation patterns and the limits of the resultant climatic regimes were the same in the past as at present. However it must be mentioned that the argument involved in the association of climate with certain lithologies is largely circular. The palaeodistributions of such climate-sensitive lithologies as evaporites and red beds have been determined using palaeomagnetic studies which have themselves been tested using palaeoclimatic parameters. In order to avoid incorrect interpretation of the presence of such lithologies, palaeoclimatic evidence must only be used in conjuction with other geological evidence (i.e. biogeography, tectonic constraints) to produce palaeoreconstructions. Therefore palaeoclimatic evidence is only used to support conclusions drawn from these other geological sources (see sections 7.1.2 a, c, d).

One example of the use (and supposed success) of the determination of continental orientations using palaeoclimatic, in addition to palaeomagnetic data, is that for the Silurian period (Ziegler et al, 1977). It was considered that the influence of climate on lithology would be particularly marked during periods when epeiric seas were widespread such as during the Silurian. Sedimentation would be dominantly autochthonous and so reflect the climate at the depositional site (e.g. evaporites, reefs, carbonates). They argued that long river systems could not develop during such times as few large land areas were found in low latitudes so climatic patterns would be more zonal than cellular. Hence the transport of clastic sediments from wet to dry belts would be precluded and even allochthonous deposits, especially thick sequences of coarse clastics, would be useful as palaeoclimatic indicators. Ziegler et al considered Silurian northern hemisphere atmospheric circulation to have been similar to present southern hemisphere patterns because of the lack of significant land influence on climate. The climatic zonal pattern (e.g. the hot-wet zone from 10°N to 10°S) was confirmed by Silurian sediment distribution on the palaeocontinents (e.g. Laurentia, Baltica, Siberia) whose orientations had previously been established from palaeomagnetic measurements.

Although Tarling (1985c) accepted that palaeoclimatic evidence is probably the second most reliable factor in forming reassemblies (after palaeomagnetism) he cautioned that it was not advisable to make too simple assessments. For example, local palaeogeography can have a drastic effect upon the latitudinal extent of any of the palaeoclimatic indicators e.g. the accumulation of a 'thick limestone reef' would be prevented if there was a high argillaceous discharge in the vicinity despite the presence of satisfactory climatic conditions for reef development. Tarling also reiterated the fear that the present may not be the ideal key to the past with regard to climatic latitude belts. This aspect of the use of palaeoclimatic indicators in the formation of reassemblies was also noted by Boucot and Gray (1983). They cautioned that climatic features of the Cenozoic and Mesozoic may depart substantially from pure association with latitude in many places and that, by analogy, they may also have done so in the Palaeozoic. However Boucot and Gray also claimed that the use of all available climate-correlated criteria in establishing Palaeozoic geography would provide estimates about latitude with which to independently check the information provided by remanent magnetisation. Boucot (1985) did use palaeoclimatic indicators to reject the late Silurian palaeogeographic maps of Bambach et al (1980) and Smith et al (1981) as the presence of major evaporitic bodies in equatorial regions was considered to be unlikely.

In conclusion palaeoclimatic information must be interpreted very cautiously when reconstructions based on palaeomagnetism are evaluated. The value of reconstructions based upon lithofacies data on top of other considerations (Ziegler et al, 1977) must also be in question. However all aspects of geological information (including that inferred from palaeoclimate-sensitive lithologies) must be resolved before satisfactory reconstructions can be produced. Another circular argument must be emphasised at this stage. In many cases one major test for the geocentric axial dipole model for the Earth's magnetic field upon which palaeomagnetic studies are based is on the distribution of palaeoclimatic-sensitive lithologies.

7.1.2 (c) The use of Palaeobiogeography in the making of Reassemblies.

Palaeobiogeographical criteria provide information about marine and terrestrial regions which maintained reproductive communication because of similarities of their biota i.e. it has the objective of defining palaeobiogeographical provinces that have specific spatial and temporal constraints. For the marine environment the reproductive communication most commonly depends on surface current circulation which is directly related to the mean wind circulation patterns. These are themselves dependent upon the relative positions of continental landmasses and proportions of continents to oceans. For the continental environment similarities of biota commonly imply the availability of both floral and faunal migration routes.

Therefore the global fauna and flora of the past and the present are not distributed in a random manner and probably have never been uniformly and homogeneously distributed. There are two orders of biogeographical controls the first being north-south correlated: temperature, light and other physical variables are related to the motion of the Earth relative to the Sun, namely seasonality and latitude. Organisms with different tolerances to these variables are latitudinally organized. Second order distribution patterns are longitudinally distinct groupings of organisms that are reproductively isolated by physical barriers e.g. ocean currents, land and salinity barriers (Boucot and Gray, 1983). Hence a study of first and second order biogeographical data should aid in the construction of reassemblies. Conversely palaeomagnetic constraints can assist in the distinctions between these controls.

The use of biogeography was introduced relatively early in the study of palaeogeographical reconstructions. Examination of Permian fauna (Vine, 1973; Waterhouse and Bonham-Carter, 1975) indicated that China, with its relatively diverse fauna was probably lower in latitude than shown in previous Pangaean reconstructions and not part of Eurasia until the Mesozoic. Much later Boucot (1985) used biogeographical information to question maps of the late Siluriam produced by Bambach et al (1980) and Smith et al (1981). He discussed both reassemblies on the grounds that surface current circulation patterns consistent with the postulated palaeogeography were unable to explain the high level of endemism postulated for that time.

The palaeontological criteria can be extremely informative, not only in terms of establishing continental positions, but also for an understanding of the nature of the organism being considered. Turner and Tarling (1982) considered the well-known distributions of the Siluro-Devonian agnathans, particularly the thelodonts, in relation to the available palaeomagnetic data. They concluded that the evidence appeared to be more consistent with these having a dispersal pattern that required land connections i.e. they lived predominantly, or even entirely, in fresh or brackish water at low latitudes and were unable to cross ocean barriers. A problem of using biogeographical, and indeed other geological evidence, is that such assessments as those made above can be largely self-fulfilling when considering areas for which no reliable palaeomagnetic data are available. This emphasizes the importance of using evidence from different geological disciplines, such as biogeography, in conjuction with palaeomagnetic data. Another limitation of the use of palaeontological criteria is that few authors have examined more than one taxonomic group for more than one or two periods (Ziegler et al, 1979) and most of the early studies recognize only three or four provinces at any one time (Middlemass et al, 1971; Hallam, 1973; Hughes, 1973; Ross, 1974).

The remarkable distributional diversities of flora and fauna also cause severe difficulties in the interpretation of biogeographical evidence e.g. the present Indo-Pacific Province spans about 180° of longitude. Conversely sharp environmental gradients (e.g. those along shelf margins) may persist in the same region for a long period of time so causing faunal changes that give no indication of the scale of geographical separation which occurred (Ziegler et al, 1979). Despite these problems biogeographical patterns have proven useful in determining the east-to-west order of the continents in the same latitudinal belt e.g. Bambach et al (1980) determined longitudinal separations between continents by integrating biogeographical relationships and plate motion constraints. They considered that the relative sequence of the continents around the equator in the Late Cambrian was clear from biogeographical evidence. Hence space constraints alone were used to fix the longitudinal positions of continents within rather narrow limits.

So the distribution of faunal provinces provides a useful check on the latitudes predicted by palaeomagnetism and can also put some constraints upon the longitudinal separation of continents. It is well known that biogeographical barriers are due to both climate and geographic distance but the interpretation remains difficult. An important need is the development of a set of biological criteria for recognizing the major climatic zones as it is difficult to differentiate between tropical, subtropical, temperate and polar fauna and flora. Ziegler et al (1981b) reduced the problem by basing climatic 'assignments' upon latitude and the orientation of coastlines and landmasses rather than on direct evidence from the fossil record. However they did also refer to obvious features such as distribution of coral reefs, occurrence of diversity gradients and presence of seasonal growth rings in trees to confirm their climatic assignments. Ziegler et al also cautioned that another factor which must be considered is the influence of east-west asymmetry in ocean current and temperature regimes.

7.1.2.(d) Tectonic constraints upon Palaeogeographic Reassemblies.

The definition of criteria for recognizing ancient continental boundaries (e.g. Dewey and Bird, 1970; Burke et al, 1977) enabled the use of varied tectonic lineaments in addition to palaeomagnetic data in making Palaeozoic palaeogeographical reconstructions (e.g. Morel and Irving, 1978; Scotese et al, 1979; Bambach et al, 1980). Tectonic and stratigraphic information from outcrop and subsurface data can be used to suggest former proximity of now separated continental fragments and also to provide information on the timing of tectonic events.

The occurrence of such features as mountain belts with strongly folded rocks and ophiolite belts indicates which previously separated regions were brought together by plate tectonic processes, under the assumption that such features mark major separation of the sutured terrains (Boucot and Gray, 1983). Where such belts cross a continent (e.g. the Ural Mountains in Eurasia) two former continents appear to have collided and been sutured (Bambach et al, 1980). Unfortunately the presence of these belts does not give an indication of the scale of the original separation of the continents. This latter may be hinted at by an evaluation of K/Na volcanics in the area to give an idea of the amount of subduction which has occurred. The rifted margins of once-associated continents are represented by other geological features, especially belts of basaltic igneous rocks associated with elongate basins bordered by normal faults. Ziegler et al (1979) assumed that subduction zones of the past would have had continuity from continent to continent. All the major Palaeozoic continents have had active andesitic volcanic chains along at least one of their margins as they have been defined that way. So the reconstructions were arranged on the assumption that most of the compressive margins were in continuous belts. They showed it was possible to do this within the constraints provided by both palaeomagnetic and biogeographic data.

Tectonic constraints clearly limit the areal extent over which individual palaeomagnetic data can be extrapolated as shown by the example cited previously (Turner and Tarling, 1982). One of the solutions to explain the distribution of Siluro-Devonian agnathans would be for a separation of the Gondwanan continents into east-west components but then there must be evidence for these two parts having become joined in the late Palaeozoic. There is no evidence for such a suture therefore this reconstruction would not be realistic and can be discarded (Tarling, 1985c).

7.1.3. Problems with Dating.

Possibly the greatest source of error in this research is the determination of the ages of the mineral deposits, the palaeoclimatic indicators and the palaeomagnetic remanences. It is difficult to quantify the total effect of these errors, as with most aspects of this research such an evaluation would involve a large degree of subjectivity. A brief summary of radiometric dating techniques is given as an introduction to this section, other methods are briefly discussed when they are mentioned in the text.

For the ^{87}Rb to ^{87}Sr decay scheme it is extremely important when measuring that all the samples had the same initial ratio at the time (t) that is being determined. Experiments show that this condition is most frequently met by igneous rocks crystallizing from the same magma. It is sometimes true for metamorphic rocks if the metamorphism was sufficiently pervasive to homogenize the $^{87}Sr/^{86}Sr$ ratios, but it is much less likely to occur among sediments with variable amounts of detritus and cements from different sources.

Serious errors in the interpretation of results will arise if at any stage since time (t) the trace element ratio changes by the introduction or removal of elements from the system. Provided that the samples are fresh and unaltered such an assumption appears to be justified for Rb/Sr and Sm/Nd decay schemes (Smith, 1981). Uranium, however, is readily oxidized to a highly mobile state so that the U/Pb ratios measured for igneous rocks are rarely the same as those acquired during crystallization. This problem may be avoided by combining equations for the two U/Pb decay schemes, forming a relationship between $^{206}Pb/^{204}Pb$ and $^{207}Pb/^{204}Pb$ in which the uranium contents need not be known. The initial presence of 'daughter' elements of non-radiogenic origin may also give rise to unreliable results.

The more important radioactive decay schemes now used for dating purposes are:

1) The U/Pb decay scheme usually involves the use of zircon which may contain minute traces of uranium and the amount of primary lead (i.e. the amount of non-radiogenic lead) is very low and the mineral itself is highly resistant, preventing the leaching of uranium.

2) The Pb/Pb scheme is reasonably reliable if the material used was initially free from ^{207}Pb and ^{206}Pb (as ^{207}Pb accumulates approximately six times as fast as ^{206}Pb). Non-radiogenic lead (^{204}Pb) is always found associated in lead ores with ^{206}Pb and ^{207}Pb.

3) The $^{40}K/^{40}Ar$ system is difficult to use because of the abundance, under normal circumstances, of non-radiogenic argon. Good results are produced even though argon is a gas which might be expected to escape from the system. It is possible to use minerals which contain only small quantities of potassium. This method has been tried for glauconte dating as attempts to date such authigenic minerals in sediments have yielded inconsistent results. The Rb/Sr technique is mainly used for Precambrian rocks as the half-life is notably long. It is a particularly valuable method for metamorphic rocks. Although rubidium and strontium are not abundant elements, contamination with primary strontium or loss of rubidium or strontium from the system are both rather unlikely occurrences, and so this method is fairly reliable.

7.1.3 (a). Dating of Mineral Deposits.

The ways in which rocks acquire a magnetisation can be used to determine the thermochemical history of a mineral deposit and may assist in the determination of both absolute and relative dating. Most mineral deposits either contain ferromagnetic minerals (e.g. pyrrhotite (Hanus and Krs, 1968), magnetite or haematite) as major constituents or they contain them in significant quantities as accessory minerals (e.g. chromite ($FeCr_2O_4$), Kropacek and Krs, 1968); the tin mineral cassiterite which has varying amounts of iron bound into its lattice (Hanus and Krs, 1965). A study of the origin of the magnetisation of these minerals can be used to investigate the physical conditions under which the ores were deposited. If the geomagnetic field changes are reasonably well documented for other rocks in the region, it should be possible to obtain relative or absolute dates for the mineral deposit by direct comparison of their direction of remanence with those rocks of known ages (Tarling, 1974). For example Evans and Evans (1977) used remanences in haematite to date mineralisation in the Mendip ore field.

A common application of the palaeomagnetic method is in determining whether a mineralisation is syngentic or epigenetic. Thus if the remanence directions of the ore and its host differ, the ore was formed epigenetically. But if the directions coincide either the ore and host remanences were formed syngenetically or both were subsequently remagnetised. The latter could be detected by testing whether the magnetisations were primary or secondary. This method may also be applicable to LSBM deposits where radiometric ages based on galena may be misleading (Park, 1983). Early studies of haematite bodies in the Lake Superior region (Symons, 1967a and b) have shown that some ore bodies have directions similar to intrusive rocks and that these remanence directions are probably residual weathering products. The accuracy of this method of absolute dating is very difficult to evaluate as precise pre-Permian polar wander curves have not yet been determined for most continents. Hence these dates are subject to the uncertainties inherent in palaeomagnetic methods, most of which have been previously discussed.

A specific example of the application of palaeomagnetic methods to determine the age of mineral deposits is the much cited work of Beales et al (1974) on Mississippi Valley-type (LSBM) ore deposits. Four deposits were sampled and the direction and intensity of the NRM were measured. Only two of the samples had measurable, although very weak, NRM in both host and ore deposits, which proved to be highly stable upon AF demagnetisation. The results showed that the pole positions for the ores and their corresponding hosts were identical within the statistical uncertainty, strongly suggesting that the ore and the host were of roughly the same age. However later work (Wu and Beales, 1981) suggested that the late Cambrian age for one deposit had suffered considerable post-depositional re-setting and/or overprinting. Using the palaeolatitudinal control upon the formation of LSBM deposits suggested in Chapter Six it may be possible to resolve the dilemma of the age of these deposits. From Figures 6.15 g (300 m.y. reconstruction) and 6.15 j

(450 m.y. reconstruction) the palaeolatitude for the younger age of the deposits appears to be the more valid. It is more in keeping with the general low palaeolatitudinal range of LSBMs given in Chapter Six, section 6.2.1. than the palaeolatitude derived from the older suggested age (i.e. the Ordovician). Hence the results here support the conclusion of Beales et al (1980) and Wu and Beales (1981) that the mineralisation for LSBMs from the Viburnum Trend, southeastern Missouri is dated about 300 m.y.

The relative ages of mineralisation may be revealed by the conglomerate test which can indicate whether the mineralisation associated with the ore is primary or secondary (Park, 1983). Mineralisation is often associated with conglomerates or breccias because of their permeability to ore-bearing fluids. The test has been applied to the copper mineralisation of the Copper Harbour Conglomerate, Michigan (Palmer er al, 1981). The results showed the mineralisation was pre-folding, possibly produced during the formation of secondary minerals and the acquisition of a secondary, chemical remanent magnetisation.

An alternative method of dating mineral deposits would be to use the radiometric techniques previously outlined, but these have limited application. The problem with many mineral deposits (e.g. some LSBMs) is that the leads in such deposits are normally of complex origin and they cannot be readily dated isotopically and so are unable to define the age of mineralisation (Sangster, 1976a). Several examples of the dating of mineral deposits using various radiometric techniques follow with comments on the solutions to the problems inherent in each method.

1) Richards et al (1985) studied the Pb-Cu-Zn mineralisation in the Northampton Block, Western Australia using both Rb/Sr and Pb isotope techniques. It was hoped that the Rb/Sr method on hydrothermally altered dolerites would date the sulphide mineralisation and that the Pb model ages on galenas would provide independent dating of the mineralisation. However

only the first of these aims was fulfilled although the Pb isotope data, in addition to the Rb/Sr data, enabled deductions as to the U, Th and Pb abundance changes during regional metamorphism and a possible source of the lead.

2) K/Ar dating of clays has been shown to be a powerful method for constraining the ages of some SDEX ore deposits, although many of the age data did not precisely define times of ore deposition (Halliday and Mitchell, 1983). K/Ar ages of clay concentrates from samples associated with SDEX mineral deposits in Ireland indicated that most, if not all, of the major ore deposits were formed during Carboniferous times e.g. Tynagh, Silvermines, Tara, Gortdrum. However in some areas Halliday and Mitchell considered that hydrothermal activity occurred during the Triassic and possibly the Permian. They also suggested that previous models which invoked major mineralisations during Mesozoic or Tertiary times were rendered invalid. The use of fine clays for dating necessitates cautious interpretation because the K/Ar age can be lowered relatively easily by tectonic and hydrothermal processes, although the clays should not lose argon spontaneously (Halliday, 1977). Hence the K/Ar ages should be regarded as minima for the timing of ore deposition. It was considered by Halliday and Mitchell in their study of Irish SDEXs that the ages might be erroneously high only where insufficient heat, stress or chemical modification had occurred to cause argon degassing from the pre-existing components. However a number of the studies have indicated that an excess of argon should not be a problem in dating clays (e.g. Mitchell and Halliday, 1976).

3) The U and Pb isotopes of samples from the Deilmann orebody, Key Lake deposit, Canada were measured by Trocki et al (1984). They proposed a two stage model of U and Pb evolution which precluded the hypothesis previously held that the 1350 m.y. event was a redistribution of U mineralisation that was emplaced during the Hudsonian orogeny. Their data showed that Pb loss (or

U gain) occurred in high grade rocks and the opposite was true for low grade rocks. Some of the results obtained provided evidence of preferential movement of ^{238}U intermediate daughters out of uranium minerals and into the matrix.

The movement of isotopes into and out of systems was also addressed by Santos and Ludwig (1983) in analyses of U/Pb isotopes in samples of ore from Highland Mine, Powder River Basin, Wyoming. They concluded that the greater discordance and lower apparent ages from the two coffinite samples were consistent with the greater daughter isotope leakage generally associated with colloform pitchblende and coffinite compared to whole rocks (Ludwig, 1979). The U/Pb apparent ages of the whole rocks should be regarded as minimum ages because of a probable loss of lead as well as radioactive daughters from the ores but Santos and Ludwig deemed it unlikely that the true ages of the deposits could be older by more than a factor of two greater than the $^{207}Pb/^{235}U$ apparent ages (3 m.y.).

It is clear that in more recent times radiometric dating methods are frequently used to determine the age of mineralisation of some mineral deposit types. There is a degree of uncertainty in using these methods such as the deduction of the isotopic composition for the initial Pb of samples which are resolved in individual ways e.g. Santos and Ludwig (1983) chose their initial Pb isotope from analyses on barren sandstones from stratigraphically equivalent rocks of the Wind River Formation. Also the ages which result have to be evaluated in the light of whether a sample has lost isotope elements or if they are representative of an older time of mineralisation.

More precise dating is possible using reversals of the polarity of the geomagnetic field. It is possible to match polarity sequences, or if the approximate age is already known, the presence of reversals may allow more precise dating than either normal palaeomagnetic or radiometric methods. Very high precision dating is possible through most of the Cenozoic as the polarity sequence is well documented from oceanic magnetic anomaly patterns (Tarling, 1974).

There is an aspect of the dating of deposits which is more specific to this research than the other, more general, sources of error which have been outlined here. Some of the deposits have been rotated through more than one set of reconstructions e.g. Meggen SDEX deposit, Germany which is dated as 375 m.y. This deposit has been rotated using the values for palaeogeographic reconstructions for 400 m.y. and 350 m.y. which are either 25 m.y. too old or too young. Obviously the rate of continental plate motion has a bearing upon the distance a continental fragment can move in a given period of time (see Chapter Six, section 6.3.3), but such a discrepancy in the ages of reconstructions available for the rotation of deposits such as this must introduce great errors into the palaeolatitudes which result. The direction of continental movement is also important as only north-south movement can be detected with some degree of accuracy using palaeomagnetic studies. To quantify the absolute east-west movement of a continent is very difficult because of the palaeolongitude problems outlined in section 7.1.2 a. Hence discrepancies derived from the palaeomagnetic data are proportional to the relative plate motion vectors.

Related to the dating problem is the question of the duration of the formation of an ore deposit. In this case, if the deposit took less than 25 m.y. to form, then to use rotation figures 25 m.y. older or younger is inappropriate. Lastly, if a deposit took 25 m.y. to form then it is a considerable period of time over which conditions (physical, chemical, climatic) must be assumed to have remained stable.

The scatter graphs (Figures 7.1 - 7.6) have been drawn to illustrate the variation between some mineral deposit types with regard to the precision of the determination of their ages of mineralisation. It is difficult to evaluate how much errors in age determinations affect the total results and one possible way to avoid these errors would have been to weight all the mineral deposit examples according to their age reliability. However when the cumulative effect of all the other errors involved in this study is considered such a weighting seems unnecessary. It would also have involved, once again, a large element of subjectivity which should be avoided whenever possible.

When there is doubt in the age of mineralisation for a particular deposit and a palaeolatitudinal control of some type upon the deposit type distribution has been proposed, it may be possible to suggest a more appropriate age for the deposit from the tie lines on the scatter graphs. For example, consider the Bulgarian and Polish LSBM deposits (tie lines e, f, h) in Figure 7.1. Perhaps these deposits should have an age close to 250 m.y. rather than being dated near 200 m.y. as at present. Their palaeolatitudes would then be more in keeping with the majority of the LSBMs. The palaeolatitudes produced for the Polish deposits (tie line h) seem to be questionable for the continental fragment appears to have moved through a considerable latitudinal distance in approximately 50 m.y. This is unlikely in view of the figures given in the discussion previously (see Chapter Six, section 6.3.3). The Bolivian SHBM deposits (Figure 7.4) have an age range of 150 m.y. so no firm conclusions can be drawn from their palaeolatitudes as regards general SHBM distribution. However it appears from the graph that the oldest age is most appropriate to enable these deposits to conform with the latitudinal distribution of the other SHBMs. From the SDEX scatter graph (Figure 7.5) only the Mae Sod deposit of Thailand appears worthy of comment in that the deposit has a very poorly defined mineralisation age range of 100 m.y. It is difficult to draw any firm conclusions about the Mae Sod deposit as its palaeolatitudes all lie within the range for the majority of SDEXs regardless of age. Hence the most likely age of the mineralisation cannot be

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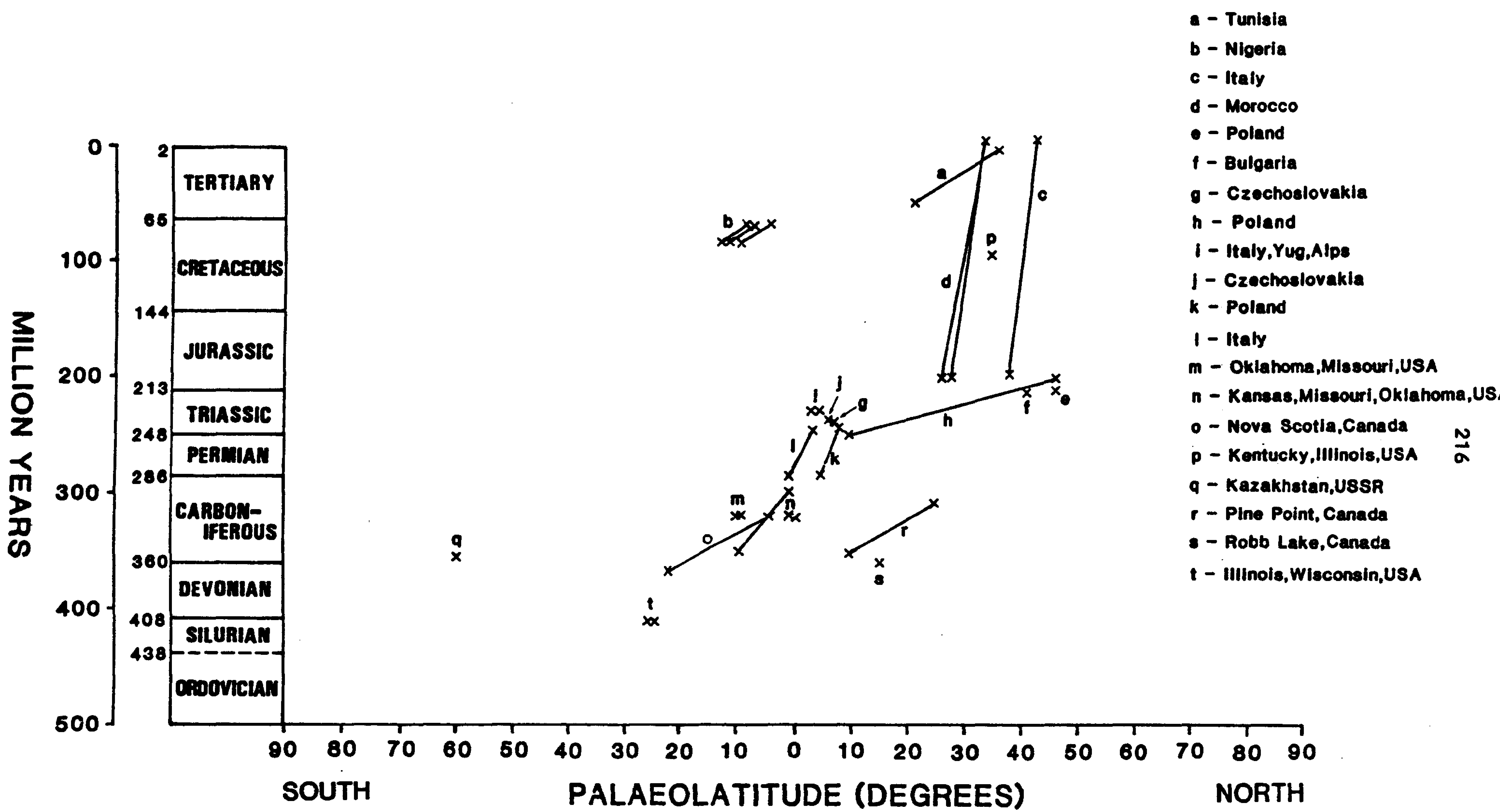


Figure 7.1. Palaeolatitude vs. Age for LSBM deposits from Tarling reconstructions. Tie lines join minimum- and maximum-age derived palaeolatitudes of those deposits with poorly defined ages of mineralisation.

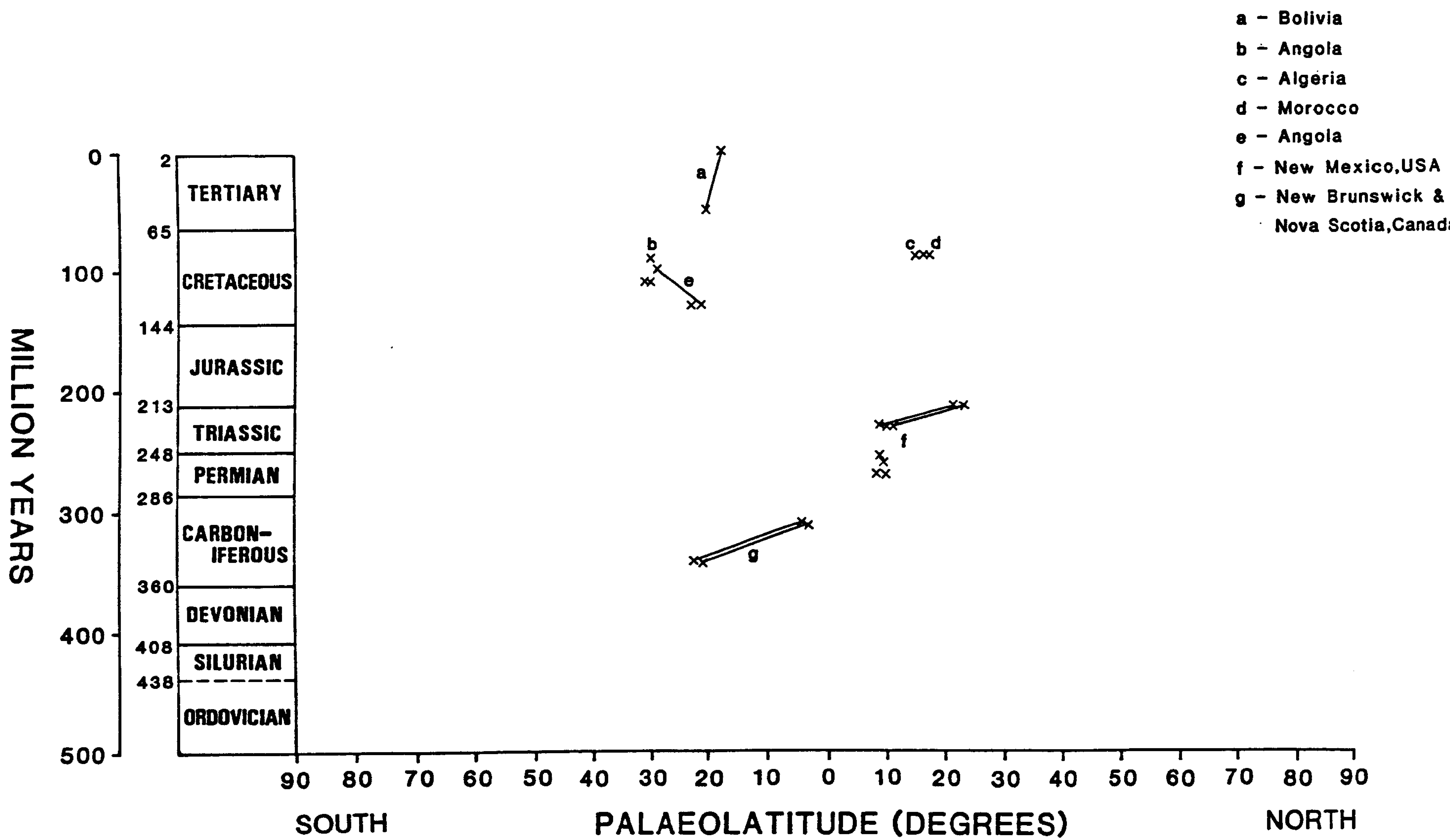


Figure 7.2. Palaeolatitude vs. Age for SSCU deposits from Tarling reconstructions. For figure legend see Figure 7.1.

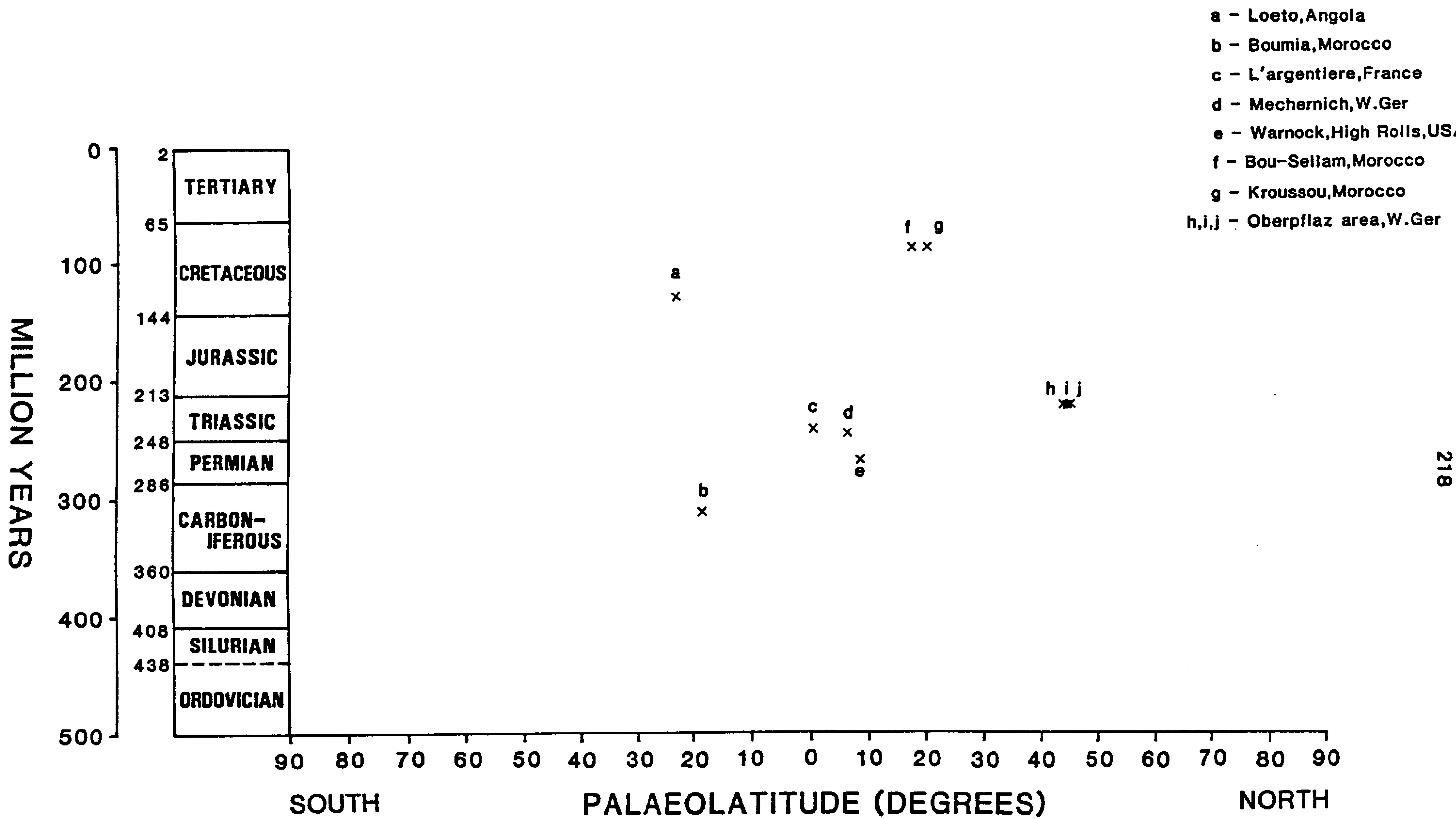


Figure 7.3. Palaeolatitude vs. Age for SSPB deposits from Tarling reconstructions.
For figure legend see Figure 7.1.

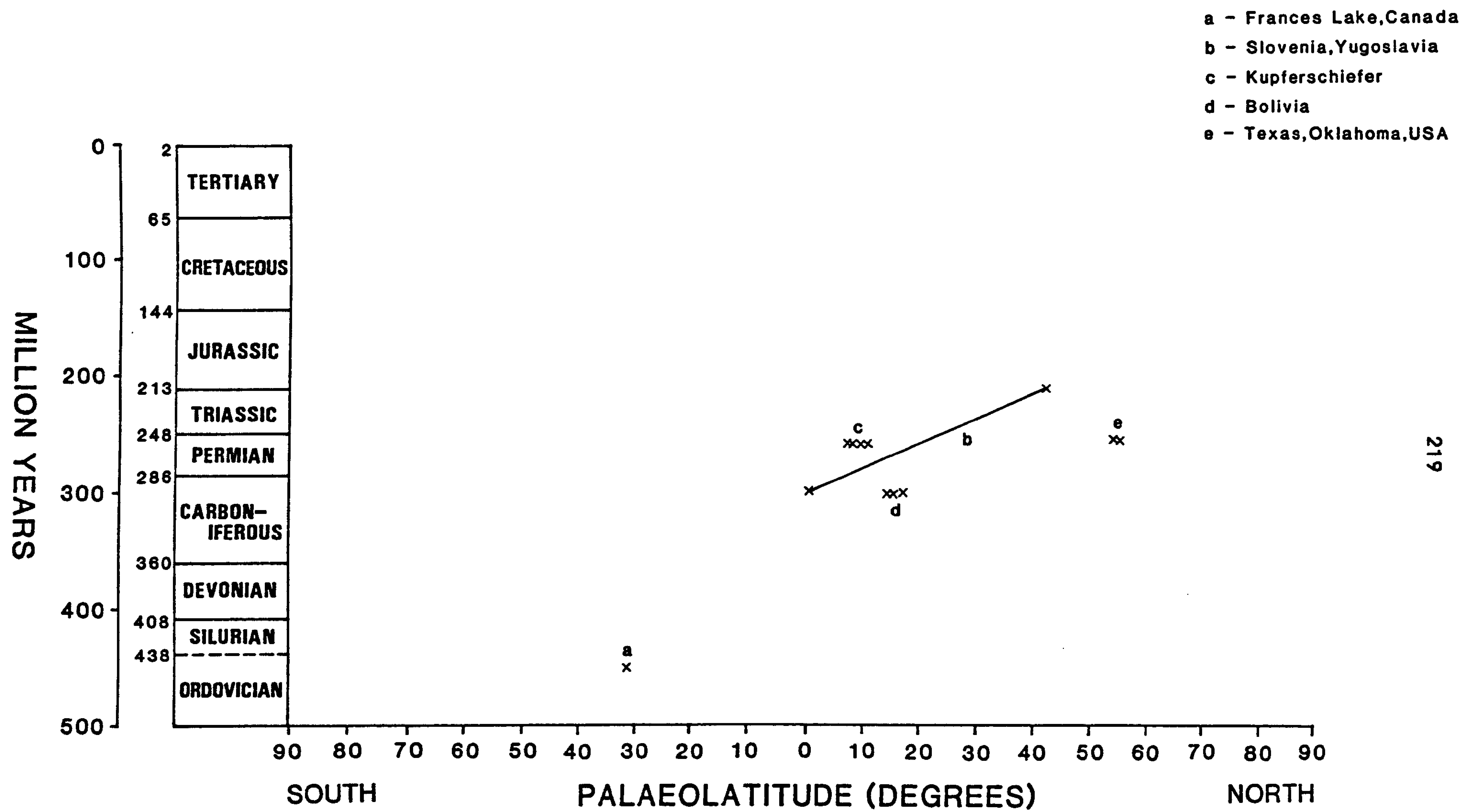


Figure 7.4. Palaeolatitude vs. Age for SHBM deposits from Tarling reconstructions. For figure legend see Figure 7.1.

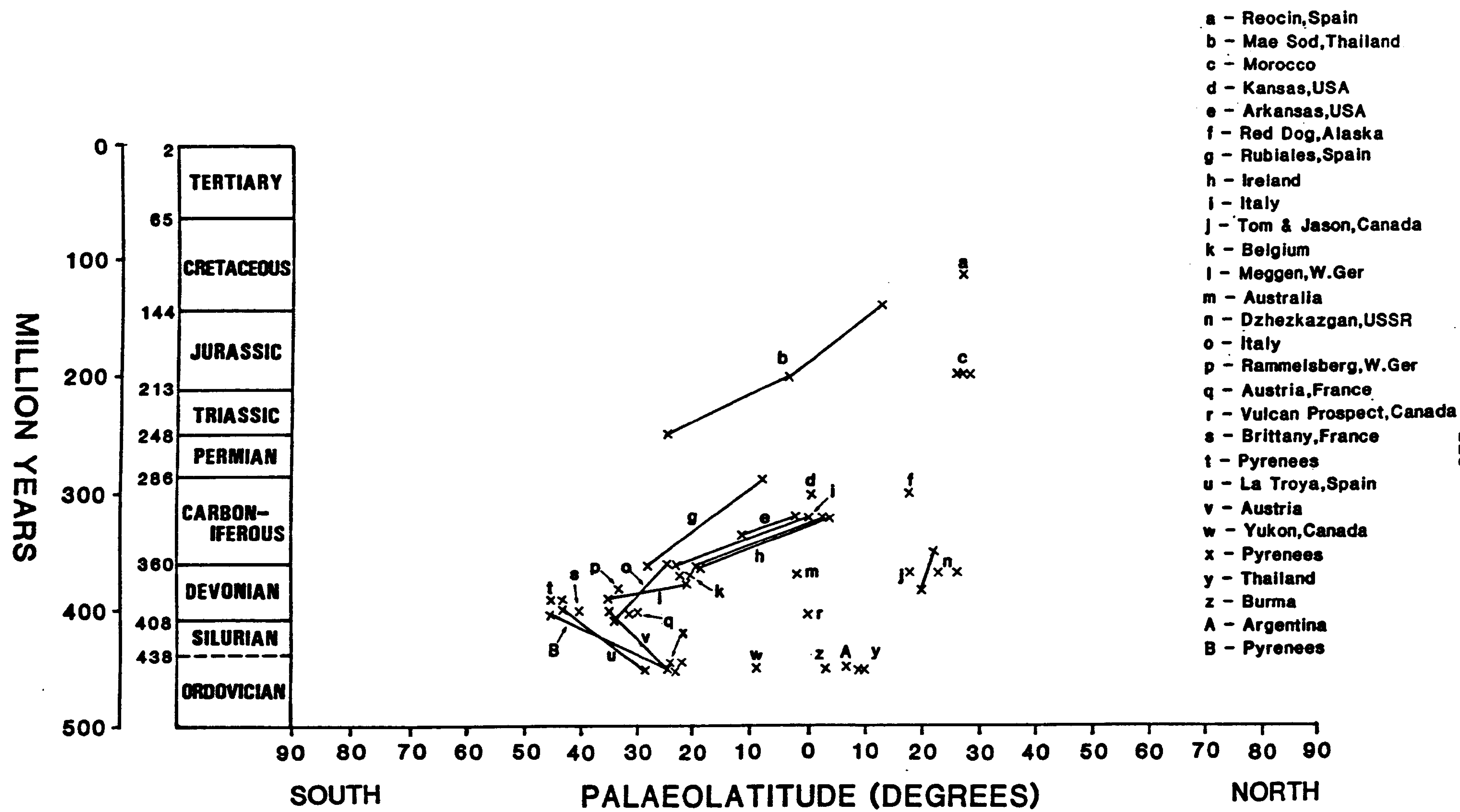


Figure 7.5. Palaeolatitude vs. Age for SDEX deposits from Tarling reconstructions. For figure legend see Figure 7.1.

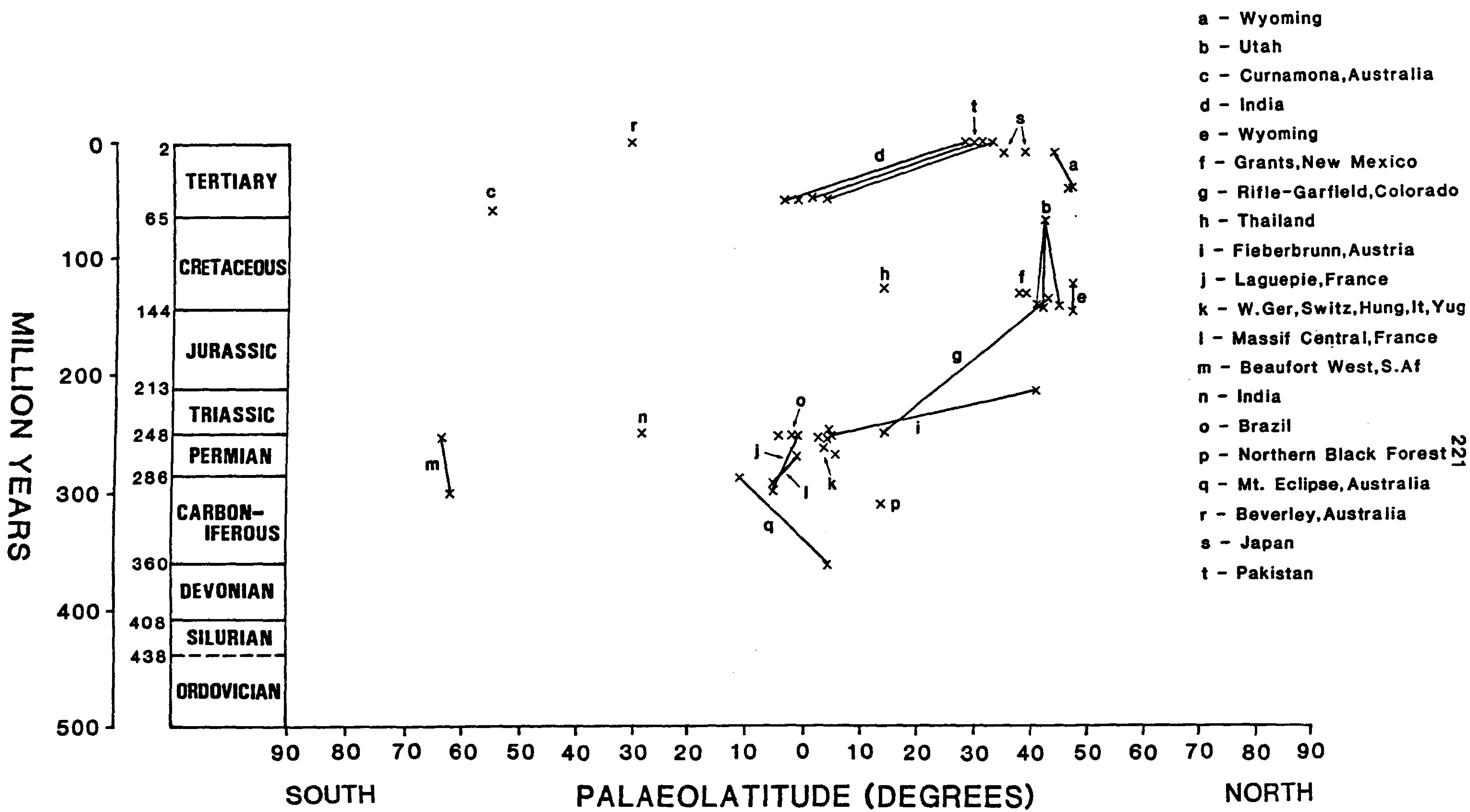


Figure 7.6. Palaeolatitude vs. Age for SSUV deposits from Tarling reconstructions.
For figure legend see Figure 7.1.

determined using the most appropriate palaeolatitude i.e. one in the typical range of SDEXs.

Half of the SSUV deposits (Figure 7.6) show a large age range but the tie lines cannot be used to assist in determining which of the ages are more likely to be the most appropriate. SSUVs are concentrated in both the equatorial and temperate rainfall belts but the age ranges encompass both of these zones (e.g. tie lines d, g, i). So it is not possible to select which age is likely to be the more accurate on the basis of palaeolatitude for SSUV deposits. The SSCU deposits appear to be quite precisely dated (Figure 7.2) and those examples with less precise ages of mineralisation have narrow ranges relative to other deposit examples e.g. LSBM and SSUV. The SSPB deposits (Figure 7.3) are all very precisely dated.

7.1.3 (b) Problems of Dating in Palaeomagnetic Methods.

The palaeomagnetic methods have been described (Chapter Two) and possible errors which they may introduce into the results have been discussed earlier in this chapter (see section 7.1.1). It is now necessary to evaluate the accuracy and reliability of the dating of the remanences and whether or not these correspond to the age of the rock under examination.

The importance of the identification and study of primary,as opposed to secondary, remanences in the palaeomagnetic techniques has been emphasized. Red sandstones and siltstones provide the main sources for palaeomagnetic data from sedimentary deposits but the origin of their remanence, on which any interpretation rests, is controversial (see Chapter Four, section 4.6.1). If a stable remanence has been identified in an igneous rock then a K/Ar radiometric age of the rock itself may be of use in assessing the age of the remanence. However this method assumes that the rocks examined have behaved as closed chemical systems since their formation. Rocks from which reliable K/Ar ages have been determined are unlikely to have been subjected to temperatures in excess of 300°C because this usually affects the argon retentivity (Dodson, 1973). Such rocks are therefore unlikely to have acquired a high temperature viscous component of remanence. However care must still be taken in the acceptance of K/Ar ages as chemical changes may have affected the magnetic minerals in the rock but not those minerals containing the potassium and argon. Hence it is extremely difficult to differentiate 'reliable' K/Ar dates from 'unreliable' ones (Tarling, 1983).

The date of the acquisition of remanent magnetisation in a rock of unknown age can be determined by comparison with the records of variation in the intensity and direction of the magnetisation itself. This dating may be carried out in three main ways summarized below from Tarling (1975). The rate of continental motion relative to the average geomagnetic pole is thought to have been in the order of 0.3°/ m.y. during the Phanerozoic. If an average geomagnetic pole position for rocks of unknown age can be determined with a reliablilty of +/- 5° then the average accuracy of dating should be approximately +/- 15 m.y. This method combines variations in observed relative motions from virtually nil to some 3°/ m.y. so the accuracy of the dating depends on the actual relative motions involved at each time. The use of a combination of sea-floor spreading and continental paleomagnetic studies is particularly useful at times in the Cenozoic. Occasionally discrepancies occur between the dates of events e.g. the extrusion of continental basalts in the Faroes, Greenland and Britain and corresponding sea-floor anomalies. It is not clear whether these age differences reflect inaccuracies in the technique or if these results are genuine. Lastly studies of secular variations of the Earth's magnetic field may be useful for dating purposes. Very short term geomagnetic changes are preserved in rapidly deposited material e.g. lake varves, and can be dated by conventional ^{14}C or other short-lived isotopes. These should give accurate absolute dating of sediments but are only really useful for Recent deposits.

It has been established that reliable ages for samples and remanences are absolutely necessary to reduce errors. However it is also important to determine whether the age of the magnetisation corresponds to that of the rocks themselves. This aspect of palaeomagnetism has been introduced by Irving and Strong (1984) who produced evidence against a large-scale Carboniferous strike-slip fault in the Appalachian-Caledonian orogen. They considered that the palaeolatitudinal off-set of Acadia (the eastern part of the northern Appalachians) relative to cratonic North America was almost certainly not tectonic but is purely an artefact of the wrong assumption of the equivalence of rock and magnetisation ages. They proposed that Kiaman (i.e. late Carboniferous and Permian) overprinting is widespread in Newfoundland and that observations from these secondary magnetisations agree with those from late Devonian and early Carboniferous rocks of the North American craton. This is taken as confirmation of the proposal that magnetisations of the cratonic early Carboniferous rocks are actually Kiaman, and not early Carboniferous in age. Confusion may have arisen because it has been shown (Irving and Parry, 1963) that the Kiaman palaeofield was almost continually reversed so uniform secondary magnetisations could have been acquired over a long period of time without the complications of reversals.

Three proposed sources of the secondary magnetisations are:

(i) secondary haematite was a common product (Hedley, 1968) of seasonal rainfall and a depressed water-table on the Pangaean supercontinent i.e. desert weathering,

(ii) mild heating during the Hercynian-Appalachian orogen could have converted oxyhydroxides of iron to haematite (Hodych et al, 1984),

(iii) uplift and cooling following deep burial could have caused the acquisition of viscous partial thermoremanent magnetisation (Chamalaun and Creer, 1964).

Livermore et al (1986) incorporated all of the factors mentioned here to evaluate some Pangaean configurations. They noted a clear off-set in age along common apparent polar wander paths of North America and Europe, suggesting a possible error in intercontinental correlation of up to 20 m.y. or widespread magnetisations of the North American formations after a similar period of time. Also paths for South America and Africa follow broadly similar northward routes, but with quite large divergences for some time windows e.g. 260 to 280 m.y. Livermore et al suggested that part of these differences may be attributable to dating errors for individual results and also to rapid polar wander. However as the discrepancies appear to be systematic a more fundamental problem concerning the accuracy of the stratigraphic correlations (upon which most of the assigned ages are based) was proposed. Also many poles were derived from studies of non-fossiliferous red sediments for which the mechanism of remanence acquisition remains to be clearly demonstrated. It was concluded therefore that if there were systematic errors in the correlation of the poles then the reconstructions under scrutiny could be viewed as artefacts of the numerous dating errors.

7.1.3 (c) Problems in Dating Palaeoclimatic Indicators

It has been noted that a number of mineral deposit types are closely associated with certain palaeoclimate-sensitive lithologies which may be used to date the deposits themselves. However the determination of absolute age within narrow limits is rarely possible; +/-5% of a mean value being the norm (Frakes, 1979). Relatively young deposits may be generally dated with more confidence than older Palaeozoic ones because the Cenozoic time scale is more closely subdivided. Also the precision of radiometric techniques, although still in the order of 5%, allows finer resolution in terms of absolute years.

There are a few problems with the application of palaeomagnetic techniques in dating deposits of palaeoclimatic significance e.g. red beds with a highly controversial origin of remanence. Limestones and dolostones generally have only a weak remanence and the remanence of the purer varieties has only been adequately measured during the last few years. Most of the diagenetic changes in these lithologies have associated migration of iron-bearing compounds, as evidenced by the growth of haematite and goethite rims. This means that the processes and mineralogy of such changes can be examined and possibly dated magnetically (Tarling, 1983) but restricts the interpretation of such observations at this stage of understanding. The formation of evaporites in basins is generally accompanied by some detrital grains, if only blown in by winds from the surrounding deserts. It is expected that the original remanences will be strongly distorted by the extensive recrystallization and mobility of these evaporite materials so few studies have been made on evaporites. Peat deposits usually contain detrital grains so it should be possible to date them. But sulphur is present in association with the organic matter and is combined by bacterial activity during diagenesis to form pyrite which often replaces fossil remains (Casagrande et al, 1979) so many of the original iron compounds are converted to non-magnetic forms. However changes in acidity and oxidation states may lead to the formation of magnetite or haematite during later stages of diagenesis. Therefore there is some potential for palaeomagnetic dating of such processes (Noltimier and Ellwood, 1977).

As palaeomagnetic and radiometric methods are fairly inadequate, palaeontological and other means of dating are more commonly used. In dating palaeoclimatic indicators by fossils it is particularly important to be aware that environmental changes and consequent migrations of certain forms are of great significance. Some forms serve as guides to geologically instantaneous changes in surface water temperature e.g. the Neogene of the North Atlantic and Southern Ocean. For correlation purposes, as opposed to strict dating, other environmental factors are sometimes used e.g. extent and thickness of lichen growth on Quaternary tills aids relative age and correlation with dated deposits (Frakes, 1979). The use of palaeontological data is most useful for marine sediments: however even this method has some limitations. For example extreme climates such as desert or polar regimes are not conducive to life and hence most strata of these regions in the geological record can be dated only within wide limits due to the paucity of fossil types and numbers.

The use of magnetic reversals in geochronology gives a precise method of dating as such reversals occur on a global basis and are geologically instantaneous, taking some 2000 to 8000 years to occur (Tarling, 1983). This makes a reversal a time-marker which is more precise than that of fossil extinctions which may be strongly diachronous when comparing different regions of the Earth. Also the reversal sequence for the last 7 m.y. is sufficiently well-known and well-dated to allow age determinations of a much greater resolution than fossil zonation (in some cases down to about 200,000 years). Although the reversal sequence is known in some detail back to the earliest Mesozoic, precise dating beyond about 10 m.y. is not yet possible because the lengths of polarity cycles are of the same magnitude as the accuracy of dating methods by the K/Ar technique (at 10 m.y., at least +/- 0.10 m.y.), Frakes, 1979. Reversals may be most useful to support previous findings e.g. to determine the extinction rates of certain organisms; to check radiometric methods of dating sediments (Ku et al, 1968); to check the degree of bioturbation mixing of oxygen and other isotopes (Hutson, 1980).

Oxygen isotopes provide another means of dating which has a wide application in dating palaeoclimatic indicators but, as with magnetic reversals, they are reliant upon calibration against an absolute time scale and are most useful as supportive evidence for other methods. The abundance of ^{18}O relative to ^{16}O in the ocean has changed over time as ^{16}O has a greater mobility than ^{18}O in evaporation and so is concentrated in polar snow and ice. The isotope stratigraphy has been extended back to cover the last few hundred thousand years (Shackleton and Opdyke, 1973). However there are obvious limitations with this, and other, dating methods in that the age ranges over which they are applicable are extremely narrow.

7.2 Biases

There are some biases which have been unintentionally introduced into the methods used here and hence into the results of this project. For example a decision was made that attention would not involve purely economic deposits because it would be difficult to assess which deposits were economic and those which were not as the economic state of a deposit is not fixed due to changing global trading markets. Also there are marked differences in economic criteria in different regions as mentioned with regard to PORCU deposits of Sumatra (see Chapter Nine, conclusions). However some of the references used to assemble data did have such a bias e.g. Cook (1976) excluded non-marine PHOS occurrences from his study as they were considered to be uneconomic.

7.2.1 Northern Continents

A study of the mineral deposit palaeolatitudes indicates a possible bias in that there is a concentration of mineral deposits in the northern hemisphere in comparison to the southern hemisphere. A number of reasons can be put forward to explain such an imbalance in the results. Firstly the northern continents have probably been the focus of more intense exploration, largely for economic reasons. Secondly the bias may have been a reflection of the poor quality of palaeomagnetic data for many of the southern continents, with the possible exception of Australia. This poorer quality data would influence the reliability of palaeolatitudes as it is involved in the development of apparent polar wander paths for continental fragments and also in the dating of deposits, as pointed out previously. Thirdly the distribution of Phanerozoic rocks is an important influence on the distribution of mineral deposits and this project was purely concerned with deposits younger than 550 m.y. old i.e. Phanerozoic in age. There is a large proportion of Pre-Cambrian rocks in many southern continents e.g. South American, Australian, Indian and African cratonic blocks, so an apparent lack of Phanerozoic mineral deposit examples is to be expected. This concept is supported by the lack of Phanerozoic examples from areas in the northern hemisphere such as the Baltic and Laurentian shields.

Lastly the imbalance in the palaeolatitude distributions may also be a result of a bias in the distribution of total land area. There are more continents in the northern hemisphere, if land area versus time is plotted, especially in more recent times. This factor has already been mentioned in the results (see Chapter Six, section 6.7). Table 6.2 illustrates that for the Lower Jurassic to the Present time the majority of land mass has been north of the equator. This bias is emphasized by the fact that there are fewer mineral deposit examples in the older geological periods when the majority of land mass was in the southern hemisphere. The relative paucity of deposits in these older periods may be due to selective preservation of younger deposits or the bias in data collection to northern continents mentioned earlier. Table 6.2 shows the imbalance in total land area is reflected by an imbalance in mineral deposit distribution. Generally there is agreement between the percentage of total land mass in the northern and southern hemispheres and the corresponding percentage of mineral deposits for a given geological period. The notable exception is during the Permo-Triassic period when a slight discrepancy occurs in the relative distributions of land mass and mineral deposits. However the distribution of mineral deposits during this time may be greatly influenced by the unusual climatic conditions associated with the Pangaean supercontinent (as discussed in Chapter Eight, section 8.2.2.3). The uneven distribution of land mass could also have a marked effect on palaeoclimate. The effect of strong asymmetry in the proportion of land in the northern hemisphere as compared to the southern hemisphere would probably be to draw the equatorial peak toward the land-dominated hemisphere e.g. at present the intertropical convergence zone occurs about 5° north of the equator. There is a tendency for the evaporite belts to occur closer to the equator in the early Palaeozoic than in the late Palaeozoic which parallels a general increase in continentality (Ziegler et al, 1981). However this phenomenon may be due to the decrease in the Earth's rotation which would have the effect of shifting the position of the sub-tropical high pressure cells further from the equator through time (Ziegler et al, 1979).

7.2.2 A Bias towards Low-Latitude Palaeomagnetic Poles

There is one aspect of palaeomagnetic methods which has not been mentioned with regard to this project and it has yet to be resolved. There is a profusion of palaeomagnetic poles with low latitudes i.e. they have a bias for sites which occur in the palaeo-equatorial zone or, conversely, there is an apparent aversion to high latitudes. There are a number of possible explanations for this;

(i) this may mean that contrary to the present, the ancient land masses had a greater affinity for the equatorial region,

(ii) or high latitude poles do not exist,

(iii) or they are not reported,

(iv) or they are discriminated against in the construction of polar paths.

The difficulty in finding high latitude poles could arise from one or more of the following causes as summarized from Lapointe et al (1978).

1) Inadequate (and biased) sampling. The palaeomagnetic record is far from complete so it is possible that the gaps within the record repetitiously

obscure a certain feature and introduce a bias. Only an examination of the full record can provide an answer to this question.

2) Tectonic Effect. It was assumed that over geological time there was a uniform distribution of poles over the surface of the Earth. One determination out of seven should then yield a high latitude (60°) pole, from the proportion of the Earth's surface (13%) above 60° latitude. However the pole position would only be accurately determined if the sampling area can be relocated exactly. Any error in the relocation of the land mass to its exact original location and orientation would result in a surplus of low latitude poles.

3) Interpretation of High Latitude Poles.

a) It may be that the interpretation of palaeomagnetic directions which are near to the present Earth's field direction influences the bias e.g. in some palaeomagnetic studies of North American rocks such magnetisations have been attributed to recent magnetisations. However these interpretations can be brought into question as rock surfaces in the higher latitude continents have been scoured by Pleistocene glaciers and so it is expected that Recent remagnetisations would be rare. It is therefore necessary to establish whether such high latitude poles are due to Recent magnetisations or are genuine ancient poles.

b) When expected and observed pole percentage ratios are near to 100, then a uniform distribution of poles does occur, or there is a bias in that the number of high latitude poles is too high in comparison with some other regions. It may be then that recent magnetisations have been mistaken for remagnetisations. Much remagnetisation does appear to take place at low latitudes because rock surfaces in tropical regimes are commonly covered by deep lateritic weathering profiles. Such weathering often develops remanence carriers of high thermal and alternating field stability which could be correctly associated with remagnetisations (Creer, 1968). An interpretation based on these results would lead to an obvious bias of palaeomagnetic poles to the equatorial zone. To conclude the remagnetisation of sediments in low latitudes would produce a predominance of low palaeolatitudes only. However the Permo-Triassic weathering in North America, Siberia and Europe occurred when the palaeomagnetic pole itself was in the present equatorial belt and such remagnetisations give more low latitude poles. Care must be taken to determine if such remagnetisations have occurred in the interpretation of palaeolatitudes from palaeogeographies of this era.

Conclusion

It is obvious that much of the uncertainty associated with palaeomagnetic results comes from doubts about the validity of the palaeomagnetic poles. Individual workers have produced apparent polar wander paths based upon their own unique interpretation of the poles which has resulted in a proliferation of apparent polar wander curves, many interpreting essentially the same data. The main discrepancies in palaeomagnetic results are due to;

a) a database insufficient to represent adequately the Lower Palaeozoic,

b) inadequate analysis of individual magnetisations which may have been wrongly identified as primary,

c) incorrect dating of magnetisations by various methods.

Any review of palaeomagnetic data of any age is now constrained by the quantity and diversity of published determinations which may include many observations undertaken before present modes of treatment and analysis became available. However to filter the palaeomagnetic data using the proper criteria outlined in Chapter Three would probably eliminate much of the currently available Palaeozoic observations. For the Lower Palaeozoic reconstructions palaeomagnetic data are sparse hence subjective evaluations must still form a major contribution for these times. The differences between individual interpretations depend on the extent to which other constraints (palaeoclimatic, palaeomagnetic and structural) are incorporated into the model. However it must be remembered that none of these disciplines can provide unambiguous reconstructions. For example palaeomagnetic data cannot distinguish longitudinal differences, climatically-dependent lithologies cannot distinguish between northern and southern hemispheres and biogeography is influenced by numerous factors (e.g. ocean current circulation). Obviously the incorporation of such elements further increases the degree of subjectivity involved in the formation of palaeogeographic reconstructions for time periods with a limited number of reliable palaeomagnetic observations.

Despite all these shortcomings and the large leeway for interpretation of palaeomagnetic results, palaeomagnetism has a reasonably sound theoretical basis which has been justified by Cenozoic and Mesozoic results. The latter have provided important direct evidence for the movement of continents from the construction and correlation of continental APW paths and continental movement is indirectly supported by the evidence for sea-floor spreading. Both sea-floor spreading and continental drift taken together provide strong proof of palaeomagnetism's validity through the correlation of the "magnetic stripe" pattern of the oceanic crust with the dated magnetostratigraphic record of reversals on land. In conclusion the application of palaeomagnetic methods as described to mineral deposit studies was deemed to be be valid as the reliability of the palaeogeographical reconstructions is of a sufficiently high standard to use them as a basis for research. Such an assessment is justified on the grounds of consistency. For example in this case the two sets of palaeolatitudes used (derived from Tarling and BP palaeoreconstructions) were very similar as discussed in Chapter Six, despite the different approaches in the interpretation of the palaeomagnetic data. Also the present distribution of some lithologies (e.g. evaporites) correlates well with their palaeo-distribution derived from palaeogeographic reconstructions so confirming a certain degree of reliability in the methods described. Consequently the palaeo-distributions of mineral deposit types which have been used as the basis for discussion in this study can be used with some confidence.

CHAPTER EIGHT

DISCUSSION: PART TWO

CLIMATE AND PHANEROZOIC SEDIMENT-HOSTED MINERAL DEPOSITS

8.1 The Possible Influences of Climate on Mineral Deposit Formation.

It is a conclusion of Chapter Six that there appears to be a palaeolatitudinal control upon the distribution of most sediment-hosted mineral deposit types considered here although the range of latitude over which such a control is excerised varies between deposit types. In Chapter Four (see section 4.8) it was argued that latitude has a considerable effect upon climate, influencing atmospheric and oceanic circulation patterns, temperature and precipitation. By association, the distribution of sediment-hosted mineral deposit types must also be palaeoclimatically controlled. Therefore the effects of the influence of climate upon weathering, sedimentology and the formation of mineral deposits must now be evaluated. There are a number of ways in which climate may be important to ore deposition, as given below.

8.1.1 The Influence of Climate on Weathering, Soil Formation and Sedimentology.

Weathering represents the response of minerals which were in equilibrium at a variety of depths within the lithosphere to conditions at or near the lithosphere-atmosphere interface. Here they are in contact with the atmosphere, hydrosphere and biosphere giving rise to largely irreversible changes involving an increase in volume, a decrease in density and particle size and the production of new minerals which are more stable under the interface conditions.

There are two basic types of weathering; mechanical and chemical. The former involves those processes such as disintegration and comminution of the original rock whereas chemical weathering refers to those processes involving decomposition or chemical alteration. Although one or other of these types is often related to certain environments this merely indicates their relative importance and does not imply that either is totally absent.

Climate has a considerable affect upon weathering influencing the rate of release of a given element from its parent minerals and its transport to a depositional site. It may result in an increased concentration of some metals in situ e.g. nickeliferous laterites and aluminium bauxites. Several workers have attempted to define the relationship between climate and weathering; one of the best known schemes is that of Peltier (1950). This began with the assumptions that mechanical weathering is almost entirely due to freeze-thaw activity and secondly that chemical weathering is so dependent on the presence of water that its intensity should bear a fairly simple relationship to precipitation. Peltier also contended that chemical weathering is accelerated by high temperature and dense vegetation cover and that some regions could be too hot or too cold for freeze-thaw to be fully effective. By these arguments he delimited climatic regions characterized by distinctive combinations of mechanical and chemical weathering as shown in Figure 8.1.

Chemical Weathering.

This is particularly effective under conditions of plentiful moisture, high temperature and abundant vegetation. It therefore reaches its maximum in the tropical zone. A secondary maximum is found in moist temperate latitudes,

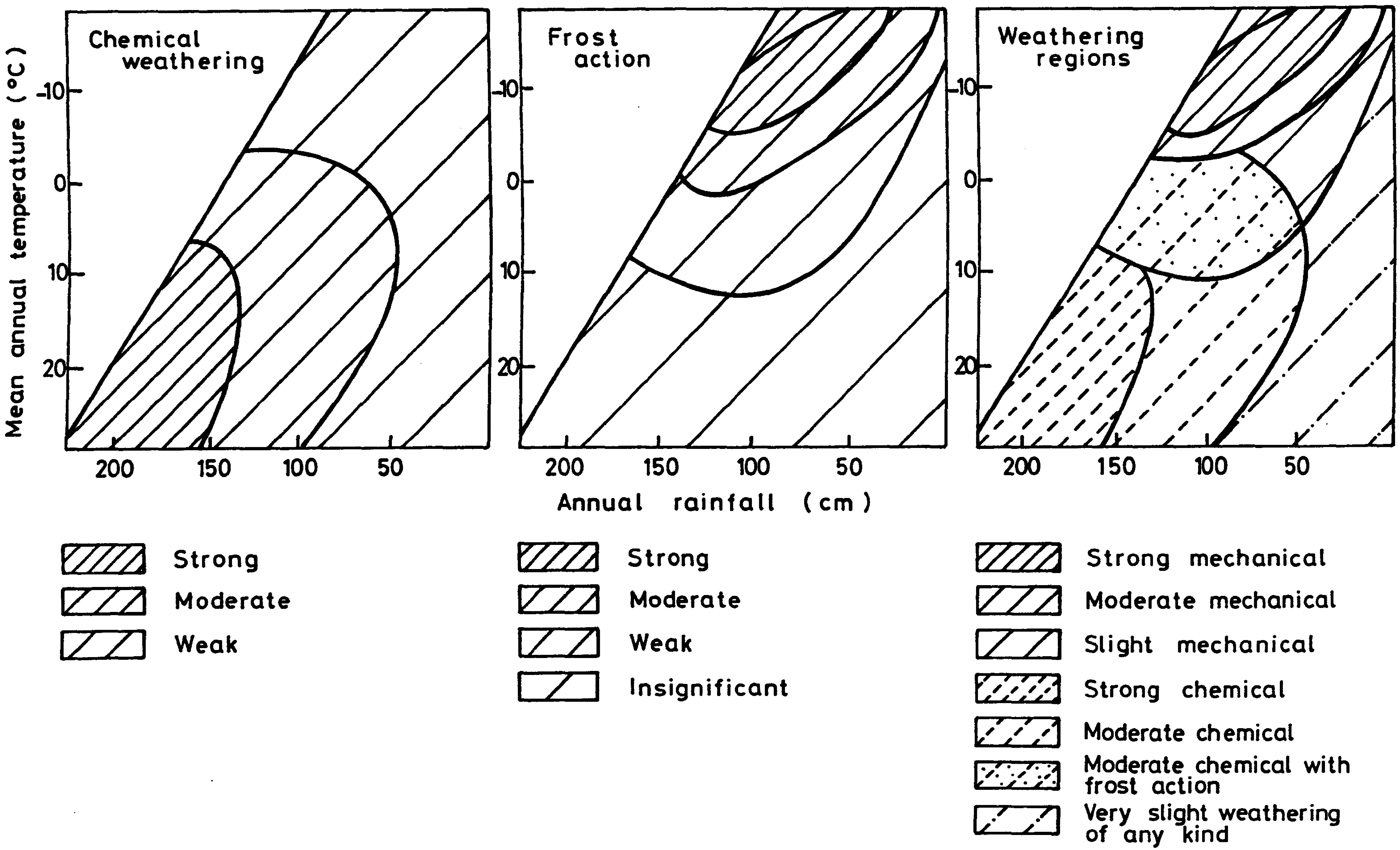


Figure 8.1. Climatic weathering regimes as defined by Peltier, 1950. (After Rice, 1977, Figure 7.4).

although the rate here is less than one-twentieth of that in the tropics (Rice, 1977).

A measure of mineral susceptibility to chemical weathering is given by the bond strengths (k cal/mol) between oxygen and the following common cations given in increased order of bond strengths: K^+, Na^+, H^+, Ca^{2+}, Mg^{2+}, Fe^{2+}, Al^{3+}, Al^{4+}, Si^{4+}. In general, minerals rich in silicon-oxygen bonds are the most resistant to chemical weathering (Chorley et al, 1984). The chemical susceptibility of clays, after weathering of the silicates, is also due to their structure of tetrahedral layers of oxygen and silicon; octahedral layers of hydroxyls, aluminiums, magnesiums etc. and of linking K^+, Na^+, Mg^{2+} and Ca^{2+} ions. Increased, continued weathering may selectively remove these ions, together with silicon and oxygen, leaving only Al^{3+}, Fe^{3+} and $(OH)^-$ in extreme conditions to yield either bauxite minerals (bohemite and gibbsite), laterites (goethite) and iron oxides (haematite and limonite). Montmorillonite and illite clay minerals have the loosest structures based on two tetrahedral layers linked to one octahedral layer. The former (especially Na-rich montmorillonite) is prone to absorption of water and expansion. However illite expands very little and is characteristic of alkaline marine conditions and environments where leaching is not excessive.

As mentioned earlier the presence of water is crucial to the degree of chemical weathering. Pure water percolating through a rock is capable of inducing three main chemical processes; solution, hydration and hydrolysis. By virtue of dissolved carbon dioxide and oxygen, rain water can regularly induce two further processes, carbonation and oxidation. Although these subjects have been described separately it is very rare for one process to operate alone and their overall effectiveness is largely the result of interactions between various chemical reactions. The vital importance of solution lies in its role of transporting the products of other weathering processes. Many solutions are dependent upon external factors for their maintenance e.g. the solubility of ferrous iron and manganese rises rapidly under slightly acidic conditions, but it requires pH 4 or less to render alumina soluble. This explains why in different localities the weathering residue from the same rock can vary widely. It also accounts for reprecipitation as the properties of the solvent change during migration through the soil and weathered mantle e.g. ferricrete, calcrete and silcrete.

During hydration many different minerals can incorporate water in their molecular structure e.g. haematite may become limonite; montmorillonite may swell greatly on becoming hydrated. The swelling factor is one of the most important aspects of hydration and it is believed to be a major cause of the crumbling of coarse-grained igneous rocks by the progressive expansion of their hydrated minerals:

$$\text{silicate}{=}O + H_2O \longrightarrow \text{silicate}{=}(OH)_2.$$

This reaction may progress to a stage when monomeric silicic acid is produced:

$$\text{silicate}{=}Si{=}O + 3H_2O \longrightarrow \text{silicate}{=}(OH)_2 + Si(OH)_4.$$

Surface hydration mechanisms of this type explain rapid dissolution of quartz in warm, humid climate soil profiles.

The process of hydrolysis can best be described using the following example. This reaction involves a virtually complete disruption of the original silicate lattice of orthoclase and the removal of potassium metal ions. The rearranged silicon and aluminium ions can accomodate more water so the final product is the hydrated clay mineral kaolinite. The sodium- and calcium-rich feldspars, olivine, augite and hornblende are all decomposed in a similar manner. Hydrolysis is accelerated by impurities in the water, especially the presence of carbon dioxide:

$$\underset{\text{orthoclase}}{2KAlSi_3O_8} + 2H_2O \longrightarrow \underset{\text{kaolinite}}{Al_2Si_2O_5(OH)_4} + \underset{\text{soluble potassic acid}}{K_2O} + \underset{\text{soluble silica}}{4SiO_2}.$$

The combination of an element with oxygen dissolved in water is one of the most frequently observed weathering phenomena e.g. iron when released by one of the other chemical processes is rapidly oxidized to the ferric state in the form of haematite or limonite (its hydrated equivalent).

Carbonation (i.e. chemical alteration by carbonated water) is possible because dissolved atmospheric carbon dioxide turns rainfall into a weak carbonic acid with an average pH of around 6. When the water percolates underground carbon dioxide is rapidly dissolved from soil air. Experiments have shown that orthoclase decomposition is greatly accelerated in CO_2-rich water, apparently due to the ready ionization of weak carbonic acid into hydrogen and bicarbonate ions:

$$\underset{\text{orthoclase}}{2KAlSi_3O_8} + 2H_2O + CO_2 \longrightarrow \underset{\text{kaolinite}}{Al_2Si_2O_5(OH)_4} + \underset{\text{potass. carbonate}}{K_2CO_3} + \underset{\text{soluble silica}}{4SiO_2}.$$

Weathering of other main feldspar groups also results in the development of clay minerals with Na_2CO_3 from the soda-rich minerals and $Ca(HCO_3)_2$ from the calcium-rich minerals. The dissolution of limestone is another example of chemical alteration by carbonation:

$$CaCO_3 + H_2CO_3 \longrightarrow Ca^{2+} + 2(HCO_3)^-.$$

The calcium and bicarbonate ions may then be removed in solution. The amount of dissolved $CaCO_3$ that may be transported by the water is very sensitive to variations in the amount of dissolved carbon dioxide.

Mechanical Weathering.

There are four major processes which may lead to the disintegration of the bedrock; dilatation, thermal expansion, crystal growth and organic activity.

1) Dilatation: As surface erosion and unloading removes the top of the rock column the confining pressures on deeply buried rocks diminish and the block

may adjust to this by upward expansion in the form of closely spaced joint systems and sheeting structures e.g. Colorado Plateau sandstones with 30 metre joints in the Monument Valley (Rice, 1977). The main significance of these joint systems is the opportunity they afford for deep penetration by other weathering agencies.

2) Thermal Expansion: Stresses are induced by the thermal gradients just below rock surfaces by solar heating. As rocks are generally poor thermal conductors, quite high thermal gradients are achieved, especially in deserts and at high altitudes. Purely thermal stresses seem insufficient to break up surface rocks, even in desert conditions, but long term effects have yet to be investigated. The existence of pronounced chemical alteration, particularly in shady sites where more moisture is available, and the expansion of exfoliation shells on boulders, shows that chemical weathering can still be relatively important even in arid areas. This is probably because night condensation still occurs. The products of desert weathering however differ from those of more humid areas being generally coarser and with a lower proportion of clay or organic material. The swelling of certain clay minerals due to absorption of water is another cause of surface disintegration. It is particularly effective in shales, mudstones and greywackes due to their high clay mineral content.

3) Crystal Growth: The most important influence of crystal growth on rock disintegration is ice crystallization which can produce a closed, high - stress system as the solution enters the solid solution phase and crystallization begins from the outside. Even where optimum conditions (rapid freezing, 80% moisture saturation) are not present, considerable frost damage can result from repeated cycles of freezing and thawing. In addition to ice crystallization the crystallization of salts (e.g. sodium chloride, gypsum, calcite) is important in rock disintegration under certain conditions. Permeable rocks such as sandstone and chalk are particularly susceptible to

disintegration by salt crystal growth whereas igneous rocks are much less so. This type of weathering is particularly effective in polar and desert areas. In high latitudes snowflake nuclei provide salt which tends to accumulate near rock surfaces because melting and run-off are minimal. In arid regions the excessive evaporation causes salts to be drawn up from depth in capillary films and crystallization to occur at the surface. Some workers have suggested that the precipitated salts may further contribute to rock shattering by expanding under the intense daytime heating of desert areas since they have higher thermal expansion coefficients than most rocks (Rice, 1977).

4) Organic Activity: The effect of organic activity on weathering is discussed in section 8.1.9 in order to discuss all aspects of the role of organic activity in mineral deposit formation in one section.

It appears that it is only in deserts and extreme northern latitudes that climatic controls encourage relatively vigorous mechanical weathering. Strakhov (1967) stressed that tectonic uplift may affect the relative influence of mechanical weathering. He argued that with increased relief amplitude, mechanical denudation in the form of surface wash becomes so intense that it finally suppresses chemical weathering altogether. However to achieve this state in the humid tropics demands exceptionally rapid uplift and is therefore much more likely to be achieved in temperate latitudes where chemical weathering is less active.

The importance of weathering in ore deposit genesis must be viewed with respect to two other main aspects of mineral deposit formation i.e. syn-diagenetic concentration and geochemistry of the basement (Samama, 1973). The concentration in situ of certain metals in very specific and localized sedimentary environments is the major factor without which many economic deposits would not exist. Also it appears that from studies of basement geochemistry of regions from which mineralized formations were derived that the richer a basement so the richer are the resulting detritic formations. It is difficult to assess the relative importance of these three factors as the separation of syngenetic processes from the other two is more obvious than separating the influence of basement geochemistry from that of weathering. Samama (1973) used the following example to illustrate this problem. Simple mechanical erosion of a copper-rich basement (x5 the clark) would produce a certain supply rate of copper. If weathering was influenced by high precipitation a majority of the copper would have been leached out before erosion so the rate of copper supply would be nearly normal i.e. less than x5 the clark. However if the weathering was of the monosiallization type a weathered cover x3 to x5 richer than the fresh rock would result i.e. x15 to x20 higher than the clark and so the rate of supply of copper to the sedimentary basin would be extremely high.

Depth of Weathering.

Figure 8.2 illustrates the range of climatically controlled weathered deposits which vary from very shallow in Arctic soils, to 1 - 3 metres in temperate regions and over 100 metres in some cases in the humid tropics. The thickness of weathered bedrock on slopes where transport of weathered material is limited depends on the balance between the depth of the weathered profile and the degree of transport limitation. The first factor is dependent itself upon rock resistance, jointing, permeability and climate. The transport limitation is due to the angle of slope, amount of surface run-off, rate of creep, frequency of slides, wetting/drying and frost heaving (Chorley et al, 1984). Thus the thickest weathered mantles occur in areas of moderate to low relief, gentle slopes, good drainage, unimpeded through-flow and rapidly weathered bedrock. However some deep soils in temperate regions may also be a relict of past tropical climates.

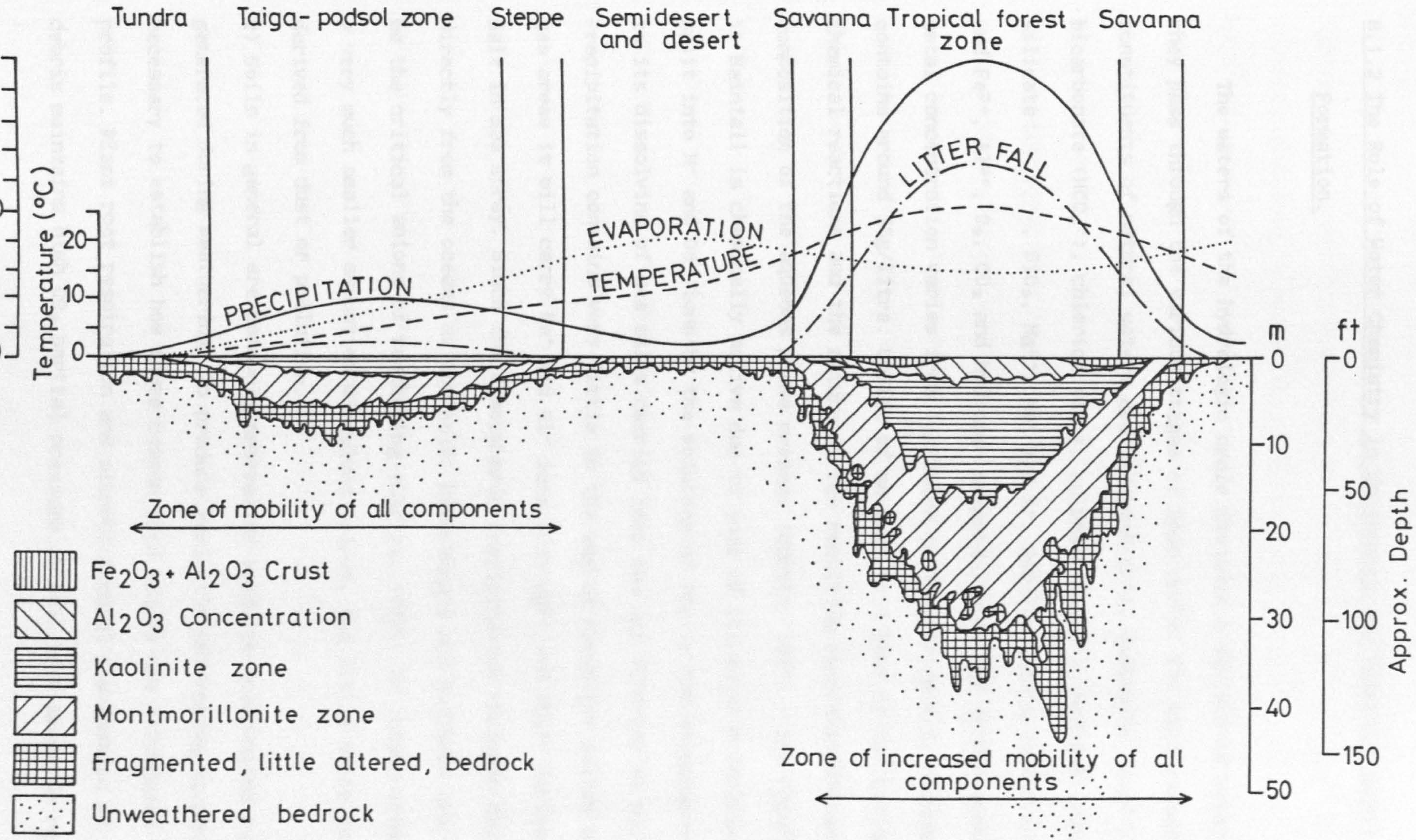


Figure 8.2. Schematic representation of the weathered mantle at various depths in different climatic regimes. (After Chorley et al, 1984, Figure 9.16).

8.1.2 The Role of Water Chemistry in Weathering and Mineral Deposit Formation.

The waters of the hydrologic cycle dissolve a variety of minerals as they pass through the various stages of that cycle. The main inorganic constituents of natural waters are hydrogen (H^+), carbonate (CO_3^{2-}), bicarbonate (HCO_3^-), chloride (Cl^-), sulphate (SO_4^{2-}), hydroxyl (OH^-), silicate , SiO_2, Mg^{2+}, Na^+ and K^+. Smaller quantities of iron as Fe^{2+} and Fe^{3+}, Al^{3+}, O_2, CO_2 and hydrogen sulphide (H_2S) are also present. The total concentration varies from 1mg/litre to 100g/litre e.g. seawater contains around 35g/litre. Uplift of an area to a zone of weathering starts chemical reactions and the nature of the resulting minerals depends on the composition of the aqueous phase present (Curtis, 1977) - see Figure 8.3.

1) Rainfall is chemically active due to some of its water molecules being split into H^+ and OH^- ions by the solution of CO_2 in the atmosphere and due to its dissolving of sea salts carried into the air (Chorley et al, 1984). Precipitation contains very little in the way of dissolved solids although in sea areas it will carry Na^+ and Cl^- ions with Mg^{2+} and SO_4^{2-} derived from the salt in sea spray. Since the atmospheric precipitation includes NaCl cycled directly from the oceans as aerosols, bicarbonate and sulphate are judged to be the critical anions of weathering (Curtis, 1977). In inland areas there is a very much smaller amount of dissolved solids, the little there is being derived from dust or pollution.

2) Soils in general are commonly neutral or acid, so acid anions must be generated during weathering. To produce realistic weathering equations it is necessary to establish how these common acid anions are introduced into the profile. Plant root respiration and microbiological degradation of plant debris maintains high CO_2 partial pressure in soil pore space so reversible

dissolution and dissociation take place:

$$H_2O(l) + CO_2(g) \rightleftharpoons H_2CO_3(aq) \rightleftharpoons H^+(aq) + HCO_3^-(aq). \quad (1)$$

This CO_2 dissolved in precipitation forms carbonic acid (H_2CO_3) which reacts with soil minerals by hydrolysis converting the frame- and sheet-silicates into clays. The result on fresh volcanic or alluvial soils is that the soil is leached and converted to an acid type. In humid tropical weather conditions soils are extensively leached of Ca, Mg, Na, K and Si. The residuum is enriched in Al and Fe, water giving kaolinite and soil sesquioxides (hydrated oxides of Fe and Al). Acid igneous rocks, sediments and metamorphic rocks are all similarly altered and quartz is often completely removed. However under cool, humid climatic conditions quartz is relatively more stable and tends to accumulate in podsols (Curtis, 1977). Given equation 1 above the weathering of anorthite can now be expressed:

$$CaAl_2Si_2O_8(s)+3H_2O(l)+2CO_2(g) \longrightarrow Al_2Si_2O_5(OH)_4(s)+Ca^{2+}(aq)+2HCO_3^-(aq). \quad (2)$$

Whether or not the soil would be acid depends on the relative rate forward of the reactions 1 and 2. High activities of the reactants (water and CO_2) and low activities of dissolved Ca and bicarbonate (the solution products) favour this reaction. These conditions are found in areas of high rainfall, organic productivity and good drainage (for solute removal). Higher temperatures increase reaction rates and organic productivity. Therefore the fact that chemical weathering is most intense in well drained areas of humid and subtropical regions is not surprising.

As mentioned earlier sulphate is also an important ionic species. The key reaction involving sulphate is the oxidation of sulphides of which pyrite is the most abundant:

$$2FeS_2(s) + 4H_2O(l) + 7^1/_2O_2(g) \longrightarrow Fe_2O_3(s) + 8H^+(aq) + 4SO_4^{2-}(aq). \quad (3)$$

pyrite haematite

Ferric iron is very stable and is highly insoluble in soil systems. Therefore the solution products form sulphuric acid which is very active in metal cation replacements similar to equation 2.

The equations 1 and 3 show the introduction of acids to soil profiles, the rate of generation of acids is one of the most vital controls of weathering. All these reactions involve dissolved species so the removal of the aqueous phase should drastically reduce reaction rates. It may be thought that chemical weathering is not important in hot desert environments, however pore solutions are often present and chemical weathering in hot arid regions contributes more to the total weathering than was once thought.

3) Small gold crystals of very high purity have been found intimately associated with iron oxide in the laterite profiles in the Yilgarn Block, Western Australia. These suggest that Au and Ag may have been dissolved, transported and redeposited during lateritization. Experimental evidence (Mann, 1984) suggests that very acid chloride solutions are generated in lateritic proflies by ferrolysis (the oxidation and hydrolysis of Fe) and these are responsible for the dissolution of the Au and Ag. In the deep and well developed laterite profiles the oxidation of pyrite (an accessory mineral which may contribute strongly to the ferrolysis process) probably occurs as two discrete reactions. The first of these occurs at the weathering bedrock front:

$2FeS_2 + 2H_2O + O_2 \longrightarrow 2Fe^{2+} + 4SO_4^{2-} + 4H^+.$

The second reaction is the oxidation of the ferrous ion at, or near, the water table:

$2Fe^{2+} + 3H_2O + {}^1/{}_2O_2 \longrightarrow 2FeOOH + 4H^+.$

As a consequence of this ferrolysis reaction the zone becomes exceedingly acid and ground water seepages with pH values less than 2.5 are common in areas such as Darling Scarp (Mann, 1983).

Whenever such ferrolysis occurs hydrogen ions must be produced and an acidic profile should develop. However if bicarbonate ions are present (from the weathering of basic rocks) this acid production may be neutralized. This is not thought to have occurred in the Yilgarn block because substantial laterite cappings and ferruginous ions are observed. So the development of low pH and the redistribution of Au and Ag in lateritic weathering profiles appear to be more common over granitic and gneissic basement as these processes may be inhibited by the presence of carbonate in the weathering zone of basic rock sequences.

4) An evolution in groundwater chemistry can be described (Smith, 1981). In deep confined aquifers the water composition will be affected by SO_4^{2-} and cation exchange with sodium-rich clay minerals will lead to a decline in Ca and Mg and a rise in Na. Calcite may be precipitated as an intergranular cement at this stage. In deeper zones Cl^- increases at the expense of SO_4^{2-} and HCO_3^- so sulphide minerals may precipitate. In deeper parts of regional confined aquifers the resulting NaCl-rich brines are sometimes hot and may be more concentrated than seawater.

5) Red beds are fully discussed in a separate section (8.1.8) but one or two points are worth mentioning with regard to the influence of water chemistry in red beds and mineral deposit formation. Mildly oxidizing saline groundwaters are commonly proposed as the metal-bearing solutions in red beds (Rose, 1976; Gustafson and Williams, 1981). Metal complexing with chloride ions for Cu and Ag, or phosphate and bicarbonate ions for U, is given as the mechanism for increasing the metal-carrying capacity of such solutions (Rose, 1976; Langmuir, 1978). However these metal complexes must be sufficiently stable to inhibit metal adsorption by ubiquitous secondary iron oxides (Zielinski et al, 1983). Also low sulphide ion concentrations and high Eh are required to inhibit the formation of relatively insoluble sulphides of Fe, Cu, Pb and Zn or low valence oxides of U and V. The leaching results of

Zielinski et al indicate more effective liberation of Fe, Pb and Zn during red-bed bleaching (leading to dissolution of secondary iron oxides) than during more oxidative leaching. This contrast is less dramatic on Cu and Co so variable ratios of Fe + Pb + Zn/ Cu + Co in sediment-hosted stratiform deposits may, in part, indicate metal fractionation related to the Eh of leach solutions. However it must be noted that such metal fractionation is only a contributory factor to the high Co content of some SSCU deposits. Those areas of White Pine, the Zambian Copperbelt and the Kupferschiefer with high Co occur above basal mafic volcanics (Badham, pers.comm.) so source rock also influences mineral deposit chemistry.

8.1.3 The Role of Water Volume in Metal Concentration and Mineral Development.

The importance of the availability of water, its volume, circulation and periodicity, to the rates and products of weathering has been stressed by Trudgill (1976). He noted that, in arid regions, the small amount of available water, the return of water to the surface by capillary rise accompanying evaporation, and the paucity of organic matter slows down weathering rates. Here, and in waterlogged situations, montmorillonite, illite and chlorite are the resulting clay products. The smectite group (e.g. montmorillonite, nontronite and beidellite) is especially notable for the way in which it takes up and loses water (i.e. its adsorption properties) and for its base exchange properties so it may be particularly important to mineral deposit formation. Smectites develop only when certain conditions are met, such as evaporation should exceed precipitation; leaching should be negligible and alkalic conditions should prevail so that a low Al:Si ratio is maintained. Consequently their occurrence is favoured by a semi-arid, warm climate. In humid regions with good drainage, intensive leaching and abundant

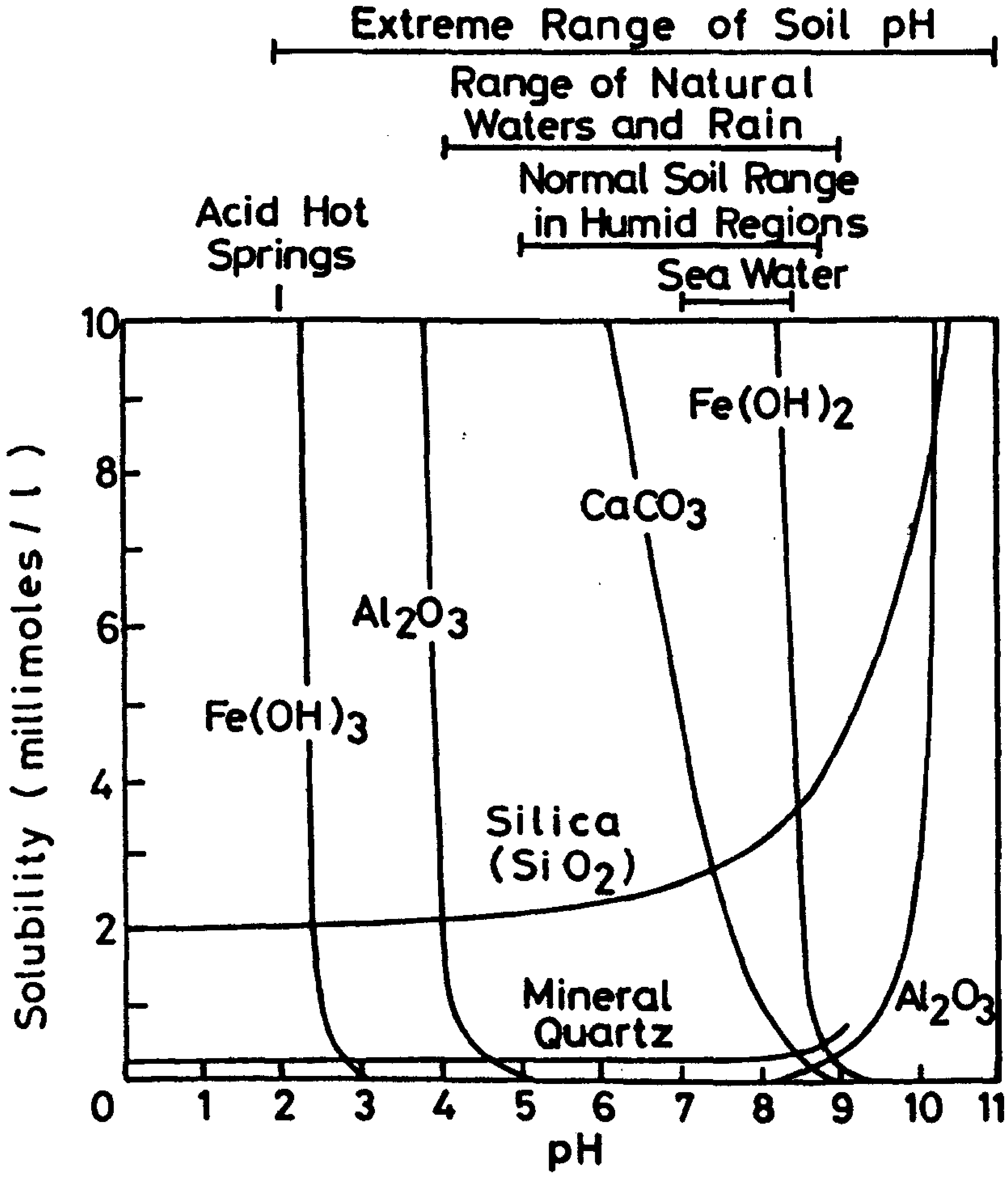


Figure 8.3. The solubility for some products released by chemical weathering in relation to pH. (After Chorley et al, 1984, Figure 9.2).

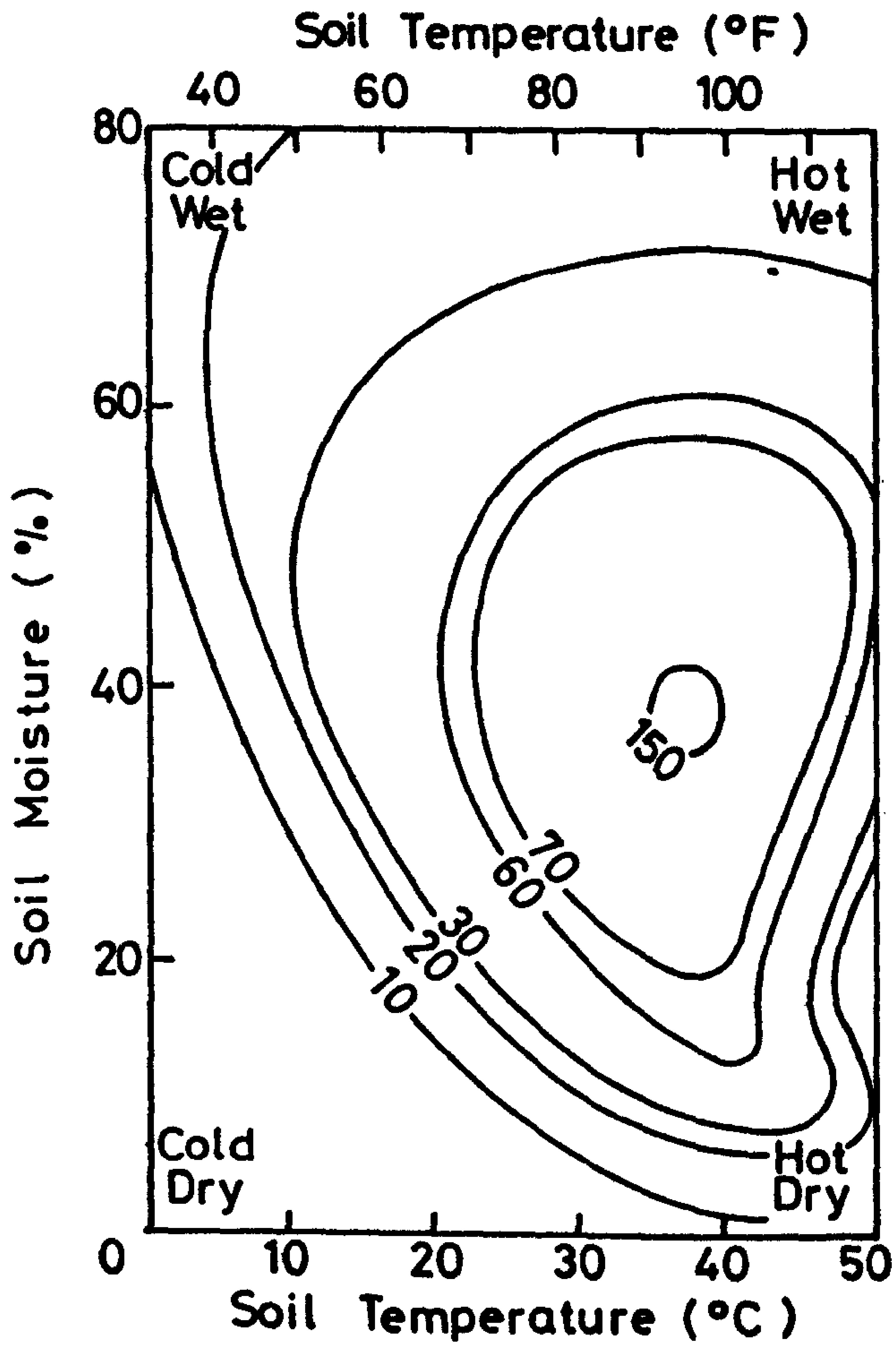


Figure 8.4. CO_2 production as a function of soil temperature and moisture content acting as a measure of organic activity. (After Chorley et al, 1984, Figure 9.9).

vegetation, deep weathering results leading to the production of kaolinite or, in tropical regions, gibbsite or goethite. Chemical weathering is most active in wet climate as water is essential to processes such as hydration, hydrolysis and carbonation (Small, 1972).

1) Weathering rates are controlled by the amount of water input into the weathering mass, water output, water chemistry, organic factors (e.g. solution and chelation) and by the susceptibility of given minerals to weathering. In environments classed as weathering-limited, transport processes (e.g. run-off and wind action) are more rapid than weathering processes so little or no soil can develop. However in transport-limited environments weathering rates are more rapid than transport processes and soil or debris cover develops. Clearly moisture and associated vegetation cover may have as important an effect upon weathering as the gross influences of climatic regime and rock type (Brunsden, 1979). Chorley et al (1984) describe two types of experiments involving the efficacy of water upon weathering. The first showed that the most effective way of weathering a coarse granite block was to use wetting and drying in association with thermal changes. Other tests show that certain rock types (e.g. chalk, limestone and sandstone) are more susceptible to salt weathering than others (e.g. igneous rocks and black shale). The samples were wetted in a salt solution and then dried. The most effective salt used was Na_2SO_4 and the rate of disintegration appeared to be related to water absorption capacity. This weathering mechanism is particularly applicable in deserts and coastal arctic areas.

2) The speed of mechanical weathering by freeze-thaw methods increases rapidly after wetting so it is greatly affected by an increase in water volume.

3) The rate of chemical weathering is dependent upon the relation between the rate of chemical solubility at a given mineral face and the rate of water

flow against that face (Trudgill, 1976) e.g. in a high solubility and slow flow environment, chemical equilibrium is achieved and the weathering rate is controlled by the solubility level i.e. the saturation value of the solution. However under lower solubility and faster flow conditions the weathering rate is controlled by the rate of flow and the solution velocity (i.e. the rate of achievement of the saturation value).

4) Trudgill (1976) also noted that the effect of the velocities of percolation on the products of weathering could be shown by the weathering of albite. In areas of bad drainage stagnant conditions develop with low percolation velocities. Equilibrium is quickly reached and not all Na^+ and Mg^{2+} cations are flushed out so those remaining react with Al^{3+} and Si^{4+} to give montmorillonite. In this way the initial weathering of anorthite produces montmorillonite and $3Ca(OH)_2$ in solution:

$$Mg^{2+} + 3NaAlSi_3O_8 + 4H_2O \longrightarrow 2Na_{0.5}Al_{1.5}Mg_{0.5}Si_4O_{10}(OH)_2 + 2Na^+ + H_4SiO_4.$$

albite montmorillonite [in solution]

In those areas with better drainage and higher percolation rates the Na^+ and Mg^{2+} are dissolved and removed. But Al^{3+} and Si^{4+} dissolve more slowly to form kaolinite:

$$2H^+ + 2NaAlSi_3O_8 + 9H_2O \longrightarrow Al_2Si_2O_5(OH)_4 + 2Na^+ + 4H_4SiO_4.$$

albite kaolinite [in solution]

Where rapid drainage occurs, percolation velocities are faster than solution velocities and only Al^{3+} remains to give bauxite while Na^+, Mg^{2+}, Ca^{2+}, Si^{4+} are all removed:

$$H^+ + NaAlSi_3O_8 + 7H_2O \longrightarrow Al(OH)_3 + Na^+ + 3H_4SiO_4.$$

albite gibbsite [in solution]

Thus, given suitable weathering material, tropical rugged relief, intense rainfall and rapid drainage tends to produce gibbsite. However in a similar way orthoclase can be weathered to illite and if continued leaching occurs, kaolinite may result:

$$5KAlSi_3O_8 + 4H^+ + 4HCO_3^- + 16H_2O \longrightarrow$$

othoclase $\quad KAl_5Si_7O_{20}(OH)_4 + 8H_4SiO_4 + 4K^+ + 4HCO_3^-.$

illite

$$2KAl_5Si_7O_{20}(OH)_4 + 2H^+ + 2HCO_3^- + 13H_2O \longrightarrow$$

illite $\quad 5Al_2Si_2O_5(OH)_4 + 4H_4SiO_4 + 2K^+ + 2HCO_3^-.$

kaolinite $\quad$ soluble potassium carbonate

5) It is well known that clay minerals have significant but different capacities to bind by adsorption, ion exchange and complexing a host of metals and non-metals (Weiss and Amstutz, 1966). It is therefore interesting to note that the distributions of different types of clay minerals in the Niger Delta have been interpreted as consequences of water volume in the region i.e. severe leaching, a rapid rate of infiltration of sediments, permeability, shale dewatering and marine/non marine water influence (Olorunfemi et al, 1985). In the sandy central delta area, permeability and porosity are high so this region has the highest groundwater flow rates and constant meteoric water flushing. With normal compaction and high leaching by meteoric water of cations, residual kaolinite and halloysite ($2H_2O$) deposits develop. In the eastern delta high Al-smectites, illite and mixed layer illite-smectite dominate. Weathering of volcanic rocks produces Mg-deficient smectites (mainly beidellites). However the occurrence of these clay minerals is also thought to be due to high organic and carbonate contents: the latter acting as a cement, suggesting lower porosity and permeability than the central delta so meteoric water flow rates are low and leaching is less effective. Shale dewatering (a normal consequence of compaction) also contributes cations (Ca^{2+}, Mg^{2+}, K^+) to smectites and illites which are found at depth. Olorunfemi et al (1985) concluded that it was possible to relate

the clay mineralogy of the Niger Delta to the hydrology or groundwater flow net as well as interaction with sea water. With a particular model-type, given the permeability and flow rates, the rainfall and flow volumes of the various tributaries, the observed clay mineral distribution patterns were satisfactorily explained.

6) Solution is an important weathering process for all rocks, but it is especially destructive to the carbonate sedimentary rocks. The susceptibility of limestone to weathering partly depends upon its purity, but even more on the absolute amount of water available. Therefore in desert areas limestone outcrops form residual hills and escarpments, whereas other rocks are more rapidly denuded (Chorley, 1969).

8.1.4 The Effect of Run - Off on Weathering.

Run-off is that portion of the rainfall which ultimately reaches streams. It consists of the water which flows off the surface, instead of sinking into the ground, together with some of the water which originally sank into the ground and joins it later in the stream. The run-off is faster and greater;

1) during heavy rain than during a protracted drizzle,
2) on clay soils than on sandy soils because of lower permeability,
3) on frozen soils than on frostless soils for the same reason,
4) in treeless areas than in forests.

Although chemical weathering is dominant in humid regions, the speed of (water) erosion affects its influence; slow erosion allows chemical reactions to occur, whereas faster erosion, through increased run-off, may inhibit chemical reactions. Also the concentration of metals during chemical weathering is favoured by relatively arid climates. The lack of water volume means that material is not diluted and flushed quickly through potential depositional surroundings. Hence the amount of free percolation in the weathering mass is crucial in determining the rate and completeness of removal of chemically weathered constituents. In this respect evenly distributed persistent showers are more effective than occasional high - intensity storms or alternate wet/dry seasons (Chorley et al, 1984).

8.1.5 The Position of the Water Table.

The water table is the upper surface of a zone of unconfined groundwater below which the pores of a rock are saturated with water. This surface is uneven and variable in depth, rising during wet weather and falling in dry weather conditions.

1) The level of water table may have a direct influence upon whether or not nickeliferous laterite deposits develop. This level is influenced itself by topography which plays an important role in the distribution of economic deposits. Ideally major enrichment of saprolite occurs immediately above and below the rim (inflection point) on slopes where the water table is low at the plateau edge. Meteoric circulation can occur to greater depths and is forced through the saprolite zone. A higher water table in regions of flatter-lying land elsewhere on the plateau, or the lowland areas, prevent nickel enrichment (Edwards and Atkinson, 1986).

2) It has been suggested (Granger, 1968) that the position of U deposits in the San Juan Basin Mineral Belt, New Mexico was defined by the intersection of the palaeo-ground water table (which was periodically fluctuating) and the uplifted Morrison aquifers. He speculated that soluble carbonaceous material was carried downward into exposed Morrison Formation sandstones by meteoric water and precipitated at the water table. The extensive kaolinization in arkosic sandstones immediately below the Dakota deposit strongly suggests that solutions rich in organic acids flowed into the underlying rocks when

the water table was lowered enough to allow drainage from the swamps above. Granger proposed that the precipitation occurred at the water table because the humate was soluble in meteoric or vadose pore waters, but not in the phreatic groundwaters.

3) For some deposits the level of water table may have an indirect influence upon ore formation. Deposits such as LSBMs are preferentially deposited in solution and karst cavities so groundwater movement is significant to their formation. The water table is the level at which cave passage formation, with concommitant roof collapse, is sometimes thought to be most rapid (Williams, 1969).

8.1.6 Temperature Influence on Weathering.

Temperature has its maximum effect on accelerated erosion indirectly through its influence on plant cover and the weathering processes. Its effect on mechanical weathering is obviously very great. Frost weathering can occur only where there are atmospheric freeze-thaw cycles. Insolation weathering requires considerable diurnal fluctuations of temperature. Freezing and thawing directly alter the structure of the soil and thus make it more susceptible to the action of wind or running water. When frozen for a continuous period, the soil is largely spared from erosion so the tundra and taiga soils are free from accelerated erosion, partly owing to the long periods of freezing temperatures.

Chemical weathering operates most effectively in very warm climates for the intensity of chemical reactions is approximately doubled for every rise in temperature of 10°C (Small, 1972). However if these higher temperatures decrease the amount of infiltration this can operate to decrease the weathering rates as the processes of chemical weathering are reliant upon water and so are most active in wet climates. Biochemical weathering is also influenced by soil temperatures as humic acid is produced most effectively by moderate soil moisture of neutral pH under high soil temperatures (see Figure 8.4). These acids ($C_{40}H_{24}O_{12}$) are active in chelation and decompose silicates and amphiboles. Fulvic acids (i.e. humic acids derived from peats) are especially important agents of weathering (Chorley et al, 1984). Hence soil zones formed by different climates exhibit different humus characteristics. Tundra soils from near the Arctic Circle have a low humic acid content, low microbiological activity and hence low cation exchange capacity. In contrast, chernozen soils (e.g. from southern U.S.S.R.) have high humic acid contents, microbiological activity and high cation exchange capacity.

Temperature changes in the soil are as important as in the air above during soil formation (Critchfield, 1983). They are conducted downward slowly in the soil so that at about two or three feet depth diurnal variations are not experienced. There is little or no seasonal variation in temperature about sixteen metres below the surface. The structure of the soil may affect its conductivity as air is a very poor conductor i.e. dry, loose soil conducts heat more slowly than wet, compact soil and solid rocks.

8.1.7 Evaporation and Evaporites.

Tropical deserts develop where the potential rate of evaporation exceeds that of precipitation and rainfall is too low or too spasmodic to support vegetation. These regions usually occur in trade wind belts north and south of the equator in the subtropical arid zones i.e. 20° to 30° north and south and in the rain shadow of mountain ranges. The transfer of water vapour in the atmosphere is polewards at latitudes higher than 20° and equatorwards at lower latitudes so exaggerating the effect of evaporation greater than precipitation as water vapour is transferred out of the subtropical arid zone (Smith, 1981). About half of the area of deserts comprises outcrop subject to erosion and deflation. The rest consists of fluvial, lacustrine, paralic and aeolian sandstones and evaporites. Desert fluvial sediments rarely reach the sea so warm, clear coastal waters are ideal for manufacture of organic carbonates (see Figure 8.5). The fluvial sediments which do reach the sea are formed into off-shore bars by longshore currents behind which evaporitic lagoonal conditions commonly develop. These lagoons become sabkhas as they are filled with algae-bound marine and wind-blown sediments. It seems likely that the sporadic occurrence of deserts in the past was controlled by the size of continents, their location relative to the equator, the presence of ice caps and the freedom of circulation of global oceanic water e.g. the Permo-Triassic was a time of extensive deposition of evaporite and dune activity (Glennie, 1987).

Evaporation is the net transfer of water molecules into the air which only occurs if there is a vapour-pressure gradient between the evaporating surface and the air i.e. evaporation is nil when the relative humidity of the air is 100%. The second control is that an external source of heat must be available as the process necessitates a source of latent heat. Once evaporation has begun, its rate is affected by wind speed, since air movement carries fresh unsaturated air to the evaporating surface. The temperature of this surface also affects the evaporation rate as at higher temperatures more water molecules can leave the surface due to their greater kinetic energy. Salinity depresses the rate in proportion to the solution concentration e.g. for sea water the rate is about 2 - 3% lower than for fresh water (Barry, 1969).

It is known that when the moisture supply in the soil is limited, plants have difficulty in extracting water and evaporation (E) falls short of its maximum value (PE). One view is that the potential rate is maintained until the soil moisture content drops below some critical value. Another is that the rate of evaporation decreases progressively with diminishing soil

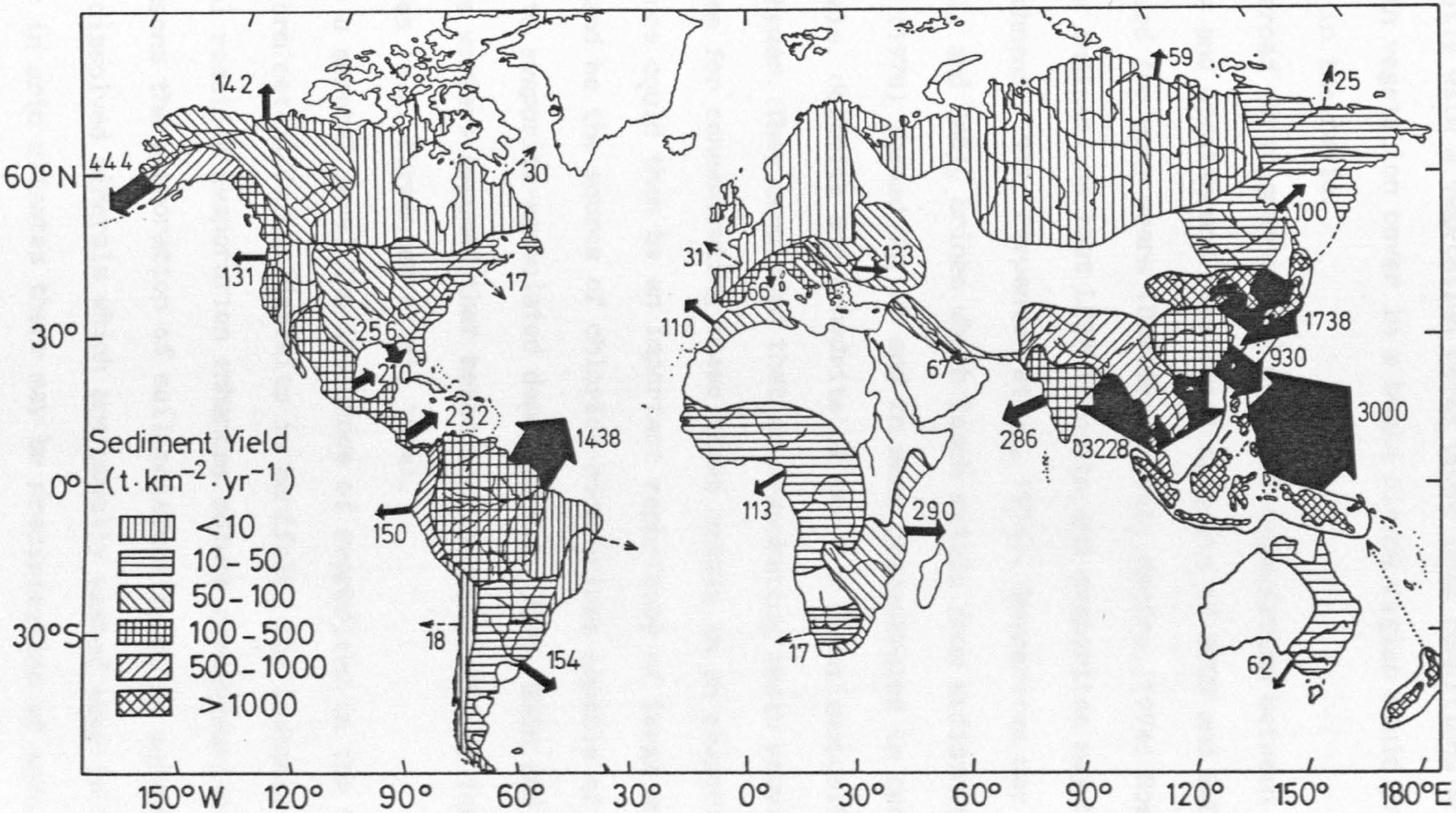


Figure 8.5. The annual discharge of suspended sediment from various drainage basins of the world. The width of arrows corresponds to relative dsicharge but direction of arrows does not indicate direction of sediment movement. Numbers refer to average input in millions of tons. Open patterns indicate essentially no discharge to the ocean. (After Milliman and Meade, 1983, Figure 4).

moisture. Undoubtedly the soil type and climatic conditions are important as the maximum soil moisture content free drainage (known as field capacity) ranges from 25mm in shallow sandy soil to 550mm in deep clay loams (Barry, 1969). Chang (1965) considered that a rapid decline in E/PE is likely in sandy soils with a vegetation cover under arid conditions whereas a heavy soil with vegetation cover in a humid cloudy region would produce different changes in the ratio.

A broad stratigraphic and regional association between evaporite deposits and stratabound base metal deposits of SSCU and SHBM types has been recognized for many years (Davidson, 1965; Renfro, 1974; Rose, 1976). There are also indications that LSBM deposits and evaporites may be related (White, 1967; Johnson, 1972; Carpenter et al, 1974). Evaporites may serve as a source for $NaCl_2$ and $CaCl_2$ brines which leach metals from sediments. Theide and Cameron (1978) showed Cu, Pb and Zn were concentrated in certain horizons i.e. shale, dolomite and anhydrite of the Elk Point evaporite sequence, Saskatchewan. They concluded that an evaporating basin provides an effective mechanism for concentrating these three metals in an evaporite sequence. Such a sequence could then be an important repository of large amounts of Cu, Pb and Zn and be the source of chloride-rich brines capable of transporting metals to evaporite-associated deposits e.g. SSCU, SHBM and LSBMs. However previous workers assumed that metal sources were outside the evaporite sequences (e.g. Carpenter et al, 1974).

On a general level the influence of evaporites in the formation of stratiform metalliferous deposits is manifold. For instance at any stage in a chemical reaction evaporation enhances solute concentration in a solution. In dry seasons the evaporation of soil moisture leads to seasonal concentrations of the dissolved minerals which are usually washed away in the wet seasons. However in arid climates there may be precipitation of nodular carbonates (Smith, 1981). A certain amount of sulphate reaches the surface cycle by the

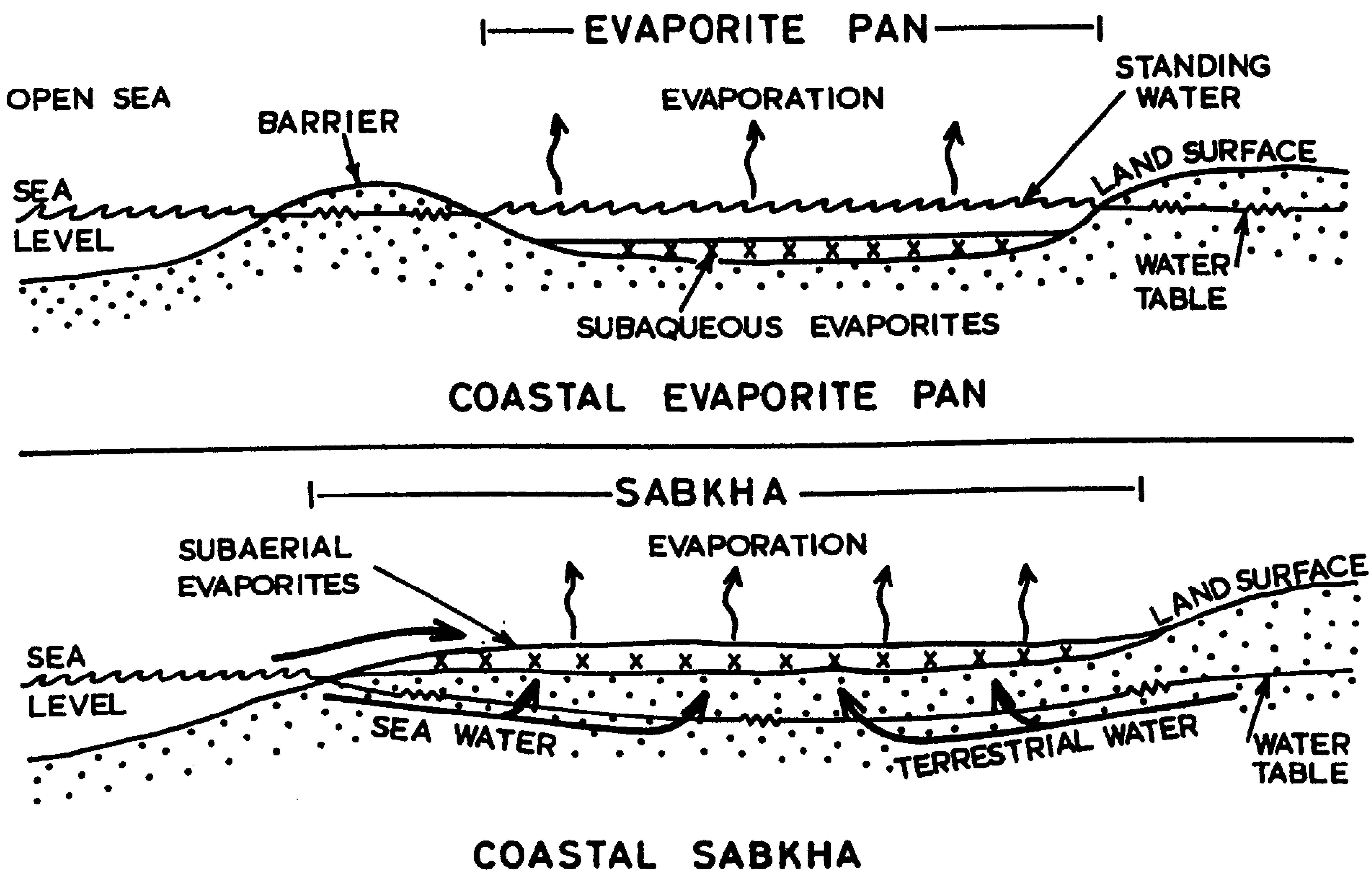


Figure 8.6. Diagrammatic cross-sections which illustrate the characteristics of a coastal sabkha as opposed to a coastal evaporite pan. (After Renfro, 1974, Figure 5b).

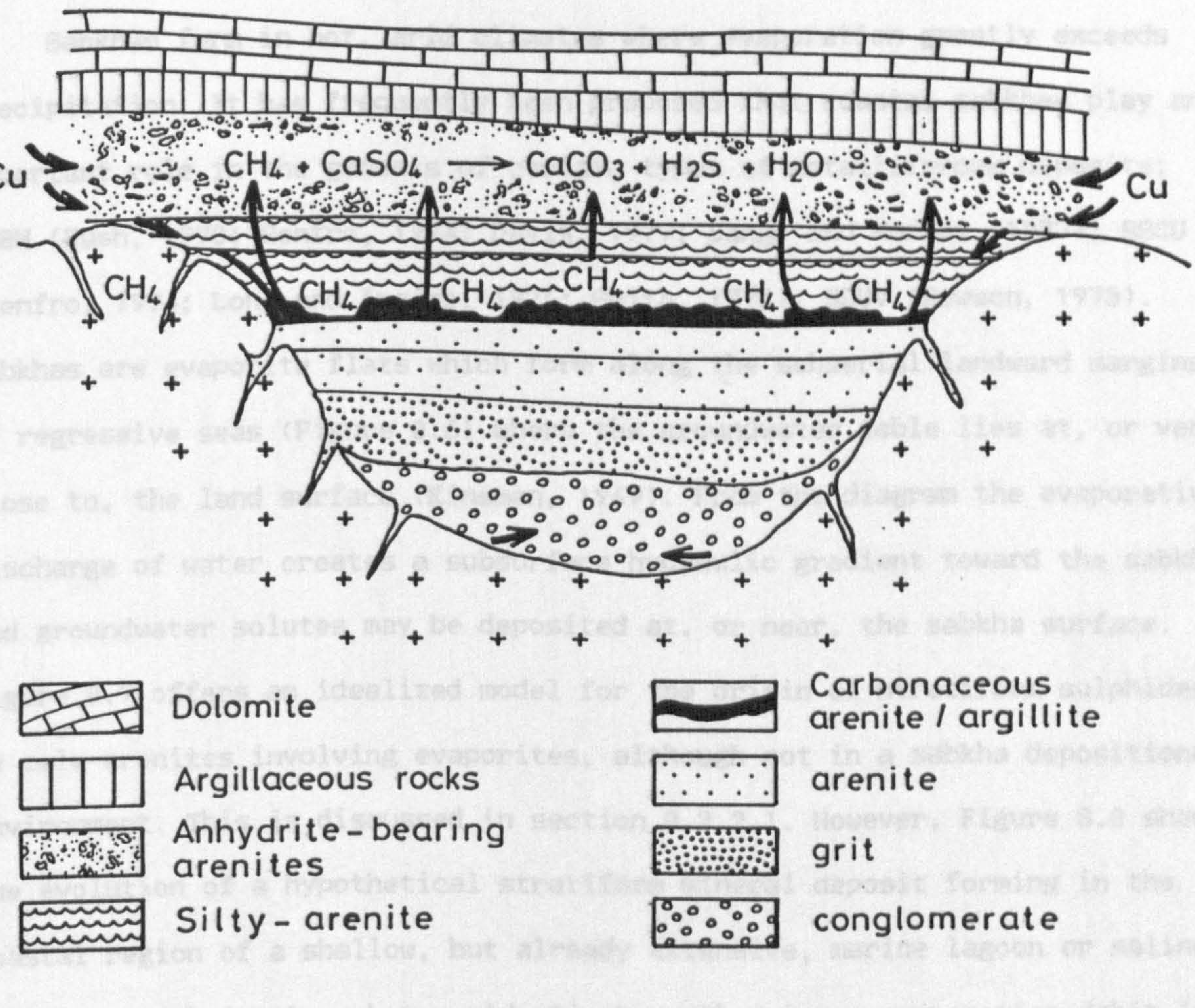


Figure 8.7. Idealized model of reduction of sulphur in anhydrite-bearing units by reaction with organically produced methane during diagenesis. (After Jacobsen, 1975, Figure 11).

dissolution of evaporite salts. Although this plays no part in such processes as chemical weathering, it does give a source of sulphur for reduction to sulphide during mineral deposit development (Curtis, 1977).

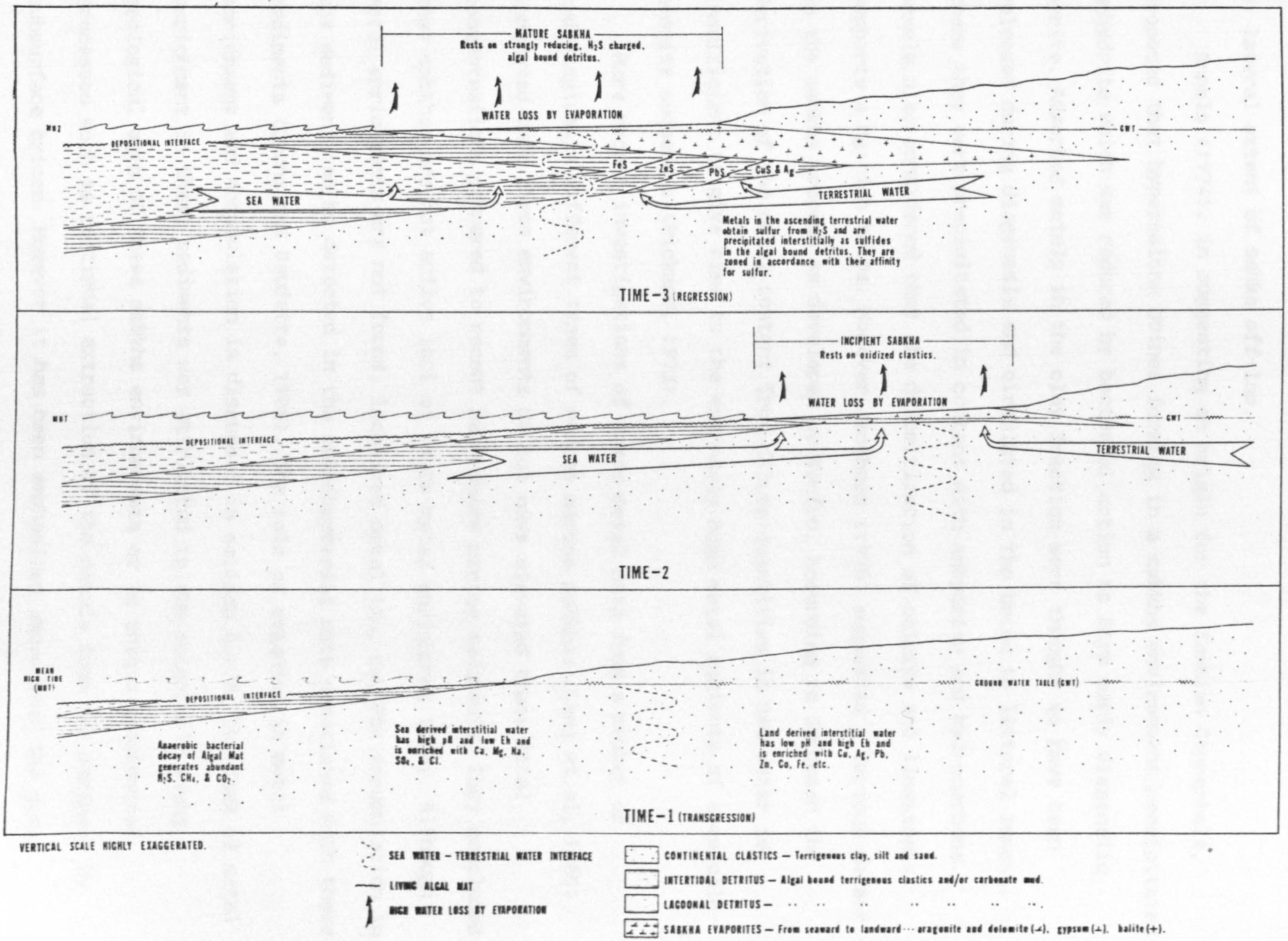
The Role of Sabkhas.

Sabkhas form in hot, arid climates where evaporation greatly exceeds precipitation. It has frequently been proposed that coastal sabkhas play an important role in the genesis of certain types of metalliferous deposits; LSBM (Bush, 1970; Renfro, 1974: Davis, 1977; Lange and Murray, 1977): SSCU (Renfro, 1974; Long and Angino, 1976; Smith, 1976): SSUV (Rawson, 1975). Sabkhas are evaporite flats which form along the subaerial landward margins of regressive seas (Figure 8.6) where the groundwater table lies at, or very close to, the land surface (Kinsman, 1969). From the diagram the evaporative discharge of water creates a subsurface hydraulic gradient toward the sabkha and groundwater solutes may be deposited at, or near, the sabkha surface. Figure 8.7 offers an idealized model for the origin of stratiform sulphides in calc-arenites involving evaporites, although not in a sabkha depositional environment. This is discussed in section 8.2.2.1. However, Figure 8.8 shows the evolution of a hypothetical stratiform mineral deposit forming in the coastal region of a shallow, but already extensive, marine lagoon or saline inland sea. Obviously a hot, arid climate with a large evaporation debit is inferred. The metals are derived from underlying terrigenous red beds and are transported from the red bed source to the sabkha trap by seaward migrating, low pH-Eh water of meteoric origin. Renfro (1974) considered that the areal extent, grade and thickness of the metalliferous dposit are dependent upon the following:

1) metal content and quantity of "source" water,

2) quantity of available reductant (e.g. organics),

3) duration of the sabkha process,



(After Renfro, 1974; figure 6).
Figure 8.8. Sequential idealized cross-sections showing the evolution of a sabkha-type sediment-hosted mineral deposits.

4) amount of "source" water that the host sediment ultimately transmits,
5) lateral extent of sabka off-lap.

Annels (1974), in suggesting an origin for the Zambian Copperbelt, proposed that hypersaline brines forming in a sabkha environment precipitated anhydrite which was reduced by bacterial action to form early diagenetic pyrite. Adsorbed metals in the clay fraction were thought to have been released during diagenesis and circulated in the brine to littoral zones where they were precipitated in contact with anhydrite and hydrocarbons. Annels also considered that the dolomitization of calcite and limestones supports a Mg-rich brine. However Jacobsen (1975) suggested a serious defect in the sabkha process as developed by Renfro. According to Jacobsen the derivation of the metal content from dilute quantities in sea water is insufficient to give rise to the extremely high metal contents of some calc-arenite sequences (Rickard, 1973).

More recent investigations of trace-metal data from a number of hydrologically different types of modern marine sabkhas (Long et al, 1985) indicated that these environments do not have elevated trace-metal concentrations compared to recent near-shore marine sediments. They concluded that sabkhas are not active loci of trace-metal enrichment today. Although metal enrichments are not found, localized metal (Zn, Cu, Pb) accumulation in the sediment can be detected in the cyanobacterial mats associated with these sediments (Lyons and Gaudette, 1985). The role of organics in metal enrichment and accumulation is discussed in section 8.1.9. The lack of metal enrichment in these sediments was attributed to the relatively young geological ages of these sabkha environments or to unique geochemical processes such as continual extraction of the metals from the sediment by subsurface brines. However it has been emphasized above that the metal enrichment of sabkhas is a diagenetic process. Hence the sampling of modern, young and thin sabkhas which have not yet developed a mature profile is unlikely to test the sabkha model of mineral deposit formation fairly.

The Generation of Chloride-bearing Brines.

The metals are thought to be widely dispersed in large volumes of evaporites and the transfer of metals into ore-forming brines requires the operation of mechanisms on very large scales.

Compaction: The expulsion of interstitial brines from the evaporite sequence during compaction would produce very large volumes of brine as modern evaporites have porosities greater than 50% (Carpenter et al, 1974) whereas those of ancient evaporites are near zero. This mechanism was proposed by Carpenter et al (1974) with metals from an outside source. However Vine and Tourtelot (1970) noted that black shales of hypersaline brines are deficient in heavy metals compared to normal marine black shales and suggested that metals could be removed during compaction of sediments.

Conversion: Brines may be generated by the release of water involved in the conversion of gypsum to anhydrite. It is thought that 0.486 cubic metres of water would be released per cubic metre of gypsum replaced (Borchert and Muir, 1964). This water would rapidly become a saturated Cl-Na-Ca-SO_4 brine through contact with, and dissolution of, NaCl-rich members of the evaporite sequence. It has been suggested (Thiede and Cameron, 1978) that metals may be freed from the dehydration reaction in a ratio similar to that determined for anhydrite yielding a Pb:Zn:Cu ratio of 5:2:1. Added to these would be metal from dissolved halite at a Zn:Pb:Cu ratio of 4:1:1. The brine produced may therefore be a saturated Cl-Na-Ca-SO_4 brine containing Pb and Zn as the significant metals.

Solution: The postdepositional solution of evaporites may be a third means of collecting metals. Gorrell and Alderman (1968) estimated that 1.25-2.5 x 10^{12} tons of salt have been lost from the Elk Point Basin. This amount of salt containing concentrations of metals seen in halite would yield up to 7.5 x 10^{5} tons of Cu and Pb and 2.6 x 10^{4} tons of Zn. The dissolution of potash-rich members of the sequence would yield Zn:Pb:Cu in the ratio of 3:2:1, whereas the dissolution of halite would yield the same metals in the ratio 4:1:1. Groundwaters moving through the sequence would become near-saturated Cl-Na-Ca-K-Mg-SO_4 brines (Thiede and Cameron, 1978).

Brine Chemistry and its influence on Metal Transport.

Ca-Na-K Chloride Brines. In saturated solutions of Na or K chloride at 18°C most chlorides dissolve in only small amounts with zinc chloride being an exception. However very saline Na-Ca chloride brines are potent solvents for base metals (White, 1968; Barnes, 1974). Most stable zinc complexes at ambient temperatures have the formula $2NaCl.ZnCl_2$ and appear to be non-resistant to sorption and precipitation processes (Sonnenfeld, 1984). The solubility of galena increases noticably at 25° - 90°C in sodium chloride brines with a reduced sulphur concentration and a pH of 8 - 13 (Hamann and Anderson, 1978). Galena is usually relatively insoluble in brines at atmospheric pressure.

Complexing. This aids the transport of metals considerably, however the composition of the complex is again vital:

1) Lead carbonate is the dominant complex in sea water (Byrne, 1981) but sufficient quantities of carbonate ions are not available in more concentrated brines.

2) Bisulphide complexes are not important near ambient temperatures (Herr and Helz, 1978). The possibility of Pb transport as bisulphide or sulphide complexes at temperatures of 100°C has been questioned (Hamann and Anderson, 1978). Clearly, these two complexes have very limited spheres of influence in the transport of metals.

3) Slightly acidic Pb and Zn chloride complexes with a low sulphur content can lead to substantial accumulations by a slight reduction in the pH (Anderson, 1973). The metals may then travel as simple chloride or as compound organo-chloride complexes. The formation of such complexes appears to increase the reactivity of the compounds:

e.g. Pb complexes with humic acids increase the solubility of galena (Lelen and Goni, 1974),

e.g. Cu and Zn form soluble metallo-organic complexes that keep copper in solution even in the presence of sulphide precipitation (Hallberg et al, 1980),

e.g. dichloride and trichloride complexes of Cu increase its solubility by up to two orders of magnitude (Rose, 1976).

In addition to metals from inflowing seawaters the metal content of evaporites is influenced by cations adsorbed onto suspended clays within the sequence (Ongley et al, 1981). Hence shale and evaporitic dolomite have increased metal contents than halite and carnallite (Sonnenfeld, 1984).

To Deposit Metal Sulphides.

In dilute solutions the pH above which metals precipitate increases in the following order according to Pasztor and Snover (1983):

(Sn^{2+}, Zr, Fe^{3+}) < Al (Zn, Cu, Cr^{3+}) < (Fe^{2+}, Pb) < (CO^{2+}, Ni^{2+}, Ca) < Mn^{2+} < Mg^{2+}.

The mobilized metals precipitate where they encounter waters with a supply of sulphide ions e.g. those associated with H_2S (sour gas) of carbonate or evaporite provenance and where the environment changes to a negative Eh and a high pH. A zonation of the metals which is so obvious in many mineral deposits occurs because the critical concentration of hydrogen sulphide varies for each metal and is dependent upon the temperature of the solution (Ganeyev, 1976). The distance of migration of a metal is inversely related to its solubility (Sonnenfeld, 1964). Where adequate amounts of sulphur are not available in a brine of high ionic strength, the complexed metals may even be precipitated in native form e.g. Pb from Jurassic Louann Salt (Carpenter et al, 1974).

In conclusion the hypersaline brines of evaporites concentrate base metals faster than they concentrate major cations and probably by biogenic means. Evaporitic precipitates therefore become major repositories of metals such as Cu, Pb and Zn which are found both in their crystal lattice (e.g. Fe and Mg in gypsum) and in point distribution (the base metals). The incorporation of these metals is thought to be a function of the rate of precipitation (Thiede and Cameron, 1978). The remobilization of such metals occurs upon recrystallization of the precipitate, and they probably travel as chloride complexes and are redeposited in surrounding porous strata in the presence of hydrogen sulphide or by replacement of pre-existing sulphides e.g. diagenetic pyrite at McArthur River, Australia (Williams, 1978).

8.1.8 The Influence of Red Beds in Mineral Deposit Formation.

Before discussing the possible role of red beds in mineral deposit formation it is necesssary to accept the assumption that red beds may be indicative of arid climates especially when they are found in close stratigraphic association with evaporite deposits. Walker (1967a and b) suggested that terrestrial water associated with sabkha evaporites bordering the northwest Gulf of California is sufficiently oxygenated to cause in situ oxidation and reddening of terrigenous clastics. Recently investigations (Zielinski et al, 1983) into the development of red pigmentation in a red bed sequence in northern Baja California have been carried out to study the mobility and distribution of heavy metals. This was based on the generally accepted view that in soils and sediments secondary oxides of Fe, especially hydrated forms, may concentrate metal ions (e.g. Wedepohl, 1970). The mechanism of uptake is adsorption, the degree of which is dependent upon the type and degree of crystallinity of the iron oxides, the solution chemistry (pH, Eh, concentration of ions and complexing agents) and the relative concentrations of oxide host and adsorbed species (e.g. Vuceta and Morgan, 1978). The increased permeability in a water-saturated environment promotes efficient interaction between mineral grains and co-existing solutions, by increasing both rate of mineral alteration and effective transport distance of dissolved species. High intergranular permeability probably aids the above so coarser grained red beds may be more susceptible to changes which result in growth or destruction of secondary iron oxides and so influence metal sorption or desorption (Zielinski et al, 1983). The leaching results also suggest that under equivalent leaching conditions the rate of metal liberation should be greatest in young red beds which contain a higher proportion of more reactive, hydrous iron oxides. Older red beds may react more slowly because of the greater proportion of crystalline iron oxides. However these older red beds provide a potentially greater amount of metals via iron oxide dissolution as the total abundance of secondary iron oxides and hence the percentage of metals associated with them increase with age. The results of this research suggest that the development of red beds which are well flushed by suitable oxidizing or reducing pore fluids may be sources of significant quantities of heavy metals.

8.1.9 The Influence of the Biomass on Metal Deposit Formation.

A number of sediment-hosted base-metal sulphide deposits have organic carbon contents considerably above average for shales and it appears that so-called "organic" processes could have played a significant, or even critical, role in the concentration of the ore metals (Roberts, 1973). Organic matter also plays an important role in the transport of metal complexes (Barker,

1982) and has been cited as important in the formation of strata-bound and stratiform ore deposits (Anderson and Macqueen, 1982; Macqueen and Powell, 1983).

The Effect of Biomass on Weathering.

Mechanical Weathering: Biological stresses which induce physical weathering are of two main classes; faunal and floral. Animals only really have an important role in disturbing the detritus already produced and thereby enhancing the efficacy of other weathering processes. However earthworms can bring an average of 43 tonnes/hectare/year to the surface i.e. a 5mm layer. This figure may be exceeded in parts of the tropics and the activity of termites may be twice that of earthworms (Chorley et al, 1984).

Plant root growth may also generate important stresses since the cellulose of cell walls is stronger than many metals so growing plant roots in favourable situations have the capacity to wedge open bed-rock joints. Tree tap-root systems commonly extend to depths of three metres and fine tree roots extend to more than twice this depth (Chorley et al, 1984). It might therefore be supposed that root growth is a major means of physical disintegration but in practice the actual force exerted by roots is difficult to evaluate as the overall role of roots involves two contrasting activities. Firstly they act as a stabilizing agent by binding weathered materials together and thereby retarding the exposure of fresh rock. In contrast they occasionally have a very disruptive effect such as when large trees are blown over.

The significance of colloidal plucking (gelatine drying in a tumbler is capable of detaching small flakes of glass) in the field has yet to be ascertained. It may be an important means of rock disintegration as a considerable fraction of soil organic material exists in colloidal form.

However it is possible that whilst effective on concave materials it does not operate on the convex form of the typical rock grain (Rice, 1977).

Chemical Weathering: Biological agencies affect chemical weathering by affecting the rates of processes e.g. the control on the quantity and quality of percolating water exercised by vegetation. The vegetation cover will regulate the amount of precipitation reaching ground level by interception. It will also influence the rate at which water moves through soil horizons by the production of humus (Rice, 1977). Maximum solution, weathering and leaching are usually associated with high intensity and high frequency of precipitation except where the surface is protected by layers of tall vegetation. Under the latter conditions leaching losses to rivers may be generally small (e.g. in undisturbed tropical rain forests) but otherwise they are at a maximum in hot and wet conditions. Temperate biohomes have generally small leaching losses, except for calcium (Chorley et al, 1984).

The most fundamental effect of organic activity on chemical weathering is the influence on water quality with regard to the supply of carbon dioxide. Oxidative bacteria decomposing organic residues, together with the respiration from plant roots, regularly raise the carbon dioxide concentration in soil air to between 0.2 - 2.0%. The actual amount of carbon dioxide varies in response to the following factors;

1) Temperature - bacterial action declines rapidly as the soil temperature falls below 10°C,

2) Water content - organic activity falls significantly when the moisture content sinks to less than 10%,

3) Soil aeration,

4) Vegetation cover.

Carbon dioxide has important consequences for the solubility of such weathering products as the iron and aluminium compounds.

Biochemical Weathering: Organic material operates within almost all weathering zones to produce a very complex set of biochemical processes including cation root exchange, reduction, chelation and production of organic acids. By these means weathering rates may be increased up to ten times those of distilled water by weak complexing acids (e.g. acetic) and up to one hundred times by strong complexing acids (e.g. citric and tartaric).

Many anaerobic bacteria obtain part of their oxygen by reducing iron from the ferric to the ferrous form. One consequence is to produce ferrous compounds which are significantly more soluble in water. This is a major method by which iron can be mobilized and removed from the soil (Rice, 1977).

Chelation involves the union of metallic cations with the hydrocarbon molecule and is a fundamental process in sustaining plant life. The roots are surrounded by a concentration of hydrogen ions which can exchange with the cations of Ca, Mg, K etc. in adjacent minerals through clay colloids. These metallic cations are then absorbed into the plant by chelation. The resulting so-called co-ordination compound is soluble in organic solvents but not in water. Copper concentrations of 3 - 10% in a peat from southeast New Brunswick have been attributed to chelation of copper (Fraser, 1961). This peat contains 60 - 80% organic matter.

Microbiological degradation products include many organic acids, themselves very effective in mobilizing metal cations. The role of organic acids in weathering is well known e.g. tartaric acid dissolves both silicate minerals and clay many times faster than deionized water; clay minerals faster than silicate minerals and Al^{3+} more rapidly than Si^{4+} in clay. Humic acids (e.g. $C_{40}H_{24}O_{12}$) are active in chelation and decompose silicates (especially amphiboles). Microfloral acids (e.g. oxalic and citric) from fungi and lichens attack silicates and clays and produce carbon dioxide. Bacterial acids (e.g. lactic and acetic) attack a wide range of minerals including magnesium carbonates, calcium and magnesium silicates, feldspars and kaolinite (Chorley et al, 1984).

Sources of Organic Matter.

At present marine organisms contribute approximately 60% of the total organic productivity donating mainly proteins, lipids and carbohydrates (from phytoplankton). Terrestrial plants make up the majority of the remainder giving resins, waxes, lignins and carbohydrates in the form of cellulose. Rates of primary productivity in aquatic environments are affected by the amounts of light present as photosynthesis takes place in the upper sixty to eighty metres of the water column. The rate of photosynthesis is at a maximum in areas of upwelling and near river mouths where abundant nutrients are available. Estimated organic productivity values are as follows (Krey, 1970): open marine waters average 50g carbon/sq.m./yr; coastal waters average approximately 100g carbon/sq.m./yr; western margins of continents with maximum values which may reach 300g carbon/sq.m./yr. However it is considered (Barnes et al, 1984) that preservation of organic matter is probably more important than productivity in determining the organic contents of sediments.

Organic Matter as a source of metals.

Biological studies have shown that metals bonded to certain organic ligands are present in all living matter and unusual accumulations may occur (Peterson, 1971). It has been stated that the total content of Co, Cu, Mn, Mo, Ni and Zn in plants alone is greater than in orebodies (Boychenko et al, 1968). Micro-organisms can concentrate elements from sea water with concentration fractions of up to 10^{6} (Trudinger and Bubela, 1967). When organisms die some of these metal-organic compounds resist bacterial action so producing an association of biogenically derived metals with carbonaceous matter in the consolidated sediment. The preference of metals for certain coordinating groups is evident (Saxby, 1976). In general carboxylate groupings form their strongest bonds with Ca^{2+} and Mg^{2+}; nitrogen atoms favour Fe^{2+}, Co^{2+}, Cu^{2+} and Zn^{2+}, while sulphur ligands prefer Cu^{2+} (Livingstone, 1965). A complexed metal may sometimes assist in the coupling of different ligands e.g. porphyrins and amino acids (Hodgson et al, 1970). During diagenesis of the sediment most metal-organic compounds will decompose. Eventually the metal content in the sediment will be more concentrated than in the original living matter with the release of metal as organic matter is lost as CO_2, volatile hydrocarbons or soluble organic compounds. The porphyrin nucleus is very stable and does not decompose at normal sediment temperatures (Saxby, 1969), although the original metal has generally been replaced by V or Ni to give a more stable complex. The resonating ring of chlorophyll is stable until extreme conditions are encountered (e.g. temperatures greater than 500°C) but ring substitutes are probably lost or hydrogenated and the Mg atom is replaced (Saxby, 1976).

Preservation of Organic Matter.

The preservation of organic matter is a function of the oxygen content of the environment. Estimates of the preservation of primary organic matter in the surficial sediments of marine environments average about 0.1% (Menzel and Ryther, 1970). This value is even lower for subaerial environments as the high oxygen content of the air favours chemical oxidation and aerobic microbial decomposition. The preservation of organic matter in regions of rapid terrestrial sediment deposition (e.g. deltas and continental slopes and rises) is relatively high because of the limited diffusion of oxygen into the sediments (Barnes et al, 1984). Generally organic material in rivers is broken down by oxidizing bacteria using atmospheric oxygen. However if the organic content is too great it accumulates on the river bed and the water becomes anoxic. Further decomposition occurs if sulphate-reducing bacteria use SO_4^{2-} in the water as an oxygen source. Hydrogen sulphide results. The maximum preservation (about 4%) occurs in anoxic environments (Deuser, 1971) where anaerobic microbial processes result in less complete decomposition of the organic matter. The products are again subject to oxidation during bacterial sulphate reduction. It can be concluded from this that rapid sedimentation limits exposure to oxygen and thus favours the preservation of organic matter. Terrigenous organic components are also better preserved than those from aquatic sources (Meyers et al, 1984).

Concentration of Metals by Organic Matter.

Complex Formation: A wide range of range organic molecules e.g. amino acids, alcohols, heterocyclic compounds containing nitrogen and sulphur are capable of binding metals through co-ordinate bonds involving nitrogen, oxygen and sulphur atoms are present in many aqueous environments. The presence of metal-organic componds in hydrothermal ore forming solutions has been verified by fluid inclusion studies (Kranz, 1968). Baker (1973) suggested another mechanism for the formation of metal complexes in a solution. This involved the attack on a particular mineral by percolating waters containing a specific organic ligand. As mentioned earlier humic acids are effective agents in the weathering and transport of certain metals e.g. Cu, Al, Fe, Zn and Pb (Ong et al, 1970).

Physical Adsorption: Physical adsorption is a weak force by which metallic ions are attracted to soluble, colloidal or particulate organic material, so that the ion is easily replacable. It has been shown that it is possible to concentrate large quantities of metals on certain naturally occurring organic materials. Studies of the reactions of Cu^{2+}, Pb^{2+} and Zn^{2+} in aqueous solutions with organic matter derived from fresh samples of green filamentous algae show that under suitable conditions, a significant proportion of the metals is removed from the solution by sorption onto the particulate organic matter of the algal suspension (Ferguson and Bubela, 1974). They found that the presence of relatively small amounts of Na^{+} and Mg^{2+} in solution reduces the sorption to an appreciable extent in even strong brines. This may be a means for the selective precipitation of Pb^{2+} from brines rich in Pb^{2+} and Zn^{2+} ions.

Chemical Precipitation: A third mechanism of metal concentration during sedimentation involves the reaction of carbonaceous material with metals or metal complexes dissolved in percolating groundwaters or hydrothermal fluids. In particular if the metallic ion can be reduced and precipitated as an insoluble species, an association of metal with carbonaceous material will be achieved, at least until the latter has been oxidized. Such a process can account for many carbonaceous U deposits where it has been suggested that U in soluble U^{6+} complexes in groundwaters is precipitated as a result of reduction to U^{4+} by carbonaceous shales (Saxby, 1976).

Diagenesis of Organic Matter.

The low temperature shallow burial diagenesis of organic matter is known as eogenesis (Barnes et al, 1984). Eogenetic changes greatly affect the physical structure and chemical stability of organic matter and influence the formation of humic compounds and hence form metal complexes which act as metal transport agents in groundwater systems. The processes are also influenced by a change in the Eh of pore waters which abruptly decreases from the oxic to the anoxic zone and is accompanied by a slight increase in the pH. This results in an enrichment of bicarbonate and ammonia and the loss of carbon dioxide leading to the precipitation of carbonates and the dissolution of amorphous silica. These reflect the breakdown of organic matter and very general activity of anaerobic micro-organisms. Iron, copper, lead and zinc may be precipitated as sulphides.

1) Generation of sulphides: Organic matter contributes to the concentration of metals but its role in generating sulphide ions (Krouse and McCready, 1979) which are vital to the formation of sediment-hosted base-metal sulphide mineral deposits is very important. At low temperatures sulphate-reducing bacteria can produce hydrogen sulphide by reaction of other iron minerals and reduced sulphur species. The latter is formed in the initial step:-

$$SO_4^{2-} + 2CH_2O \longrightarrow HS^- + HCO_{3-} + H_2CO_3.$$

The factors which limit the amount of pyrite development are the amount of organic matter that can be metabolized by sulphate reducing bacteria; the available iron compounds that are within the sediment and the maximum rate at which sulphate can diffuse into sediment from the overlying depositional water (Berner, 1970). Typical conditions for the existence of sulphate-reducing bacteria are: Eh (+350 to -500mV), pH (4.2 - 10.4), pressure (1 - 100 atm), temperature (0 - 100°C), salinity (,1 - 30% NaCl), presence of an energy source and trace elements (Ca, Mg, K, Fe, P, Cl, N) (ZoBell, 1963). The formation of completely biogenic sulphide ores has been questioned from a consideration of the rates of sulphate reduction, organic matter production and ore deposition (Rickard, 1973).

2) Fermentation: This process occurs whenever sulphate-reducing bacteria are inactive e.g. in anoxic fresh-water sediments or in rapidly deposited marine sediments below the depth of diffusive sulphate penetration. It is limited to the availability and amount of utilizable organic matter (Curtis, 1977). The common product of fermentation is methane.

3) Carbonates: Oxidative bacteria close to the sediment-water interface, sulphate-reducing bacteria (down to a few metres) and fermentation bacteria (below the level of complete sulphate depletion in more rapidly deposited sediments) all produce bicarbonate. This is found in the diagenetic carbonates calcite, dolomite, ankerite and siderite. Shallow burial

environments therefore must be considered as likely locations for recrystallization and precipitation of various carbonate minerals with carbonate being supplied by biogenic primary carbonates and bacterial degradation of organic matter (Curtis, 1977).

Organic Matter in Sabkhas and Evaporites.

The inward flow of waters into an evaporite basin is very rich in biota whereas the bottom brines are hostile to life forms. As the brine becomes more concentrated, the number of species which can survive decreases and there is a concurrent decrease in oxygen solubility. This leads to an elimination of bottom-dwelling forms and anaerobic bacteria proliferate. The organic matter which now enters the evaporite basin is not oxidized and so accumulates in the brine. Ultimately it attaches to precipitates, is deposited in the sediment, or is flushed through a suitably permeable substrate.

Sabkhas are usually bordered by intertidal mudflats and lagoons on the seaward margin which are coated by blue-green algal mats. Beneath the upper surface is a zone of interbedded decaying algae and detrital sediment. Through time sabkhas prograde seaward over these algal mats which become saturated with hydrogen sulphide by anaerobic bacteria upon burial. Ultimately terrestrial formation water must pass upward through these strongly reducing algal mats to reach the evaporation surface (Renfro, 1974). Initially such terrestrial waters are characterized by low pH and high Eh so they can mobilize and transport metals such as Cu, Ag, Pb and Zn. However when the waters pass through the hydrogen sulphide enriched algal mats these metals are reduced and so precipitated as sulphides. There is a zonation in the suite of metals from landward to seaward according to their relative solubilities in the presence of hydrogen sulphide.

More recently, Long et al (1985) have found that modern marine sabkhas do not have elevated trace-metal concentrations compared to recent near-shore marine sediments and are therefore not active loci of trace-metal enrichment. However they do acknowledge that localized metal accumulation in the sediment can be detected. This is again centred on the cyanobacterial mat associated with evaporite sediments which are thought to play an important role in the initial concentration of metals such as Zn, Cu and Pb. During burial and subsequent diagenesis the metals can be reconcentrated (Lyons and Gaudette, 1985). The lack of metal enrichment in sediments of modern marine sabkha systems has been attributed by Long et al (1985) to either the relatively young geological ages of these environments hence diagenetic processes have not yet concentrated metals or the intervention of a specific geochemical process e.g. the continual extraction of the metals from the sediment by subsurface brines.

In general the biota within a concentrated brine in an evaporite basin accumulate some metals in their bodies and release them upon decay. Such metals may initially enter the crystal lattice of evaporite minerals but are removed by recrystallization and then travel into the subsurface with migrating hypersaline brines.

Conclusion.

From the previous sections the importance of climate to mineral deposit development is undeniable. It has been shown that climate affects the degree of weathering, the amount of available organic matter for reduction, evaporites for a sulphur source and the volume and chemistry of potential metal-transporting fluids.

There are two main varieties of sediment-hosted mineral deposits which are under examination here; syngenetic (primary) and epigenetic (secondary) deposits. The influence of climate on mineral deposit development may also be divided. For syngenetic deposits it appears that syn-ore climate is a vital factor e.g. for adequate weathering of source rocks, release of metals and transport in an adequate solution. However for epigenetic deposits, both syn- and pre-ore climates may be influential. Pre-ore climatic conditions would dictate the products of such a climate in sediments which may control the later processes e.g. influence depositional conditions (the presence of evaporites, organic matter). However syn-ore climate affects other factors which are influential in ore development such as quantity and quality of groundwaters for metal transportation.

Climate is only one of many possible enhancements (e.g. initial metal source, metal enrichment, suitable depositional environment) that must occur to allow mineral deposit formation. In some instances the climate may have appeared to have been suitable, but no mineral deposits have developed because another of the necessary enhancing mechanisms did not function. An example of the necessity for a combination of enhancements was inadvertently stressed by Ferguson and Bubela (1974). Their results suggested that certain types of particulate algal matter could sorb sufficient quantities of metal to form an ore deposit only if the weight of organic matter available was of a similar order of magnitude to that of the inorganic sediment in the deposits. However as metal sorption is an equilibrium reaction, the metal "saturation" values could be reached only in solutions with initial metal contents at least two orders of magnitude above those of seawater. The magnitude of these enrichment factors indicated that the saturation values could be achieved only in solutions already enriched in metals. So regardless of other factors that may be involved (e.g. climate) if the initial solution was not carrying sufficient metals, or the depositional environment was not sufficiently rich in organic material, then no accumulations of metals could occur.

8.2 Individual Deposit Types Discussed.

8.2.1 Carbonate-hosted Deposit Types.

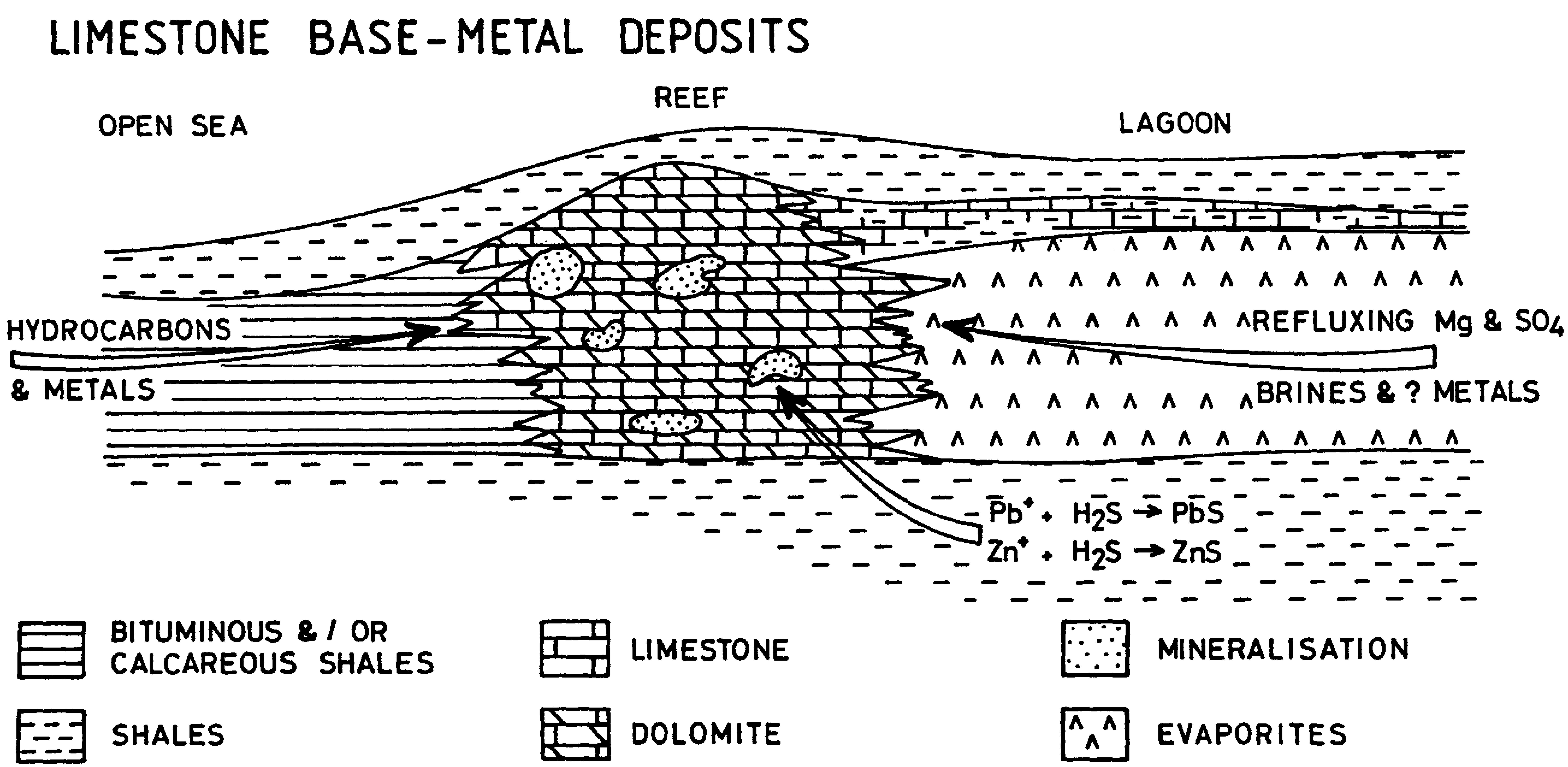
8.2.1.1 Limestone Base-Metal Deposits (LSBM).

The main characteristics of LSBMs have been given in Chapter Three, section 3.2.7 and so are not discussed here. However the genesis of these deposits should now be mentioned in order to introduce the possible influence of climatic conditions upon their formation. There are three aspects of the origin of LSBMs which have been the centre of controversy for many years. Firstly, the ore is frequently found in close association with dolomitization but it is not known how closely these two events are related in time i.e. was the dolomitization caused by the precipitating ore fluids or did the increase in porosity associated with dolomitization simply prepare favourable pathways for later, unrelated ore deposits? Secondly was the ultimate source for the metal-bearing brines the adjacent sedimentary basin to the deposits i.e. due to fluid expulsion of basinal shales? Alternatively they could be surface waters that acquired salinity as a result of evaporation during a period of hot, semi arid climatic conditions and/or by contact with evaporites in the lagoonal facies of LSBM deposits. Lastly the dilemma is whether or not the metals and the sulphur for reduction were transported to the depositional site in the same fluid. The problem is that it is not possible to transport metal chloride complexes and reduced sulphur species in the same solution as they would react. Hence it is thought that two independent fluids using different pathways must meet and mix in order to precipitate sulphide minerals.

Hoagland (1976) listed seven essential requirements for the formation of LSBMs which included such factors as the coincidence of time and space i.e. the necessity for a two-fluid system to mix within the channels for a sufficiently long period of time to produce such large LSBMs as found today, and conditions which could produce and sustain a hydraulic gradient capable of driving metal-rich brines from source area to the discharge area. The first point i.e. mixing of two fluids has very little to do with climate, although the second i.e. hydraulic gradient and heat flow may be greatly influenced by climatic enhancements (see section 8.1.3). The different facies of LSBMs are discussed below in terms of the possible influence of climate on certain features (see Figure 8.9).

The Open Sea/Basinal Facies.

The conditions in the basin which may be affected by climate are the low clastic input (so shales with a high clay mineral content dominate), the clay type itself (see section 8.1.3) and high biological productivity. The latter increases the organic content of the associated rocks which may also influence the metal content of the shales and hydrocarbon development. The ore fluid for LSBMs is generally considered to be a chloride-rich saline brine at 50° to 150°C (Roedder, 1976) expelled from the basinal shales (Hoagland, 1976; Dunsmore and Shearman, 1977). Such waters may become enriched in heavy metals if they have a sufficiently high chloride content (perhaps from the dissolution of evaporites) and basin geometry is such that sufficient quantities of water are able to flow through the shales to allow a high degree of metal release by solution. These metal-bearing brines have been reported from sedimentary basins (Carpenter et al, 1974) with 15% NaCl, and containing 100ppm Pb and 350ppm Zn at 90° - 160°C. Long and Angino (1982) have also noted substantial leaching of metals from shales by brines. However Ohle (1980) questioned that shales were the major metal source arguing they were very impermeable when lithified. He also doubted that their relatively high metal content could be made available once the initial flush of water had occurred.



CLIMATIC ENHANCEMENTS

OPEN SEA	1. Low clastic input, high biologic productivity promotes metal rich shales & hydrocarbons
REEF	2. Thick reef growth is climatically dependent because of temperature, light & low clastic input
LAGOON	3. Evaporite formation is temperature dependent
LAGOON	4. Metal content high with hot chemical weathering of hinterland

Figure 8.9. Idealized model of the formation of LSBM deposits showing possible sources of brines and base-metals and their transport to the site of deposition.

The utilization of waters expelled during compaction of underlying sediments to move brines upwards to suitable depositional sites was proposed by Noble (1963). However this was questioned (Lange and Murray, 1977) on the grounds that underlying rocks may have become compacted by the time of mineralisation e.g. the southeast Missouri Lead district (Snyder and Gerdemann, 1968). Bethke (1986) argued that compact-driven flow may be a dominant ore-forming process in districts associated with shaly, rapidly subsiding basins (e.g. Ouchita Basin) because compaction-flow would be rapid and over-pressuring common. However he suggested that in other instances e.g. the Upper Mississippi Valley district, gravity-driven flow would be more effective because of the moderate flow velocities of metres per year through deep aquifers, permeability structures and high heat flow. Compact-driven flow in this area would not be effective in transporting heat within the basin because migrating fluids moved too slowly to avoid conductive cooling: the basin was not over-pressured during subsidence because of low burial rates, low shale content and the presence of deep aquifers. So if the theory of basinal shales as the origin of metal-bearing brines is correct the climatic influences mentioned at the onset are obvious. However even if alternative sources are sought climate may still be influential (see below).

The influence of climate on biological productivity has been discussed previously so the possible genetic roles of organic matter in the formation of LSBMs are to be examined. They provide a source for metals in their initial concentration as metal-organic complexes which are then released during thermal maturation. Organically bound sulphur (0.2 - 2%) may also be utilized in metal sulphide formation. A proportion becomes available as H_2S through bacterial attack during consolidation of the sediment. The residue is gradually released during thermal breakdown of the buried organic matter (Brooks, 1971). Skinner (1967) proposed that hot brines carrying metal chloride complexes caused release of H_2S by thermal degradation of sulphur-containing organic compounds. This may be less important than bacterial sulphate reduction but still accounts for the metal sulphide-organic matter association seen in LSBMs. The relationships between bitumen and a sulphide-bearing barite deposit in the Lodève Basin, France have been examined (Connan, 1977). It was concluded that the degree of bacterial biodegradation of bitumens was linked to the abundance of sulphides (e.g. galena and bornite) in the vicinity. In this case biodegradation by sulphate-reducing bacteria was regarded as the sulphur source for the deposit. The transport of metals as organic complexes might assist in a reduction of large volumes of fluids required to develop these deposits (Anderson, 1975) by alleviating the low solubility of chloride complexes (Macqueen, 1979).

The association of many LSBMs with hydrocarbons has been frequently mentioned (e.g. Dozy, 1970; Macqueen and Thompson, 1978). These hydrocarbons may provide an important source of hydrogen sulphide i.e. supplied directly by reduction of sulphur-bearing compounds such as resins or asphaltenes in the petroleum (Macqueen, 1979). An alternative source of H_2S is from sulphate-reducing bacteria which feed on petroleum, oil or bituminous sediments and which produce continuous volumes of H_2S (Jackson and Beales, 1967). Such bacteria can be active up to temperatures around 80°C (Dunsmore and Shearman, 1977).

The Reef/Carbonate Facies.

One of the most consistent features of LSBMs is the ubiquity of limestones as hosts for the mineralisation. In most deposits dolomites are of regional importance within the mining district and in a few cases e.g. the Tri-State district of east Tennessee, dolomites are restricted to the vicinity of the ore body (Sangster, 1983). Alternatively the ore may be restricted to the environment of dolomitization as mentioned earlier.

The formation of carbonate sediments is enhanced by certain climatic conditions e.g. warm sea water, so it is to be expected that a plot of the LSBM deposit host rocks shows a preference for low latitudes (Dunsmore and Shearman, 1977; this thesis). As the warmer climates of low latitudes also encourage the development of reefs, an association between reefs and LSBMs is to be expected too. The formation of carbonates is considered to be limited by factors which influence the development of hermatypic corals and calcareous green algae which are among the most important carbonate contributors (Milliman, 1974). Both these organism types depend on light for photosynthesis hence clarity and depth are considered important. However temperature is also often assumed to be a limiting factor because present hermatypic reefs do not occur poleward of the 18°C minimum isotherm (Wells, 1957). Ultimately the above are controlled by insolation and the refractive index of sea water. Ideally in low latitudes, 100% of the radiation is refracted into the sea, above 30° of latitude the amount of refracted light diminishes rapidly and above 65° all light is reflected back into space. The carbonate belt does not expand poleward during warmer intervals (e.g. the Cretaceous) which suggests that light refraction, and not temperature, is the limiting factor for carbonate environments (Ziegler et al, 1984). This is obviously a latitudinal phenomenon and some climatic conditions as a consequence of latitude may also reduce the effective radiation in a number of ways. Firstly, surface roughness (in the form of waves) influenced by prevailing winds and cloud cover combine to reflect about 20% of the low latitude radiation. Also turbid conditions associated with clastic run-off dominate the western sides of oceans in tropical and subtropical zones, so carbonates are resticted to shallow shelf margin sites. On the eastern side of oceans, turbid conditions associated with upwelling and high surface productivity restrict carbonate formation to protected inner shelf situations (Ziegler et al, 1984).

The role of carbonates in LSBM development is manifold. Carbonate rocks are often enriched in hydrogen sulphide which is an effective metal sulphide precipitator (Macqueen, 1979). They are also often rich in organic matter, whose uses have been previously discussed in full and whose presence may also enhance the H_2S content, or the organics may serve as a metal source. Perhaps the most important role of carbonates is that concerning their porosity. Hoagland (1976) stressed that a carbonate host rock environment with sufficient open space to permit the necessary volume of ore minerals to be deposited was essential to LSBM formation. Carbonate sequences are usually lithified early with a high initial porosity and permeability e.g. crinoidal bioherms (Hagni, 1976). These zones are usually connected to both the basinal shales (fore-reef zone) and the evaporites in the back-reef area and can thus easily accomodate products of their greater compaction (Badham, 1981b).

The development of secondary permeability and porosity during dolomitization or karstification may be even more important than the initial porosity of carbonates. Prominent erosional surfaces which represent extensive karstification characterize all major LSBM deposits. These usually occur less than a few hundred metres above the ore body with the exception of southeast Missouri district (Sangster, 1983). The probable significance of these features is that they have provided sites of high permeability for the passage of metal-bearing brines. Hagni (1976) noted that part dissolution of limestone originated through groundwater action associated with the development of karst topography on the post-Mississippian, pre-Permian erosion surface of some Tri-state deposits. This subsurface groundwater activity aided in the development of additional solution-related fractures which facilitated the introduction of subsequent mineralisation solutions. Kyle (1981) also described palaeo-solution structures related to subaerial exposure of the reef barrier complexes in the Pine Point deposit which were important ore hosts. These features were also thought to have been caused by dissolution of limestones by meteoric water and which were then filled with dolostone by the mixing of fresh and sea water during a period of subaerial exposure.

Where leaching of carbonate beds has occurred it appears to have taken place prior to, or during, ore deposition (Heyl, 1983). In some cases this has resulted in sagging and collapse of the overlying beds thereby further increasing permeability. Brecciation and open-space filling is a very important process in LSBM formation, where solution of carbonate by ore-related brines has allowed collapse and brecciation of the overlying beds (Ohle, 1985). The reef environment is therefore particularly important because it influences the porosity of the host rocks and may act as a source of metals, organic matter and hydrogen sulphide. Thick carbonate/reef growth is obviously climatically dependent because of light penetration, water temperature and a low clastic input.

The Lagoonal Facies.

Dunsmore and Shearman (1977) suggested that the proximity of evaporitic sediments to reef carbonates was a prerequisite for LSBM formation. Although this has not been proven as some deposits do not show such an association, there are a number of advantages to LSBM development because of the presence of evaporites. If evaporites are an integral part of LSBM development then the absence of any evidence of leached evaporites in some deposits requires an explanation (Ohle, 1980). Evaporites may provide large volumes of refluxing brines for the carbonates. Lange and Murray (1977) proposed that tongues of very dense hypersaline brine seeped downward by reflux from over-lying evaporite deposits and displaced any existing pore fluid. This mechanism required a two stage evaporite development. The first stage provided the deep brine which became heated and initially dissolved the metals whilst the second stage provided the refluxing brine. These evaporite brines may contain Mg to aid dolomitization and hence porosity (Badham, 1981b). Carpenter et al (1974) suggested that chloride-bromide and potassium-bromide relationships in brines of the central Mississippi region indicated that they originated from evaporation of sea water past the point of halite deposition. The brines probably originated as interstitial fluids in the Louann Salt and were expelled upwards as a result of loading by younger sediments.

Another role for evaporites in LSBM development may be as a source of base metals, particularly if they are interbedded with shale deposits (Badham, 1981b). Olade and Morton (1985) proposed that the most probable source of mineralising fluids for the Benue Valley, Nigeria was the evaporitic shale deposited contemporaneously with, or prior to, the main mineralising event (Ford, 1981). Thiede and Cameron (1978) provided evidence that certain components in an evaporite sequence contain appreciable Cu, Pb and Zn contents. The possibility that evaporites may be an important metal source and an alternative to basinal shales, is supported by the highly saline nature of LSBM fluid inclusions. However there is little doubt that evaporites are a likely source for the sulphur to precipitate ore metals as sulphide (Dunsmore and Shearman, 1977; Kyle, 1981). The lagoonal facies of LSBMs indicates that a climatic influence may be of importance in that the formation of evaporites is particularly temperature dependent as evaporation must exceed precipitation. Also if they are a metal source, then the metal content of the lagoon environment must be high due to hot, chemical weathering of the hinterland e.g. ore metals for the Benue Trough, Nigeria are thought to have been leached from alkali feldspars from surrounding regions (Grant, 1971).

8.2.1.2 Oolitic Ironstone Deposits (OOFE).

The main point of discussion concerning the genesis of OOFEs is the relationship of oolites associated with an agitated, well-oxygenated environment in tropical climates with iron which is usually expected to be deposited in quiet, low-oxygen water. OOFEs are not environmentally diagnostic so an interpretation of their depositional environment must be based upon fauna and any notable sedimentary structures. A low diversity of fauna composed of large numbers of individuals often indicates abnormal salinities e.g. Ostrea at the present day is found in lagoons and enclosed bays which are kept at moderate salinities (>30ppt) by tidal renewal (Parker, 1960). The great abundance of Deltoideum and the scarcity of other fossils in the OOFEs in the Upper Oxfordian of southern England is suggestive of the restricted conditions under which these deposits were formed (Talbot, 1974). The presence of belemnites and ammonites also shows these OOFEs must have been connected to the open sea, whereas other lithologies which occur (e.g. a muddy sediment with much burrowing and little turbulent reworking) is in keeping with deposition in a calm, sheltered environment. The occurrence of an Aponogeton-like plant impression in beds immediately above OOFEs at Bida in the Middle Niger Valley was considered by Adeleye (1973) to confirm tropical conditions during OOFE sedimentation as its modern representatives are restricted to tropical and subtropical zones.

No general model has been developed for the formation of OOFEs which explains fully all the various features shown by these deposits. This may be due to an absence of modern analogues. The problem has been approached in two ways; the origin of the ooliths and the possible sources of the iron.

Sources of the Iron.

The main problem of iron source and transport is based upon the fact that the Fe^{3+} ion is insoluble and the Fe^{2+} ion is soluble at Eh and pH conditions of the Earth's surface. Hence Fe occurs as a residue in weathering profiles (e.g. as laterite) or is transported clastically and so is unavailable for oolites.

It is unlikely that the iron was derived from sea water in which its concentration is generally very low at present i.e. less than 0.01ppm. There is no obvious spatial or temporal association of volcanics and OOFEs so volcanism is also not thought to have played a direct role in the origin of OOFEs and an alternative source of the iron is required. One of the most popular hypotheses is Fe may be derived from adjacent land areas by the normal processes of weathering and erosion (e.g. James, 1966; Hallam, 1975). Indeed Lunar and Amoros (1979) suggested that iron for the OOFEs of northwest Spain may have been derived from the continent. Adeleye (1973) proposed that the prevailing tropical climate in southwest Nigeria caused intense chemical weathering of the igneous and metamorphic basement complex which supplied iron to the OOFEs and that some secondary ferruginous matter may have been derived by leaching of older, alluvial swamp deposits of the Middle Niger Valley. The released iron and its associated terrigenous clasts were considered to have been eroded from source by normal fluviatile processes. The Upper Calcareous Grit Fe-bearing sediments of southern England are thought to have been deposited in locally restricted water bodies of reduced salinity which suggests that the iron was indeed derived from rivers draining adjacent land masses (Talbot, 1974).

Weathering of some rocks, particularly under humid conditions may lead to the liberation of large quantities of iron which accumulate as insoluble residues in oxide and hydroxide form e.g. laterites. If it is assumed that transport of iron to the depositional site is mainly by river water, two possible mechanisms can be proposed (Talbot, 1974). Firstly Carroll (1958) suggested that large quantities of iron could be carried as iron oxide adhering to the surface of clay minerals. Such oxide-coated clay minerals are very common products of subaerial rock weathering and soil formation. Ferguson et al (1983) proposed that ferric oxyhydroxide coatings on quartz grains in the aquifer system explained the high iron concentration in ground water and low iron contents in sediments at the same level in Fisherman's Bay, South Australia. The second mechanism is to transport the iron as colloidal suspensions in solutions. The quantity of organic matter available probably plays an important role in this process as large amounts of iron may be carried as, or adsorbed by, organic colloids (Gross, 1965). Boyle et al (1977) noted that such iron oxide-organic matter colloids were rapidly precipitated in estuaries and are thus in accordance with the model of semi-restricted, near shore basins for OOFE development. Such suspensions would best develop during chemical weathering of a land mass of low relief with dense, swampy vegetation cover and a humid tropical or subtropical climate. According to some models of sedimentation (e.g. Muller and Forstner, 1973) the formation of large iron deposits requires a preconcentration of iron on the land before transportation of the continental iron materials to the basin occurs. Such a preconcentration can be related to the leaching out of iron during the development of a podsol. Muller and Forstner (1973) considered this kind of regime would not require a well developed flora when combined with climatic factors that produced a slow decomposition of organic matter.

There are two possible ways in which the transported iron may then be released to form OOFE deposits. Firstly the Fe^{3+} particles may be transported to the basin where upwelling of anoxic basinal waters onto the aerated shelf caused reduction during diagenesis and the Fe^{2+} ions are released (Borchert, 1960). This reduction of the Fe^{3+} may also be caused by downward moving ground water from overlying organic-rich muds (Kimberley, 1979). An alternative method of making Fe available is by weathering of Fe minerals and transport of Fe^{2+} to the basin by ground water under abnormally low pH conditions. The source of Fe is clearly problematical. However which ever of the solutions given above ultimately proves to be the most appropriate, all are potentially influenced by the prevailing climatic conditions at the time.

The Origin of the Ooliths.

One hypothesis for the origin of ooliths is that they are detrital, being eroded soil ooliths and pisoliths from lateritic terrains (Siehl and Thein, 1978). Nahon et al (1980) described OOFEs from West Africa and suggested they were formed by lateritic weathering of sedimentary rocks, especially glauconite-rich sandstones based on similarities with ooids characteristic of exposure crusts in carbonate rocks (Carozzi, 1973). These deposits have been shown to be related to the so-called "Continental Terminal" which is considered to be the product of the lateritic weathering of Mesozoic-Cenozoic marine sediments (Tessier et al, 1975). These OOFEs with an Fe-rich argillaceous matrix described above would represent palaeoexposure crusts in a tropical climate. However these conclusions may be difficult to justify as many ooids formed on skeletal parts of marine organisms.

Kimberley (1979) revived the hypothesis that the oolites were originally calcareous. They were initially composed of aragonite and were replaced by downward-percolating meteoric waters to chamosite and goethite. The third theory of oolitic origin is that chamositic ooliths formed in relatively quiet water (Knox, 1970) whilst goethitic ooliths were formed in more agitated waters (James and Van Houten, 1979). Alternating bands of chamosite and goethite have been explained by reworking of ooliths from one depositional environment to another and back again (Maynard, 1983).

Possible climatic conditions at the time of OOFE formation have been mentioned. Lunar and Amoros (1979) suggested that the fine-grained sediment texture and mineral composition of the deposits of northwest Spain indicated they were accumulated in a cold, humid climate with a predominance of chemical over mechanical alteration. The semi-arid climatic regime of the

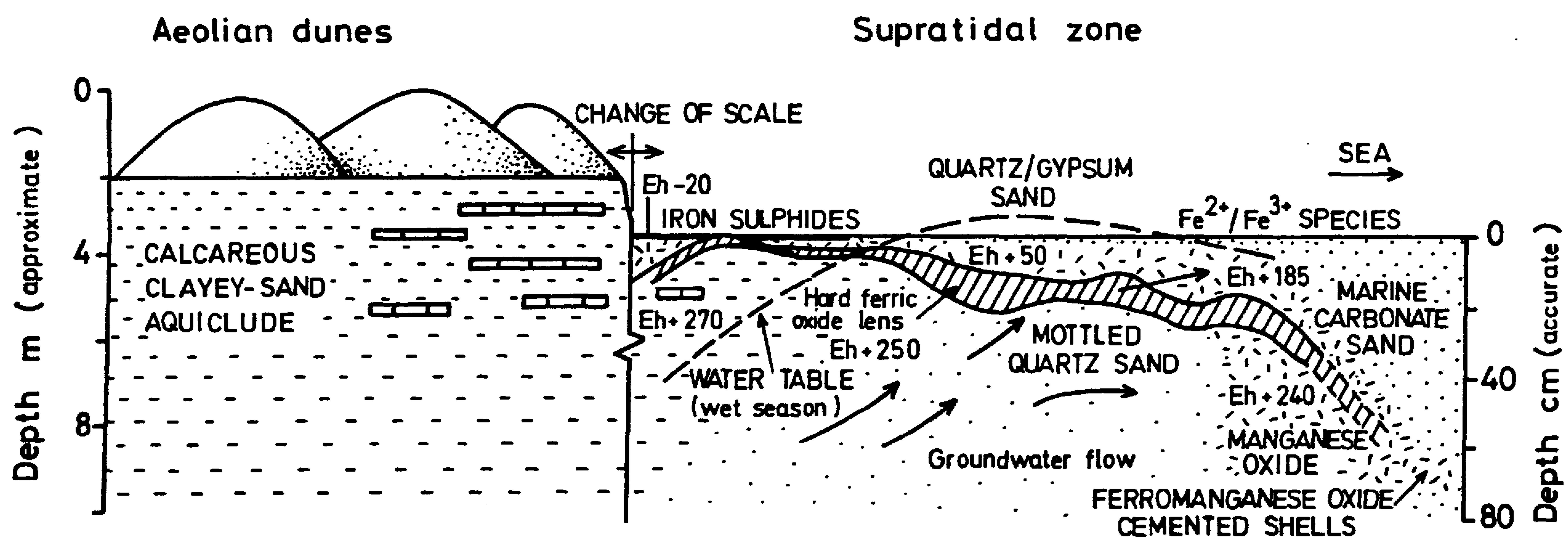


Figure 8.10. Schematic cross-section of the iron mineralisation, Fisherman Bay, South Australia. Note the relationship of the iron lens to incoming groundwaters, the underlying Pleistocene and overlying marine carbonate sands. (After Ferguson et al, 1983, Figure 3c).

Spencer Gulf, South Australia greatly influences the OOFE deposit development of Fisherman's Bay (Ferguson et al, 1983). Surface run-off from the coastal plains is severely limited so drainage is mainly subsurface. Seaward groundwater flow therefore occurs largely through aquifer systems in the surface aeolian sands and the underlying Pleistocene sands and clayey sands (see Figure 8.10). In Fe-rich areas the regressive coastal zone is 1km wide, consisting of intertidal marshes of cyanobacteria. The prevalence of tropical conditions during OOFE sedimentation in the Middle Niger Valley is confirmed by the presence of kaolinitic pellets, oolites and pisoliths (Adeleye, 1973). Recent kaolinitic sediments are almost exclusively restricted to tropical and subtropical zones (Rateev et al, 1969). Talbot (1974) deemed that hot, humid conditions prevailed during the production of Oxfordian OOFEs in southern England with sufficient rainfall to maintain sediment-transporting rivers and to support a land flora. The temperatures had to be high enough to allow hermatypic coral growth and permit the widespread precipitation of calcium carbonate in ooids and grapestones. These climatic conditions would be expected to have promoted deep weathering of land areas and perhaps the development of Fe-rich suspensions.

The derivation of iron from the adjacent land mass and transportion to site of deposition by river water is hinted at by the results shown in Chapter 6, section 6.2.2. OOFEs showed a marked lack in number in the tropical arid zone, suggesting the necessity of surface precipitation for river transport of iron.

8.2.2 Clastic-Hosted Deposit Types.

8.2.2.1 Sandstone-Copper Deposits (SSCU).

Strata-bound copper deposits appear to form a single genetic group which tend to be confined to shallow lagoonal or lacustrine environments. The host rocks are characteristically terrigenous and supra-tidal to sub-tidal sandstones with abundant evidence for fairly broad tidal flats and features of transgressive cycles (Jacobsen, 1975). Evaporites and pyritic shale sediments are frequently associated with the cupriferous sequence, if not actually part of it and red-bed type sediments are found in some areas. Kirkham (1986) suggested that SSCU host rocks were deposited in arid and semi-arid areas (because of interbedding with evaporites) within 20° to 30° of the palaeoequator. He stressed that SSCUs show a close relationship to environment of sedimentation even though most are probably products of diagenetic oxidation-reduction processes. Silver is frequently associated with these deposits (Maynard, 1983) so Cu and Ag are discussed together.

Geochemistry of Copper.

At low temperatures the geochemical behaviour of Cu and Ag is dominated by Eh-pH changes. At surface temperatures in aerated fresh water Cu is soluble in significant amounts (1ppm) only at pH values less than about 6.0 (see Figure 8.11). As Fe is less soluble under these conditions this is an effective mechanism for separating Cu from Fe. Figure 8.11 also shows that under reducing conditions Cu can precipitate as a sulphide or native Cu. Hence Cu will tend to migrate from areas of oxidation and concentrate in areas with reducing conditions. The redox behaviour of Ag is very similar to that of Cu (Maynard, 1983).

The transportation of Cu and Ag relies heavily upon chloride-complexing which greatly enhances the solubility of these metals (White, 1968). Rose (1976) has shown that the solubility of Cu in particular is greatly enhanced by the formation of such complexes as $CuCl_2^-$ and $CuCl_3^-$ (see Figure 8.12). Contact of these chloride solutions with pyrite or H_2S will precipitate various Cu and Cu-Fe sulphides. The possible importance of organic complexing to SSCU genesis and of slightly acid solutions caused by oxidation of

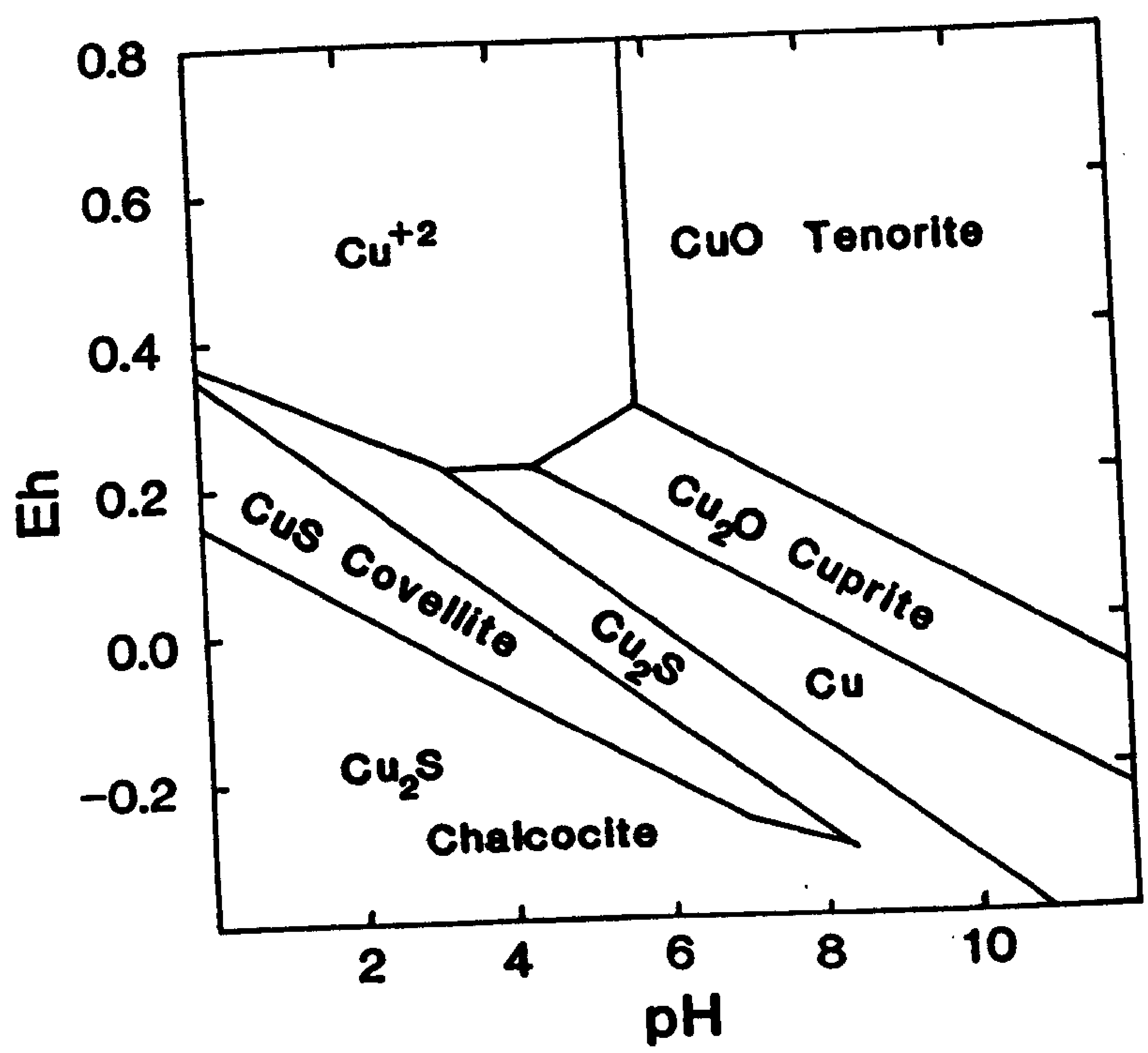


Figure 8.11. Eh-pH diagram for copper oxides and sulphides showing region of solubility under oxidizing conditions at pH less than 6. (After Maynard, 1983, Figure 3.1).

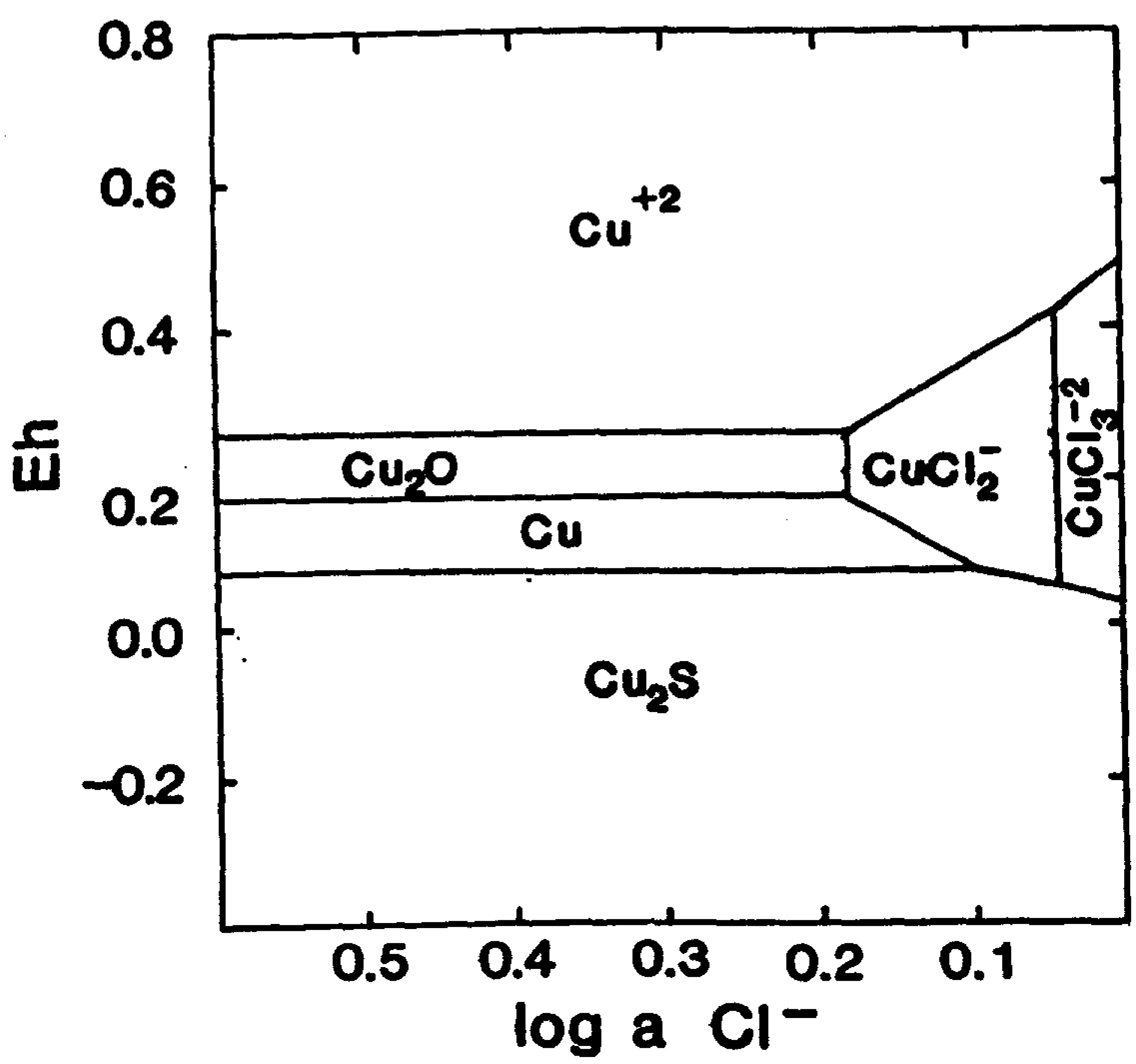


Figure 8.12.Effect of Cl^- complexing on copper solubility.

(After Maynard, 1983, Figure 3.3).

silicates and oxides should be recognized. Also the significance of adsorption of Cu and other heavy metal ions on haematite and colloidal Fe oxides (Rose, 1986). In fact Maynard (1983) noted that the transport of sufficient Cu to form large orebodies in the absence of chloride complexing would be difficult to envisage unless some pre-existing enrichment e.g. supergene alteration of porphyry coppers, was present.

Models of SSCU genesis.

Perhaps the most widely accepted theory as to the genesis of SSCU deposits is that of syn-diagenesis with evaporites. However other models have also been proposed.

The Epigenetic Theories: Ahlfeld (1967) distinguished between ores associated with organic remains (Colorado Plateau) and those ores in red-beds devoid of such remains (Corocoro, Bolivia). He suggested that metals were introduced after diagnesis by circulating saline groundwaters. These were thought to contain $CuSO_4$ in a slightly reducing environment as evidenced by presence of sulphides in cell structures of plant remains. Smith (1976) also proposed an epigenetic origin for SSCUs - in this instance for deposits of North Texas. Cu in the form of chloride complexes was thought to have travelled vertically along faults or fractures and spread out laterally along porous sandstone beds. Deposition may have occurred when mineralising solutions came into contact with H_2S in organic-rich zones e.g. tidal channel-fill facies and algal mat facies. However the main problem with this theory is that no evidence of faulting or fracturing has been detected in the area of the study, unless the faults were of an opening and closing nature in which case they would not be recognized.

A second epigenetic theory involving thermal diffusion and metasomatic processes has also been investigated (Garrels et al, 1949). Micro-porosity may be developed in lithified, deeply buried rocks as a result of intergranular micro-fissuration during which fluids are released and the solubility of metals is increased (Dandurand et al, 1972). Such a process involving principally diffusion would necessitate very slow movement of Cu, and the presence of H_2S for a long time and so may not be a satisfactory explanation for some SSCU deposits e.g. those of the San Angelo Formation, North Texas (Smith, 1976).

The Sabkha Process: This model attributes the formation of evaporite-associated SSCUs to syngenetic and diagenetic processes of coastal sabkhas (Renfro, 1974). A hot, arid climate with a large evaporation debit is inferred in the coastal region of a shallow, marine lagoon or saline inland sea. The land area of the model has relatively low topographic relief dominated by typical desert landforms e.g. dunes, alluvial fans and sheet-wash plains. It is underlain by unconsolidated terrigenous clastics. With regard to the sabkha process only, the composition of the continental clastics is irrelevant as long as they are porous and permeable. During transgression (Figure 8.8) the hydrogen sulphide-laden algal facies must on-lap the sterile, oxygenated sediments of the adjacent land. When sedimentation and subsidence reach equilibrium, transgression ceases and a sabkha develops immediately landward resting directly upon terrigenous clastics of the desert. Eventually regression occurs and the sabkha progrades basinwards across a wedge of strongly reducing, organic, inter-tidal, lagoonal sediment. The dilute metalliferous solution must pass upward from its oxygenated source beds through the overlying hydrogen sulphide-charged algal mat in order to reach the sabkha surface. This algal mat causes the trace metals in the ascending water to be precipitated as sulphide minerals.

Climatic Enhancements encouraging SSCU development.

1) Climate: The sabkha model outlined above requires the mobilization and transport of trace amounts of Cu, Ag, Pb and Zn from continental red-beds by

low salinity terrestrial formation waters of low pH and high Eh. Rose (1976) has shown that low-salinity waters are relatively ineffective in transporting dissolved Cu, so unless the terrestrial formation waters of Renfro's sabkha model were saline, Cu would not be sufficiently soluble in them to give rise to deposits of appreciable size. Ferguson and Burne (1981) tested the feasibility of aspects of these genetic models through an investigation of interactions between saline red-bed groundwaters and peritidal carbonates of the Spencer Gulf, South Australia. They concluded that terrestrial chloride-rich groundwaters capable of transporting high concentrations of Cu, Pb and Zn are generated within continental red-beds of semi-arid climates. Groundwater sediment interactions within the red-bed aquifers were capable of mobilizing extensive quantities of metals from Fe-oxide grain coatings.

Strakhov (1970) proposed a tectonic-climatic model whereby Cu sulphides were leached from actively dissected highland that lies in a moist climatic zone. Woodward et al (1974) favoured a model similar to that of Strakhov where Cu is carried in solution by water draining into an arid intermontane basin where Cu mineral deposition occurs. The association of Cu ores with coarse-grained clastics appears to be related to changes of water pH in streams moving from humid to arid environments.

Van de Poll (1978) concluded that from the available evidence the palaeoclimate during Carboniferous sedimentation of eastern Canada was of subequatorial monsoon type with seasonal changes in precipitation rates. During Windsor time the prevailing climate was arid. The gradual increase in preservation of plant remains by Late Mississippian-Early Pennsylvanian time indicated a return to more humid climatic conditions with seasonal changes in precipitation rates and warm subtropical temperatures. This suggests that the SSCU deposits of Windsorian age in eastern Canada experienced a clear palaeoclimatic control upon their formation.

2) Biomass:

a) Algal Mats: The importance of algal mats to the sabkha-diagenetic model of SSCU genesis has previously been emphasized. These are associated with semi-arid/arid environments in which sabkhas are formed and they act as reduction zones in which sulphide minerals are deposited. The recognition of abundant carbonized debris of algal mats in the sediments at Mufulira, Zambia led Garlick (1981) to conclude that the bulk of the mineralisation was by syngenetic deposition of sulphides associated with decaying transported algal material in marine lagoons.

b) Plant Remains: In areas of more humid climatic conditions abundant plant debris is more important to the deposition of SSCU deposits in a manner similar to its role in SSUV formation. Caia (1976) described two types of SSCUs in the Lower Cretaceous sandstones of Africa, the first being characterized by fine plant debris with 5-6% Cu related to the abundance of vegetable debris. Where no such debris occurs the host rocks are barren. The second type of ore deposit recognized by Caia contains coarse plant debris e.g. branches and tree trunks. Large scale deposition of sulphide minerals in Triassic sandstones of New Mexico is associated with fossil log-jams within palaeochannels (Woodward et al, 1974) and fossil plant remains have also been cited as the focus for mineralisation in the SSCU deposits of the Corocoro Basin, Bolivia (Ljunggren and Meyer, 1964). In the Carboniferous SSCUs of eastern Canada plant remains are not commonly found. However where they occur Cu sulphide mineralisation is also present. This association led Van de Poll (1978) to conclude that the presence of organic debris appeared to be the controlling influence in the clastic red-bed, Colorado Plateau type of SSCU deposit.

3) Evaporites: The presence of evaporites is virtually essential to the definition of SSCUs as the association is so frequently found. For example the Flowerpot Shale at Creta, Oklahoma is interlayered with, and overlain by,

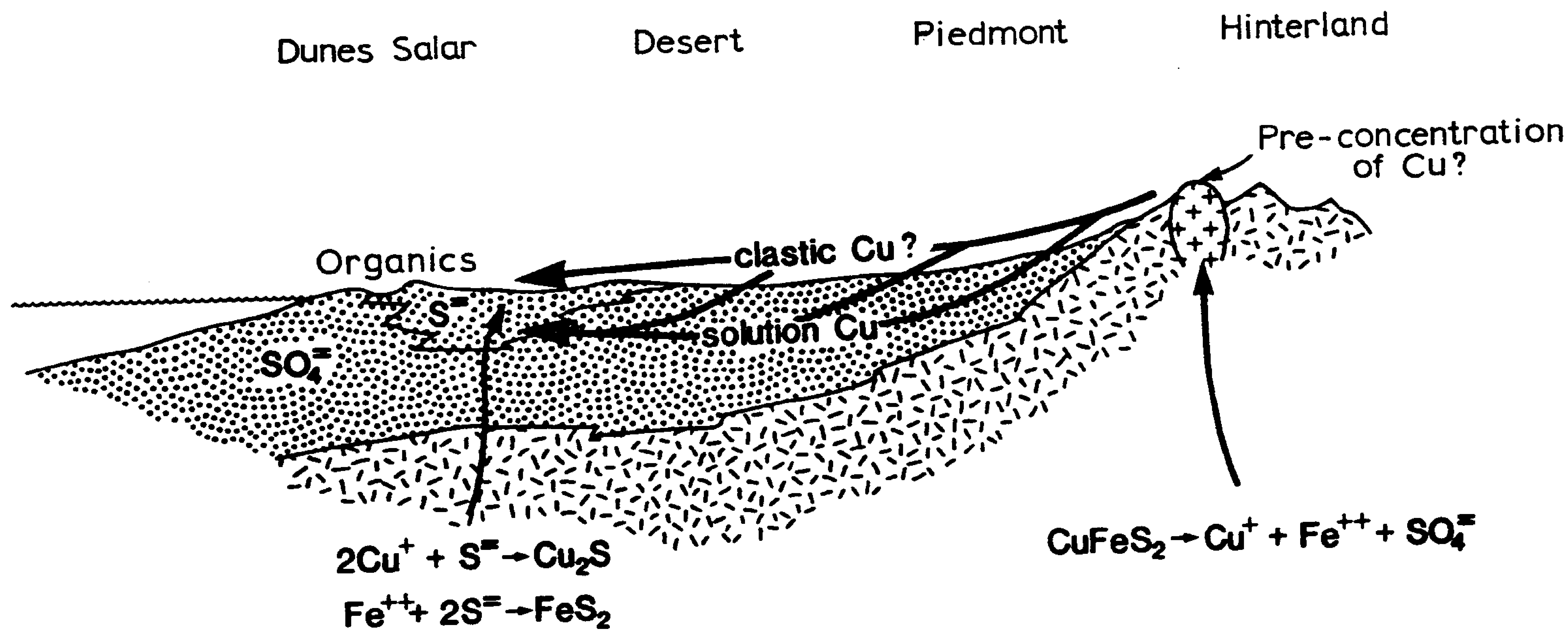
evaporites (Johnson, 1974); in the Permian of New Mexico evaporites, including halite, are common (Rose, 1976); in the Corocoro Basin, Bolivia gypsum and halite occur in the units containing SSCUs (Ljunggren and Meyer, 1964) and in SSCUs of Nova Scotia, New Brunswick and the Redstone area, Canada evaporites are also known (Kirkham, 1974). Anhydrite occurs as a cement in footwall quartzites at Mufulira, Zambian Copper belt (Annels, 1974). Garlick (1972) concluded it had formed interstitially during exposure of muddy sand flats or from detrital gypsum accumulated with wind blown sand dunes. It is envisaged that the original nodular anhydrite formed during early diagenesis under conditions of supersaturation of $CaSO_4$ in a supratidal sabkha environment.

As mentioned in section 8.1.7 the value of evaporites to the deposition of sulphides is mainly two fold; evaporitic groundwater promotes Cu solubility and evaporite sulphate and organics provide sources of sulphur for reduction to sulphide.

4) The source of copper: The hot climatic conditions proposed for the time of SSCU deposition would be responsible for intense chemical weathering of the surrounding hinterland. Woodward et al (1974) proposed the ultimate source of Cu may have been older deposits in Precambrian rocks of the Uncompahgre highland in northern New Mexico and Colorado. Ljunggren and Meyer (1964) suggested Cu-bearing Tertiary basalts of the Altiplano or porphyry copper deposits of the western Andes as the source of Cu for SSCUs in the Corocoro Basin. Samama (1973) concluded that only the heavy metals which had been enriched during weathering on the continent could be found concentrated in the depositional basins. The source of Cu could also be the Fe oxides in red-beds as outlined in section 8.1 8.

The Exotica orebody, Chile lies at 22° 20'S and is found some 2km. below the Chiquicamata porphyry copper orebody (Mortimer et al, 1977). The Exotica gravels are well-stratified thin beds of coarse sands and angular fragments, channels within the gravels are infilled with Fortuna gravels. The cementation in the upper parts of the alluvium by gypsum and iron oxides is light and irregular. However near the bedrock surface the cement is of gypsum and copper oxide minerals. Mortimer et al (1977) considered the genesis of the Exotica mine to be dependent upon the existence of the Chiquicamata mother deposit and, to a lesser degree, upon the prevailing climatic conditions. If the climate was more "pluvial" the water-table would have been much nearer the ground surface, the zone of leaching, oxidation and enrichment in the porphyry deposit would not have been so deep and the volumes of solution generated would probably have been less. Also any Cu that was dissolved would have been in a solution too weak to be an effective reagent. However if the climate were extremely arid, chemical weathering would be much less active and sufficient Cu would not have been released to form the Exotica deposit. The semi-arid climate which prevails at present with infrequent but heavy rains would be ideal for the release of sufficient Cu and the derivation of a sufficiently large volume of solution to form a SSCU deposit. It would also create a suitable permeable piedmont gravel deposit for movement of Cu-bearing solutions and to act as host for the deposition of Cu minerals.

Figure 8.13 shows that Cu sulphide minerals may have been transported in river sediment and deposited as at Chanaral Beach, mouth of the Rio Salado, northern Chile at present. One implication from the occurrence of such "sulphide placers" is that Cu orebodies may have a placer origin (Clemmey, 1978). Rapid erosion of a supergene blanket during a rhexistatic period is thought to be responsible for the release of natural sulphide. The results in Chapter Six, section 6.3.1 indicate that SSCUs occur from 0°-35° from the palaeoequator. Hence they appear to form in both humid and arid climatic



CLIMATIC ENHANCEMENTS

1. Heat promotes dissolution of Cu
2. Evaporitic groundwater promotes Cu solubility
3. Evaporite sulphate & organics provide sulphide
4. Low clastic supply minimises dilution - for SHBM deposits

Figure 8.13. Idealized model of the formation of SSCU deposits showing possible source, modes of transport and precipitation of copper.

zones where a source for Cu is available and sporadic periods of evaporation and precipitation occur.

8.2.2.2 Sandstone Lead Deposits (SSPB).

Sandstone lead mineralisation is usually hosted by basal quartzitic sandstones with some organic matter. Depositional environments range from continental e.g. Yava, Canada to shallow marine e.g. Baltic Shield deposits. It is thought (e.g. Rickards et al, 1979; Bjorlykke and Sangster, 1981) that these form a separate group of mineral deposits from either SSCU or LSBM, although they have certain features in common with both.

Genesis of SSPBs

The genesis of SSPB deposits has been explained by two major models; Groundwater Transport Model: Jerome et al (1965) and Samama (1976) described the model in five basic steps for the origin of the L'Argentière deposit of France;

1) prolonged weathering of the basement,

2) progressive formation of a pediment by active mechanical sedimentation,

3) metal enrichment of saline groundwater percolating through the pediment; weathering becomes more intense so K-feldspars are destabilized and Si, Pb and Ba are released,

4) precipitation of metals at the contact between groundwater and marine water at the edge of the basin. Precipitation of Si is probably due to re-equilibration of the oversaturated water (20 - 80ppm SiO_2) in the pediment. In the most reducing zone (caused by the presence of organic matter) SO_4^{2-} is completely reduced as S^{2-} and HS^- and Pb and Zn are precipitated as sulphides. In less reducing zones both S^{2-} and SO_4^{2-} are present, Pb and Zn are precipitated as sulphides and barium is precipitated as sulphate,

5) lastly there is diagenetic and epigenetic reorganization of the ore minerals.

Basin Brine Model: Such a model was proposed by Rickard et al (1979) for the genesis of Baltic Shield SSPB deposits and comprises three basic steps;

1) dewatering of sediments in the basin. Water would have a high salinity and metal content, the latter carried as chloride complexes,

2) metal bearing brines would then move upwards and outwards through permeable sandstones to the margins of basins,

3) precipitation of metals occurs because of decreases in temperature and pressure and/or mixing with sulphide-bearing groundwater.

This model has many similarities to that for LSBMs by Jackson and Beales (1967), however they invoked compaction as the principal driving force whereas Rickard et al (1979) proposed compression by overriding nappes. In favour of this model is the apparent similarity in the composition of fluid inclusions in SSPB deposits to that from recent basin brines (Rickard et al, 1979). However Bjorlykke and Sangster (1981) considered that such a similarity could be readily explained by a late diagenetic resetting of the inclusions as suggested previously by Bernard (1973). A common feature of SSPB deposits is the presence of an overlying shale (e.g. Christofferson et al, 1979). It is an obvious source of reductant (to precipitate sulphides) and possibly sulphur. This shale may also act as a permeability barrier which would restrict mineralising fluids to the host sandstone channels. Hence the shale overlying SSPB deposits may be an integral part of the SSPB genetic model.

Source of Lead from Continental Weathering?

The basement rocks to SSPBs are usually granites or granitic gneisses. In some cases (e.g. the L'Argentière and Oberpfalz deposits) the basement has a higher average lead content than the average granite as cited by Wedepohl

(1974). It has been proposed (Christofferson et al, 1979) that the SSPBs at Vassbo were deposited during a marine transgression over a deeply weathered basement. This basement has a characteristic morphology of basal SSPBs i.e. it is a nearly perfect peneplain with very few irregularities. Samama (e.g. 1976) suggested that such deep weathering was responsible for determining the chemistry of the L'Argentière deposit of France (i.e. Pb and Zn without Cu, U and V). He distinguished two geochemical areas within the Lower Triassic Formation; one with Cu and U and another in which Cu is virtually unknown. Such differences could not be explained in terms of local palaeogeographical conditions or by the geochemistry of the basement. Instead Samama (1973) evoked a climatic opposition between an area of pre-Triassic weathering conditions in which Cu and partly Zn and Pb were concentrated in the weathering profile and another area with conditions of bisiallitization when Pb was more selectively concentrated. Hence the physiographic conditions prevailing between the end of Permian and the beginning of Triassic sedimentation could have produced a drastic separation of Cu (and U) which are completely leached out and Pb (and Zn) which are enriched in the residual formation. In more detail (Samama, 1969) suggested that the following weathering processes influenced deposition of SSPB at L'Argentière.

1) The basement first experienced constant weathering which was limited to the destabilization of the most unstable mineral species i.e. the plagioclases and ferromagnesian minerals were destroyed. Cu, U and partly Zn were leached out whereas Pb and Ba were concentrated in resistates with K-feldspars. The weathering products were illite, kaolinite, vermiculite and montmorillonite-illite interstratified clays.

2) The weathering process continued and leaching of K-feldspars by percolating waters yielded Si, Ba and Pb. This selectively enriched water carried these elements to the basin margins where deposition occurred.

The Contribution of Organic Matter to SSPBs.

Terrestrial organic debris generally as plant remains is found in most of the Phanerozoic deposits (e.g. Moroccan, German and L'Argentière) although the Cambrian deposits of the Baltic Shield only have very minor carbonaceous material (Bjorlykke and Sangster, 1981). The presence of such organic debris would have maintained chemically reducing conditions in the sandstones. This is supported by the characteristic grey colour of SSPB deposits. As with other deposit types precipitation of metals directly from groundwater would take place under conditions of low Eh and high H_2S content such as those found in a sandstone with abundant terrestrial organic debris e.g. at Yava where the grey or white host sandstones are rich in plant debris and are underlain by carbonates and evaporites. In this instance rising, sulphate-rich groundwater flowing through the organic-laden host sandstones may have produced sufficient bacteriogenic H_2S to precipitate Pb as sulphides from groundwater (Bjorlykke and Sangster, 1981). The spread of sulphur isotope values within individual deposits has lead some authors (e.g. Rickard et al, 1979) to conclude that sulphur was probably not part of the metal-bearing solution. It may have been present as marine sulphate in connate water already trapped in the sediments. Bayer et al (1970) suggested a possible source of sulphur in the Mechernich-Maubach area of Germany would be bacterial reduction of gypsum in the Muschelkalk followed by downward transport of H_2S by groundwater. The presence of even low concentrations of biogenically produced sulphide would preferentially precipitate galena and allow most of the Zn present to remain in solution and pass toward the marine basin (the solubility of galena is approximately 10^4 less than that of sphalerite under reducing conditions).

Palaeoclimate and SSPB formation.

It has been reported (Bjorlykke and Sangster, 1981) that all SSPBs for which palaeomagnetic data are available show a low latitude (0° - 30°) position at the time of formation. The results presented in this project are similar in that the majority of deposits occur in low latitudes, although not all are confined to within 30° of the palaeoequator. It appears that palaeoclimatic conditions for the formation of SSPBs varies somewhat with the majority being deposited in a semi-arid, warm environment. For example evaporites associated with Lower Triassic and Cretaceous deposits suggest such a climate. However abundant organic debris in Yava, Canada, Oberpfalz, West Germany and Kroussou, West Africa may indicate more humid climatic conditions. It has even been suggested that SSPBs in the Baltic Shield were deposited in a relatively cool climate based on evidence from their fossil assemblages (Bjorlykke and Sangster, 1981). Unfortunately no Cambrian palaeogeographic reconstruction was available for this research to support or refute such a suggestion.

Samama (1973) proposed that SSPBs occur in arkosic formations characterized by the lack of iron hydroxides and the scarcity of wood fragments which indicates a drier climate (arid tropical) corresponding to weathering of the bisiallitization process. Such an arid climate would produce illite as a stable phase in highly saline ground water. However Bjorlykke and Sangster (1981) suggested that the evidence indicated humid palaeoclimatic conditions for the formation of Baltic Shield, Oberpfalz and Yava deposits. (As calculations show that weathering rates aremore critical than the metal-transporting capacity of the groundwater, highly saline solutions are not necessarily required for the formation of the ore). One of the most important aspects of Bjorlykke and Sangster's groundwater table model for SSPBs (see Figure 8.15) is the maintenance of a stable and prolonged interface between laterally moving groundwater (with the metals)

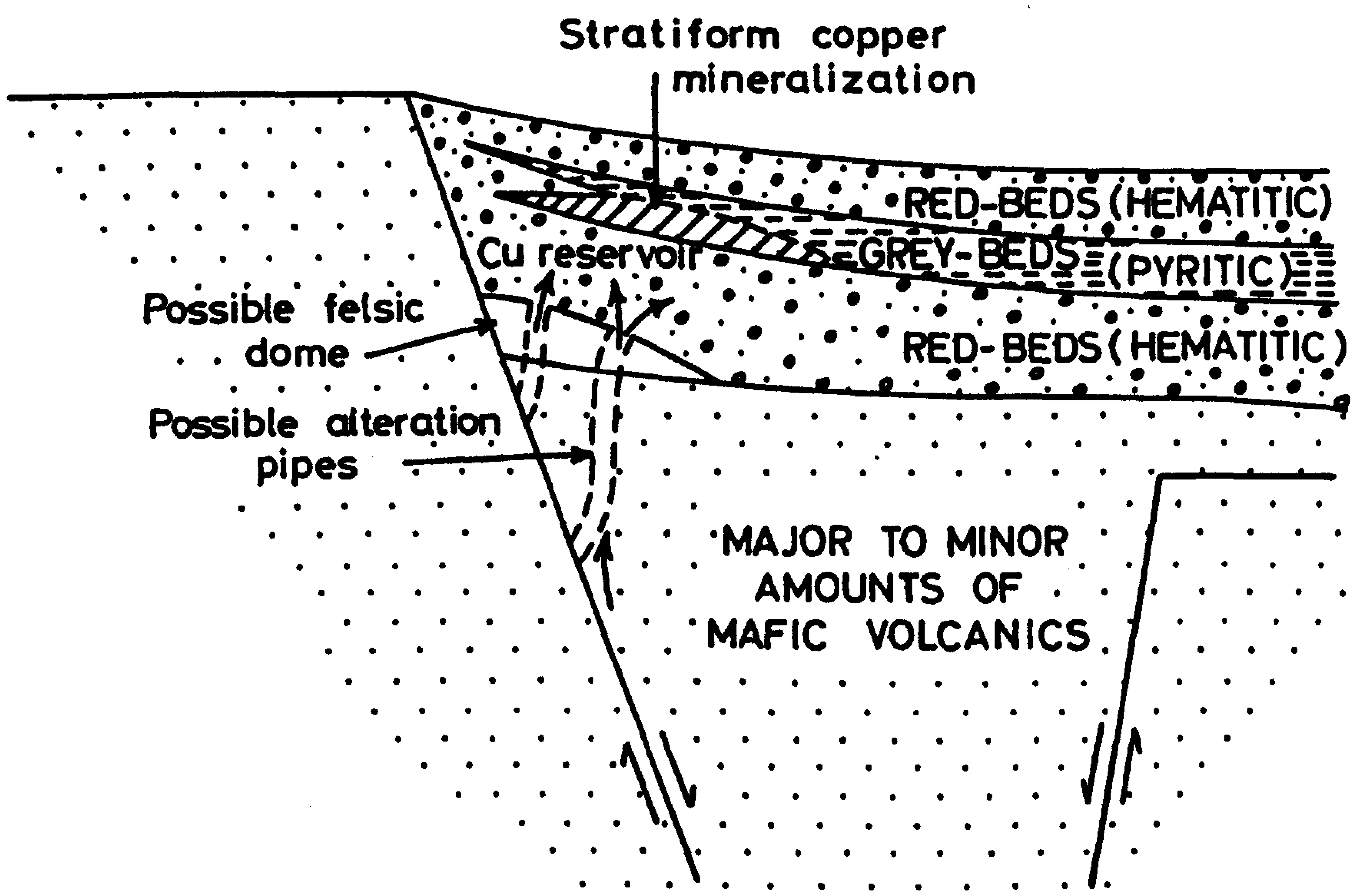


Figure 8.14 Pene-exhalative model for the genesis of SHBM deposits related to intra-cratonic rifting. (After Brown, 1984, Figure 2).

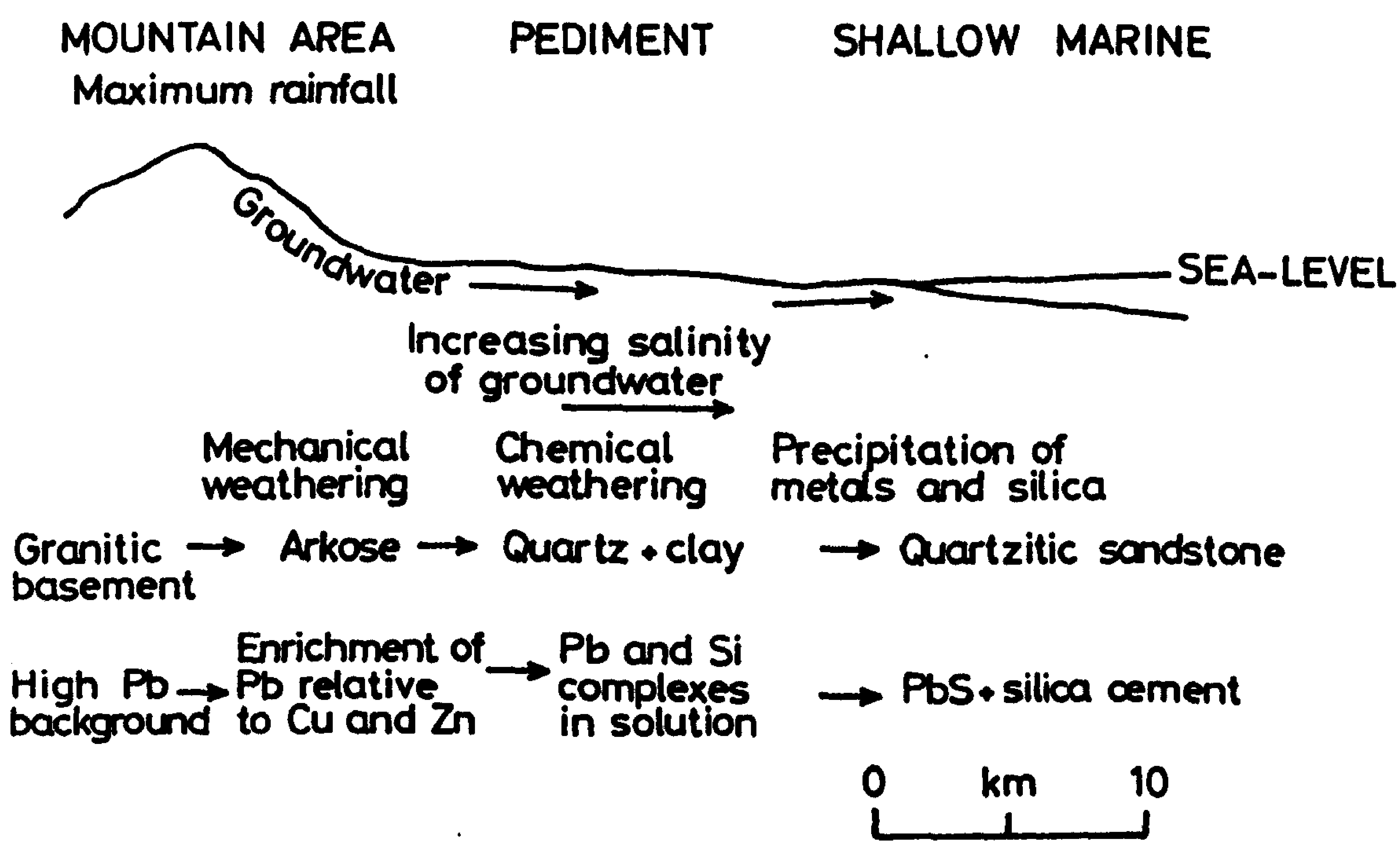


Figure 8.15. Schematic representation of the main features in the groundwater transport genetic model for SSPB deposits. (After Bjorlykke and Sangster, 1981, Figure 26).

and the basin water in which H_2S production occurs. If the groundwater influx was too high, sulphate reduction would be inhibited by the relatively oxygenated fresh groundwater. However if groundwater influx was too low then insufficient metal would be available to support the model. Obviously such fluxes are directly influenced by climatic and physiographic conditions.

Christofferson et al (1979) concluded that SSPB mineralisation in both the Vassbo and Laisvall areas was controlled by the palaeopermeability of the host sandstones. These are thick, coarse, relatively pure sandstones deposited in channels on the palaeosurface. They have a much greater permeability than the under-lying, finer-grained calcite-cemented sandstone or the overlying Alum Shale. Hence any solution entering the region after sedimentation would have preferentially moved along this horizon. The lack of mineralisation in some porous sandstone horizons above basement highs is due to a reduction in solution flow because of the relative thinness of these horizons. This permeability control was highlighted by Rickard et al (1979) as the common denominator between SSUV, SSCU and SSPB deposits.

The results (Chapter Six) suggest that both arid and humid conditions are suitable for SSPB development although the sample size was too small to show that the majority preferred one set of climatic conditions to another. Evaporation in a dry climate would increase the salinity of the groundwater and thereby increase its base metal content - an advantage if a low volume of groundwater was present as in arid regions. An increase in the base-metal content of the groundwater would be achieved in a more humid climate by a high rate of weathering and a higher volume of water.

In all areas the SSPBs appear to be the result of prolonged periods of stable tectonic conditions which caused deep chemical weathering and produced regional peneplanation. Ultimately Pb would be released from the breakdown of K-feldspars and transported to the site of deposition in tidal sands by highly saline groundwaters, or a large volume of less saline waters in humid conditions. Deposition of metals as sulphides occurred in sufficiently permeable sandstones of low iron content in a reducing environment probably by abundant organic matter. Whether the sulphur and the metals were transported in the same, or different, solutions is still a matter of debate.

The relationship between SSPB, SSCU and LSBM deposits has been frequently referred to e.g. Garlick (1969) thought some SSPB and SSCU deposits were closely related; Heyl et al (1974) suggested that the Laisvall SSPB deposit was a variety of LSBMs. Bjorlykke and Sangster (1981) considered that these three groups were separate deposit types although it was proposed that they illustrated basin development. SSCUs were hosted by rapidly deposited, immature rift-generated feldspathic red-bed sandstones with early release of Cu from the more readily altered mafic minerals during mild chemical weathering. Peneplanation from prolonged weathering in stable tectonic conditions led to a slow marine transgression onto the continent, the development of relatively pure quartzitic sandstones and the deposition of lead sulphides in SSPB deposits. Bjorlykke and Sangster (1981) considered the Zn-rich nature of seawater (Goldberg, 1975) to be reflected in the Zn-dominated nature of LSBMs. The Zn concentration would be enhanced by groundwater passing through the pediment environment. If the more continental sandstone environment filtered Pb and precipitated it in preference to Zn as mentioned earlier, then groundwater reaching the marine environment would tend to have a high Zn/Zn+Pb ratio.

8.2.2.3 Shale-hosted Base-Metal Deposits (SHBM).

There appears to be a close association between SHBM and SSCU deposits in a number of deposits e.g. the Zambian Copper belt and the Kupferschiefer of Europe, which has led to the theory that there may be a genetic relationship between the two. The major characteristics of SHBMs have been given in Chapter Three, section 3.2.7 together with a detailed description of the association mentioned above. However one of the features of SHBMs that should be emphasized is the regular occurrence of evaporites and red-beds with these deposits. Naturally such a close stratigraphic relationship is of interest to this study because of the inferences regarding climatic conditions at the time of ore deposition that may be made.

Genesis of SHBM deposits

Exhalative Model: Brown (1981) proposed that stratiform copper deposits (SSCUs and SHBMs) originated from exhalative-related processes similar to those which form SDEX deposits (see Figure 8.14). In favour of this hypothesis he cited the repeated association of predominantly mafic volcanics and/or sills with some deposits e.g. White Pine. For this particular deposit, rapid clastic infilling of a rift may have been accompanied by circulation of exhalative metalliferous brines into the host rocks during a minor waning phase of felsic volcanism which may also explain the high Co content. In the Redstone area, NWT, Canada disseminated copper mineralisation in tidal-flat units may be due to the introduction of thermal cupriferous brines associated with rift-margin faults (Brown, 1981). Later Brown (1986b) noted that some features of stratiform copper deposits (such as overgrowths and replacement textures) indicated that metal deposition was postsynsedimentary so excluded a direct exhalative model for metal emplacement. He proposed a model of pene-exhalative hydrothermal activity for SHBM deposits in which the metalliferous fluids encountered porous oxidized strata (e.g. red-bed clastic units) before reaching the sediment-water interface. These ascending ore-fluids would form a reservoir typically confined beneath very fine-grained highly impermeable beds rich in organic matter as described above for SSPB deposits. Such grey beds contain sulphides (or sulphates which may be reduced biogenically during very early diagenesis) and form a chemical sink for copper. Brown also proposed that copper was introduced from the adjacent red-beds by infiltration or diffusion. Dunham (1964) suggested that heavy metal enrichment in the Kupferschiefer was due to submarine exhalations similar to those of the present Red Sea. The concentration of metals from submarine hydrothermal springs was also proposed as a model for SHBM genesis by Degens and Ross (1969).

Other Syngenetic Models: These include the concentration of metals;

1) from open sea waters (Brongersma-Sanders, 1968),

2) from seawater where a lagoon and aerobic sea are separated by a barrier (Haranczyk, 1972),

3) by cation exchange between oxic and anoxic waters with a continental source of metals (Wedepohl et al, 1978). Wedepohl (1971) favoured a source of metals from surrounding red-beds for the Kupferschiefer.

Epigenetic Models: These include the concentration of metals;

1) by means of the sabkha process (Renfro, 1974) as described for SSCU deposits in section 8.2.2.1,

2) from brines descending from the associated evaporites which leach base metals from magmatic sulphides at depth and redeposit them in host rocks (Davidson, 1965 for the Kupferschiefer deposit). A major objection to this model is the very large lateral extent of this particular deposit (Vaughan, 1976),

3) from solutions ascending from the "molasse intraorogenic deeps" (Rentzsch, 1974),

4) during diagenesis of the tidal-flat sediments with the sulphide supplied from modified connate waters and metals supplied from hypersaline brines from below. Such tidal-flat deposits are thought to be excellent traps for sulphophile metals because of their high organic matter and sulphate contents, low iron content, high initial porosity and their association with evaporites and chloride-rich brines (Bartholome et al, 1973),

5) by the mixing of two brines (Kucha and Pawlikowski, 1986 for the Kupferschiefer-type deposits of Poland). The upper cold brine (pH >7) originated from overlying evaporites and was rich in Na, Ca, Cl and SO_4. The lower hot brine (pH <7) formed in sediments in the central part of the Zechstein Basin and was rich in Mg, K, Cl, SO_4 and CO_3. This brine became enriched in heavy metals by leaching of rocks underlying the Zechstein host rocks.

The palaeolatitudes of SHBM deposits examined here reflect that a certain degree of climatic control may exist on the formation of these deposits, although the results are not conclusive owing to the unreliability of some of the palaeogeographic reconstructions (see section 6.3.3). However a low latitude control is inferred and supported by the lithologies associated with SHBM deposits.

The Red-bed Association.

Gustafson and Williams (1981) mentioned that sulphides in the Kupferschiefer occur within a few metres of the underlying Rotliegend red-beds over areas of thousands of square kilometres. Galena and sphalerite are common in the underlying reduced sandstones which contain varying amounts of copper mineralisation (see Figure 8.16). Indeed the majority of Kupferschiefer ore is in this bleached sandstone (known as the Weissliegende) beneath the Kupferschiefer shale. Both the Zambian Copper belt and the Kupferschiefer were classed by Gustafson and Williams (1981) as being first marine transgressions over red terrestrial successions. The deposits occur at the base of the overlying reduced marine sedimentary rocks and the red clastics lie on an eroded basement. They are thought to have been deposited in an arid environment as evidenced by the presence of anhydrite and gypsum. However above the Weissliegende the reduced sandstone, ore grades in the shale are at a maximum which is evidence for the diagenetic reduction of the

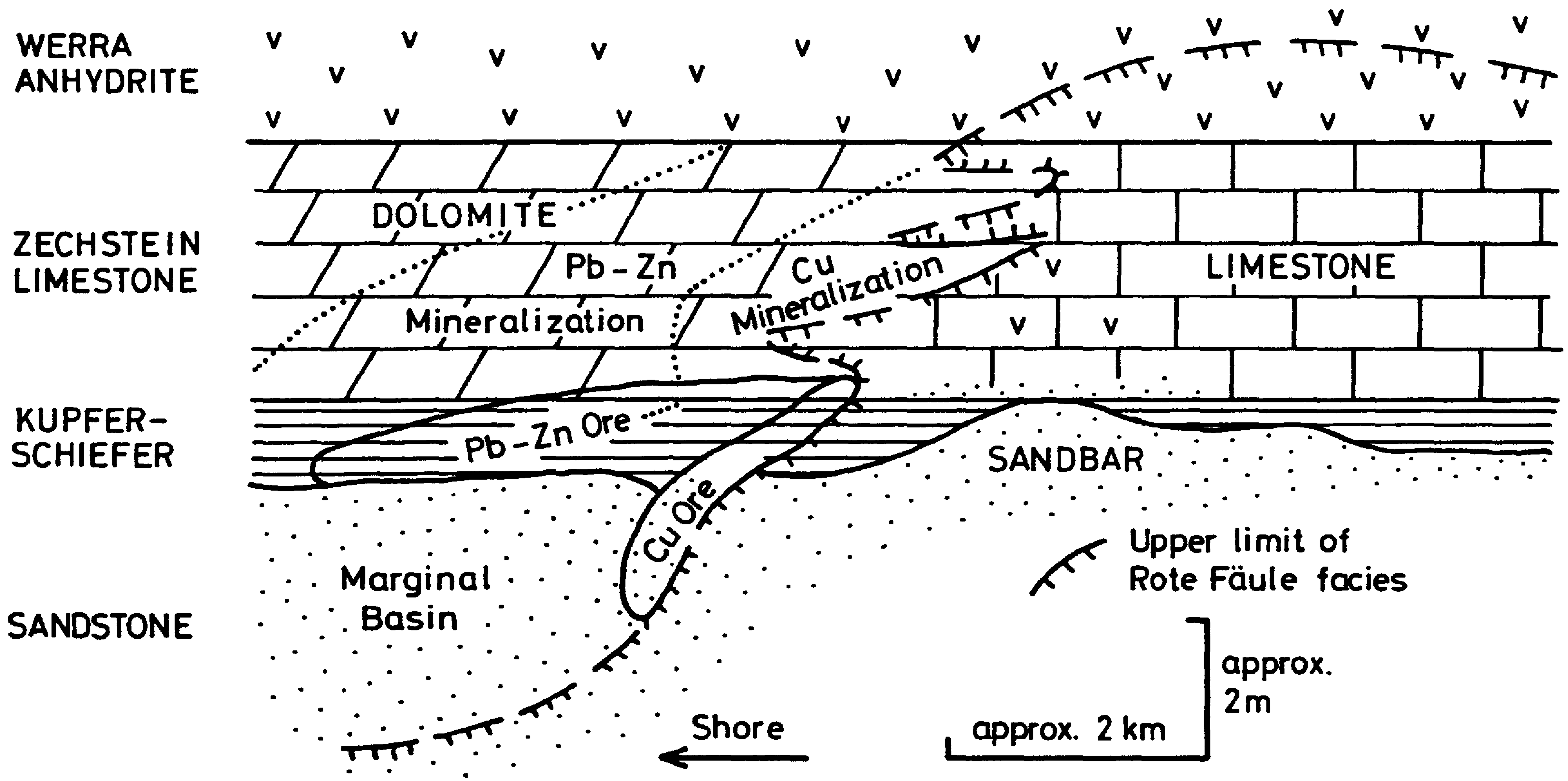


Figure 8.16. Typical stratigraphic sequence of the basal Zechstein with Rote Faule facies alteration. Sulphide mineralisation is concentrated in the adjacent overlying unoxidized zone with copper next to the Rote Faule. (After Rentzsch, 1974; Brown, 1978, Figure 8).

red sandstone by the mineralising fluid. Annels (1984) listed features of the Zambian Copperbelt in common with intracratonic rift environments (as reviewed from Rowlands, 1974; Rowlands et al, 1978, 1980; Lambert et al, 1980) the following infer particular climatic conditions;

1) evidence of saline to hypersaline waters and thus an arid climate,

2) abundance of evaporitic mineral phases and nodular concretions of anhydrite,

3) the presence of stromatolites,

4) continental non-marine to prograding marine sabkhas.

The Evaporite Association.

The Flowerpot Shale in Permian Cu-bearing shales of southwest Oklahoma is characteristically reddish-brown with thin interbeds of gypsum (80-90% of evaporite sediments), dolomite (10%), siltstone, sandstone and greenish-grey shale (Johnson, 1976). At Rokana in the Zambian Copper belt Clemmey (1978b) interpreted the orebody zoning sequence as resulting from several cycles of shallow submergence and emergence in a very quiet epeiric marine or lacustrine environment with a climate conducive to evaporitic conditions. The association of the Kupferschiefer with evaporites has been mentioned previously with regard to models of genesis. Wedepohl et al (1978) considered that evidence of a dry climate during the Lower Permian was given by the occurrence of typical red-bed and evaporite lithologies within the sequence.

Low Detrital Supply.

It appears likely that a low clastic supply (due to high relief or low rainfall) to the depositional site is important to minimize dilution. Annels (1984) proposed low relief hinterlands to the depositional basins of the Zambian Copperbelt, the latter being slowly subsiding with a balance between subsidence and the supply of detritus. This aspect of SHBMs was also mentioned by Wedepohl et al (1978) with reference to the Kupferschiefer. They considered that the moderate and equal thickness of the Kupferschiefer bed from England to Poland indicated that the topography and climate did not allow much erosion of the areas adjacent to the Kupferschiefer depositional basin, resulting in low detrital accumulation. Poole and Desborough (1981) recommended a relatively slow rate (<10m/m.y.) of detrital deposition and/or chemical precipitation to enable the accumulation of low temperature black shales in Ordovician and Devonian rocks of Nevada, USA. These three examples suggest that a more arid than humid climate would be favourable for SHBM development.

The Organic Association.

In a review of the literature on deposits which have been classed here as SHBM the association with organics is repeatedly mentioned e.g. Preidl and Metzler (1984) described Cu-bearing shales of the Sudetic Foreland deposited in lagoonal areas in which an oxygen deficiency was associated with a great amount of organic matter. Indeed the Kupferschiefer itself may be described as an oil shale of marine algal origin. Poole and Desborough (1981) proposed a high organic carbon content (>5 wt.%) for Ordovician and Devonian black shales of Nevada consisting of high sapropel content (mostly from marine algae) and low detrital land plant material (unlike SSCU deposit). Vaughan (1976) suggested that euxinic bottom conditions existed at the time of deposition of the Kupferschiefer together with abundant organic matter suitable for enabling bacterial sulphate reduction. This role of bacteria for the reduction of sulphate ions in sea water was also proposed as a mechanism for producing sulphides by Johnson (1976) for Permian Cu-bearing shales of southwestern USA and by Saxby (1976) for SHBM deposits in general. Brongersma-Sanders (1969) suggested that the metals were derived initially from ocean water and were concentrated by living plankton as metal complexes, as outlined in section 8.1.9. Coveney (1979) also suggested organic activity as a source for Zn in Carboniferous black shales of Missouri and Kansas. The presence of organic matter has been used by Kucha and Pawlikowski (1986) to explain metal zonation in Polish Kupferschiefer-type SHBM deposits. They suggested that low pH due to oxidation of organic matter (producing carbonic acid) in the host shale and the top of the white sandstone promoted Cu concentration. The horizon above this shale, enriched in H_2S, favoured PbS deposition. The uppermost zone of the deposit was within the base of the upper cold brine originating from overlying evaporites, so high pH conditions existed and ZnS minerals were precipitated.

It is obvious from the above that organic matter plays an important role in the genesis of SHBM deposits. However this organic matter may be algal in nature unlike the large, plant debris associated with certain SSCU deposits. Quiet depositional conditions with low clastic supply and abundant organic activity are required for SHBM deposits and especially conditions where evaporites may develop. Hence a more arid climate is inferred with evaporites providing organics and sulphate, a low clastic supply minimizing dilution and intense weathering of hinterland perhaps supplying the metals. But the SHBM palaeolatitudes show a tendency for SHBMs to occur between 0° and 10° north and south of the palaeoequator. According to present conditions at these latitudes this suggests that a more humid than arid climate prevailed during SHBM formation. However the majority of the SHBM examples used in this research are Permo-Triassic (250 m.y.) in age. The presence of the supercontinent Pangaea across the equatorial region during this time may have greatly influenced the climatic conditions experienced by particular palaeolatitude zones.

The phenomenon of continentality (see section 4.5.2.2) must have influenced the climatic conditions experienced over the supercontinent of Pangaea. The main characteristics of low humidity, extremes of temperature and low clastic supply would have occurred particularly towards the interior of the continent. These conditions are very similar to those proposed above as being the most favourable for SHBM formation. Another consequence of the vast, continuous land area is that a monsoonal circulation was probably dominant (Robinson, 1973). The term "monsoon" means alternating circulation between high pressure in the winter and low pressure in the summer e.g. as experienced in the interior of Asia today. During the Permo-Triassic the contrast between high and low pressure belts which occurs during the northern summer at present may have occurred in both summer and winter due to the huge latitudinal extent of Pangaea. The low pressure cell in the northern summer over northern Pangaea (caused by high temperatures) would have contrasted with a continental high in the south and vice versa during the southern summer. Such an extreme form of monsoonal circulation would have kept the equatorial region very dry and maximized the seasonality of the coastal regions of the Tethys (Parrish et al, 1982).

8.2.2.4 Sedimentary-Exhalative Deposits (SDEX).

Amongst the most well-known and extensively researched SDEXs are a number of Proterozoic age e.g. Mt. Isa, Hilton, McArthur River and Sullivan. So any discussion on the genesis of SDEXs must include mechanisms proposed for these deposits, regardless of their age. However it should be remembered that only Phanerozoic examples have been included in the results given in Chapter Six, section 6.3.4. The SDEX sedimentary host rocks are very variable and range from dolomites through dolomitic siltstones e.g. Mt. Isa to silty argillites and shales e.g. Rammelsberg (Large, 1981b). It is not clear whether active volcanicsm played a major role in their genesis, but evidence of contemporaneous volcanic activity is commonly found in the immediate host rocks or in stratigraphically younger rocks e.g. glass shards in the McArthur River deposit (Lambert, 1976); tuffites within the host sequence at Tynagh

(Morrissey et al, 1971). The questions concerning SDEX genesis which are still to be resolved echo those given for other deposit types. Namely, the source of the metals and the sulphur and nature of the metal-bearing ore solutions.

Source of Sulphur.

The sources of sulphur for SDEXs are generally discussed with reference to sulphur isotope studies. One of the most popular hypotheses proposes a dual sulphur source and was first used to explain the different isotopic behaviour of pyrite to galena/spalerite in the Rammelsberg deposit (Anger et al, 1966). It has subsequently been used by Smith and Croxford (1973) for McArthur River; and by Taylor and Andrew (1978) and Coomer and Robinson (1976) for Silvermines. It has been suggested (Large, 1981b) that the sulphur in SDEX galena, sphalerite and pyrrhotite is deep-seated being introduced to the site of mineralisation by the same hydrothermal solution that transported the metals. However as with other sediment-hosted deposit types already described (e.g. LSBMs) the problems of transporting metals (as chloride complexes) with sulphur (for reduction to sulphide) in the same solution without reacting needs to be resolved, so casting doubt on this hypothesis. The ultimate source of sulphur is probably sea water sulphate which Hajash (1975) thought to be inorganically reduced at temperatures greater than 200°C during convective circulation through the underlying prism. Sea water was also given as the source for sulphate in the Vulcan deposit (Mako and Shanks, 1984) which was reduced by thermal decomposition of organic matter during maturation of hydrocarbons. The pyrite-sulphur may be the product of biogenic reduction of sea water sulphate for McArthur River (Smith and Croxford, 1973) and Tynagh (Boast et al, 1981).

Source of Metals.

There are a number of possible sources for SDEX base-metals.

1) Russell (1983) suggested that the thick sediment pile at Mt. Isa was the source of metals which were leached by descending sea water acidified by early reaction with clays and feldspars.

2) Leaching from sediments and volcanics. Bischoff and Seyfried (1978) showed that sea water trapped in basalt tends to become increasingly acid at higher temperatures and is therefore able to leach metals from the basalt. Williams (1978) suggested that metal-rich solutions were expelled from the sedimentary pile. These solutions ascended up major faults as the result of compaction e.g. at McArthur River (Rye and Williams, 1981).

3) Subaqueous metal-rich fluids released during volcanic activity (Murray, 1975). Large (1976) proposed that some of the Pb and Zn may be derived from an isotopically homogenous magmatic source e.g. at Sullivan, the presence of cassiterite may be indicative of partial melting of continental crust (Garson and Mitchell, 1977).

A Volcanic Origin for SDEXs?

Coats et al (1980) proposed a hydrothermal origin for the metal-bearing brine of the Aberfeldy deposit which may have been associated with igneous activity;

a) thermal energy associated with a rising basic magma could have stimulated leaching of the underlying crust by meteoric or connate waters,

b) the brine could be of juvenile origin derived from unrecognized or concealed igneous source rocks.

The heating of groundwater by igneous activity has been proposed by other workers e.g. Russell (1968) and Lambert and Scott (1973) whereas Kraume et al (1955) implied direct introduction of magmatic solutions in SDEX formation. However one of the major problems with the proposals is that directly beneath most SDEXs there is generally a lack of plutons of either sufficient size to heat the groundwater or of appropriate composition to exsolve Pb and Zn solutions (Russell et al, 1981). With reference to the Mt. Isa deposit Russell (1983) mentioned that the mineralisation occurred for a considerable period of time. However if the Sybella granite acted as a heat source and was the driving mechanism of convection then it would be expected that mineralisation would lessen with time. The widespread and long-lasting nature of the mineralisation and the lack of significant alteration pipes beneath the ores or of any evidence of local hydrothermal circulation all point to a non-volcanic origin. However the homogeneous, non-radiogenic Pb isotope values for some SDEXs does suggest a magmatic source for at least some of the ores (Badham, 1981a).

Models of SDEX genesis.

The two most popular theories of SDEX genesis at present include the basinal compaction (Gustafson and Williams, 1981) and the hydrothermal convection models (Russell, 1983).

Basinal Compaction Model: Pb and Zn are extracted from silicates in the surrounding rocks by highly saline formation water. These metal-bearing brines are then expelled towards the surface along fault zones during compaction (Gale, 1983). This model fits most neatly for deposits which have evaporites within their immediate stratigraphy (e.g. McArthur River) so that highly saline fluids were almost certainly present within the sediment during the early stages of diagenesis. The evidence that metals in the HYC deposit emanated from fault zones (Williams, 1978) lends further credence to the model. However Edwards and Atkinson (1986) highlighted one of the most important problems with the basin compaction model: if it is assumed that a geothermal gradient of 40°C/km existed the bulk of the ore fluid would be expelled at temperatures between 70 to 100°C. But it has been suggested that the temperature of the McArthur River ore fluid was in the range 120° to 240°C (Rye and Williams, 1981).

Hydrothermal Convection Model: Russell (1978 and 1983) is perhaps the most vigorous exponent of the downward-penetrating convection cell model of SDEX ore formation. He suggested that SDEX deposits are characterized by foundering of the sea floor due to extensional strain in the upper part of the crust. This led to enhanced permeability which allowed convective circulation at relatively low temperatures. Continued cooling and fracturing of the rocks resulted in convection extending to increasing depths (Russell et al, 1981). However there are problems with this model, too. There is little evidence of foundering in such deposits as Mt. Isa, indeed the Urquhart Shale suggests uniform conditions of sedimentation (Edwards and Atkinson, 1986). Also Pine and Batchelor (1984) have shown that downward growth of microfaulting associated with hydraulic injection is dependent upon the presence of a strongly jointed rock. Edwards and Atkinson (1986) cast doubt as to whether the partly lithified rocks of SDEXs would have had sufficient rigidity for joint formation.

Climatic Enhancements for SDEX formation.

The SDEX deposits examined in this research show a concentration in low latitudes, and although there is an preference for warm arid regions, they appear to have formed in humid, equatorial regions too. These warm environments encourage extensive chemical weathering of the hinterland to give a high metal input into the sea basin.

In some SDEX deposits there are evaporites in the stratigraphy or evidence of the presence of the previous existence of evaporites. For example Croxford and Jephcott (1972) proposed an arid palaeoclimate for the McArthur River deposit. They considered the scapolite-rich rocks of the Cloncurry-Mt. Isa district (Edwards and Baker, 1954) to be metamorphosed evaporites.

Further evidence of hypersalinity (due to aridity) is indicated by locally abundant barite at McArthur River and Mt. Isa which may have been transported by saline waters (Davidson, 1966; Dunham, 1966; White, 1968). Also at Mt. Isa halite infillings of shear zones are found in dolomite rich rocks. Mathias and Clark (1975) suggested that Cu deposits of Mt. Isa represented dolomite-chert facies indicative of an oxygenated, shallow water environment. They also proposed that some of the silica-dolomite may be original algal material found in semi-arid environments. Where evaporites do occur they are important as a likely source for the vast amount of sulphur that is eventually fixed as sulphide and for the high salinity required to move metals at relatively low temperatures. Strongly seasonal climates, with extended dry periods, but not necessarily complete year-round aridity can produce hypersaline lake waters that would be an adequate supply for SDEX deposition (Gustafson and Williams, 1981). Also evaporites in the vicinity of SDEXs have been cited as a source for chloride-rich brines derived from their dissolution (Walker et al, 1978). However the presence of evaporites has not been established in every case (e.g. Meggen, Rammelsberg and Sullivan) so it appears that this may not be a vital aspect of SDEX formation, just another enhancement for some deposits. Russell (1983) suggested an indirect role of evaporitic brines via density flow and trap, in addition to their direct role in shallow water deposits.

In other deposits there is some evidence of shallow water conditions, such as stromatolitic carbonates and reefs with their climatic implications as described for LSBMs e.g. Meggen, Lady Loretta. However the deposits themselves are confined to small basins characterized by carbonaceous, finely laminated silts and carbonates lacking evidence of biogenic activity in the photic zone (Finlow-Bates and Large, 1978).

The "third-order" basins in which SDEXs are characteristically found (Large, 1981b) are themselves situated within larger, fault-bounded basins (see Figure 8.17). These faults appear to have acted as conduits for early

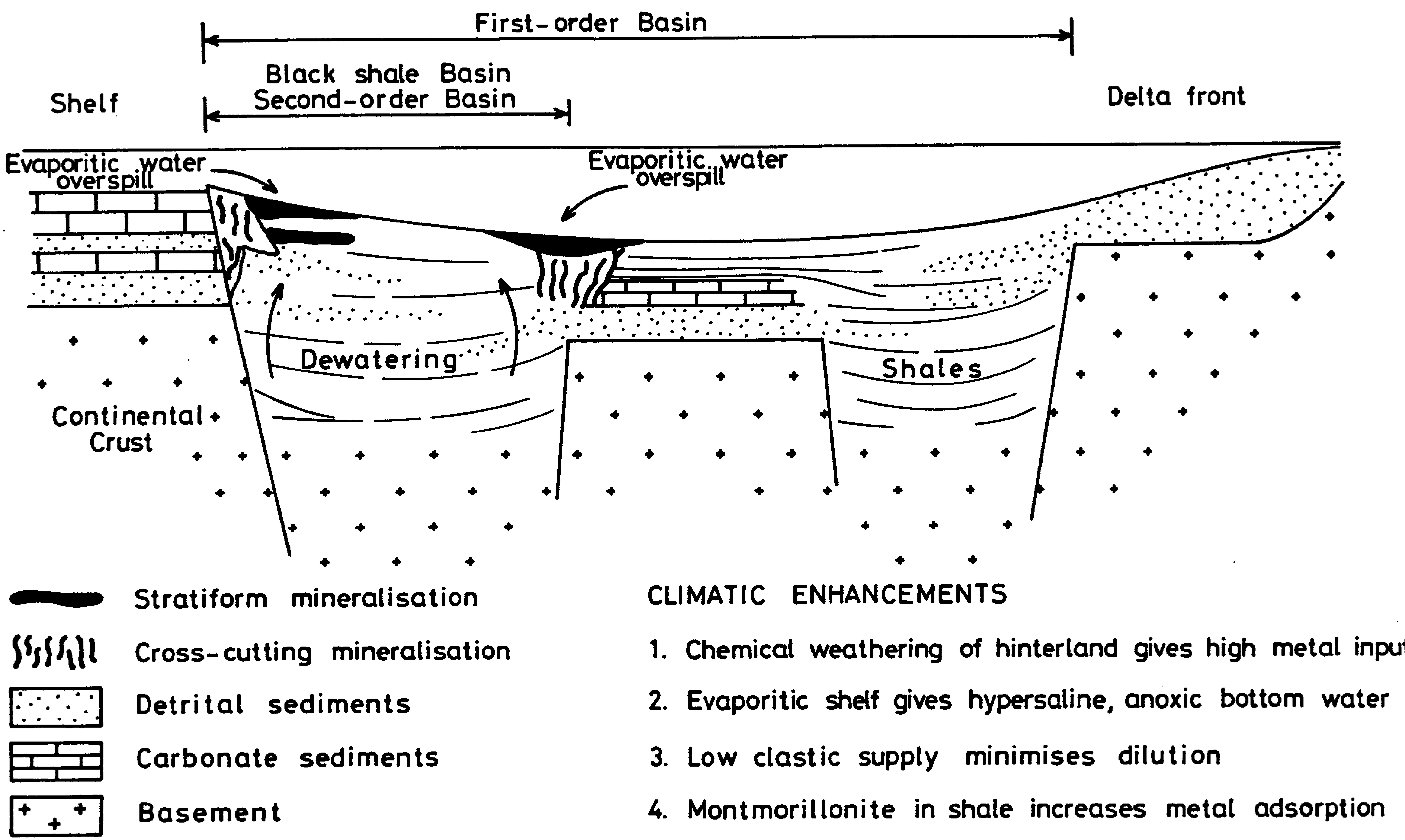


Figure 8.17. Idealized model of SDEX formation illustrating the two main possible sources of metals (evaporites/basin dewatering). (Modified after Large, 1981, Figure 3).

mineralisation in many deposits e.g. Irish examples (Badham, 1981a). The host rocks in these basins are usually fine-grained clastics i.e. shales and siltstones which show no evidence of having been deposited in high energy sedimentary environments such as ripple marks, cross-bedding or erosive channel margins. Instead they are characteristic of sediments from quiet, euxinic environments (Large, 1980). Precipitation of FeS_2, ZnS and PbS is thought to have occurred in dense cooling brines beneath a sulphate-sulphide reduction zone with carbonates being formed from the brine at very high salinities or precipitated from warm overlying marine waters (Coats et al, 1980). The mineralogy of these shales is also very important to metal deposition, as a high clay mineral (e.g. montmorillonite) content in shales considerably increases their metal adsorption properties. In section (8.1.1) it was shown that the prevailing climatic conditions at the site of erosion could affect the mineralogy of the weathering products. The presence of shales in these basins also suggests a low clastic supply to the basin (again as a function of climate) and they minimize dilution of metal concentration within the basin. Badham (1981a) mentioned that early erosion on the continental side of the basin would probably result in rapid peneplanation leading to a cessation of the clastic supply in the absence of any organic processes. Also as basin evolution progresses central areas become distal or separated from sources of clastic sediments.

In the vicinity of Tom, McArthur River and Meggen autochthonous shales in the depositional basins have relatively high concentrations of organic carbon (up to 5%) as compared with the rest of the sedimentary sequence, perhaps reflecting a high productivity zone (Large, 1980). This is in keeping with the theory of biogenic reduction of seawater sulphate in the reducing environment of the depositional basin as a source for pyrite-sulphur. The bacterial reduction of sulphate to sulphide is thought to be affected by such fatcors as sulphate concentration, temperature, pH, fO_2, concentration of bacteria and availability of nutrients (Saxby, 1976) many of which are influenced indirectly by the climate.

One aspect of SDEX mineralogy is particularly important because it is of immense economic importance and so is of genetic concern - whether or not the deposits contain significant quantities of Ag e.g. Rammelsberg with 120g/T (30-35°), Mt. Isa with 149g/T, Meggen with 3g/T (20-25°/35-40°), Tynagh with 28g/T (0-5°/15-20°). Unfortunately no rigid pattern can be discerned in the SDEX palaeolatitudes which can distinguish between the Ag-poor and Ag-rich types. There is a slight suggestion that Ag-rich deposits occur in more arid environments than Ag-poor ones, but this is not an obvious association.

It has been suggested that there is a genetic link between SDEX and LSBM deposits e.g. Mako and Shanks (1984) with reference to the Vulcan Prospect deposit. They suggested that both deposit types could be the result of basinal fluids migrating up active faults at the carbonate platform to shale basin transition. The primary distinction is that in LSBMs the mineralising fluid is introduced into a lithified carbonate host whereas in the shale basin the fluid is exhaled onto the sea floor during deposition of the host sediments.

8.2.2.5 Sandstone-hosted Uranium-Vanadium Deposits (SSUV).

As mentioned earlier in the classification (Chapter Three, section 3.2.1) there are two varieties of U accumulation of particular interest to this research; the roll-front and tabular types. The host rocks are mainly fluviatile sandstones. The dominant sedimentary forms are alluvial fans within which braided streams are the main channels for sediment transport and deposition. Most workers consider that U has been introduced into the host fluviatile sandstones shortly after deposition by migrating, well-oxygenated meteoric water. Some of the factors of importance to the development of SSUV include source rock, rates of uplift in the source area, sediment supply, rainfall, stream gradient, vegetation growth and rates of sediment accumulation.

The Geochemistry of Uranium.

The transport of U is in the 6^+ or, less commomly, the 5^+ states and its solubility can be greatly enhanced by the formation of complexes with other ions in solution. U^{6+} (as UO_2^{2+}) complexes with the following in order of decreasing strength of association; CO_3^{2-}, HPO_4^{2-}, OH^-, F^-, H_2PO^{4-} and SO_4^{2-}. The strongest complexes for the uranous (4^+) species are OH^-, HPO_4^{2-}, F^- and SO_4^{2-}. At the concentrations found in most natural waters (Table 3.2) carbonate and phosphate complexes are the most effective transporting agents with fluoride becoming more important at low pH (Maynard, 1983). Lukacs and Florjancic (1974) proposed that U was transported as complex uranyl carbonate ions, as the uranyl-hydroxide ions and as colloidal aggregates of uranyl solutions in oxidizing solutions with a pH value of 7 and 8.

The geochemical behaviour of U is largely controlled by three factors:

1) Oxidation-reduction reactions are a crucial factor in the development of most U ores i.e. low Eh leads to the precipitation of uraninite or coffinite.

2) The amount of CO_2 in the system. There is a sharp increase in U solubility resulting from carbonate complexing (Figure 8.18). Uranium is also enriched by dissolution of dispersed metals by oxidizing solutions.

3) If appreciable amounts of V or phosphate are present uranyl deposits can form under oxidizing conditions. Carnotite (the least soluble uranyl mineral) may be deposited under neutral to slightly acid conditions implying it would be impossible to transport V and U in the same solutions (see Figure 8.19). A more detailed outline of U geochemistry can be found in Langmuir (1978) and Nash et al (1981).

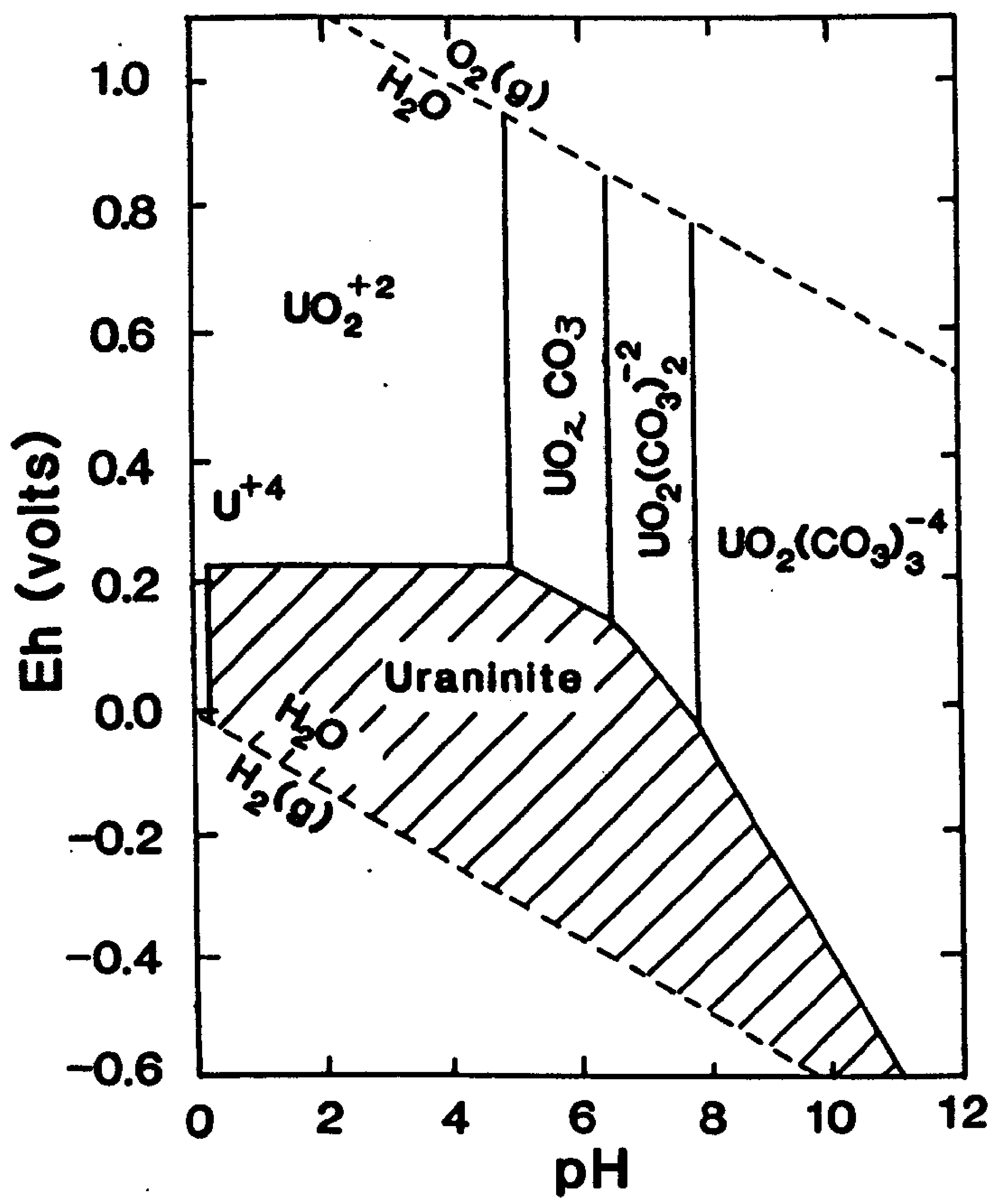


Figure 8.18. Eh-pH relations for uraninite and aqueous solution with CO_2 demonstrating the necessity for reducing conditions for precipitation of uraninite at most pHs. (After Langmuir, 1978, Figure 14).

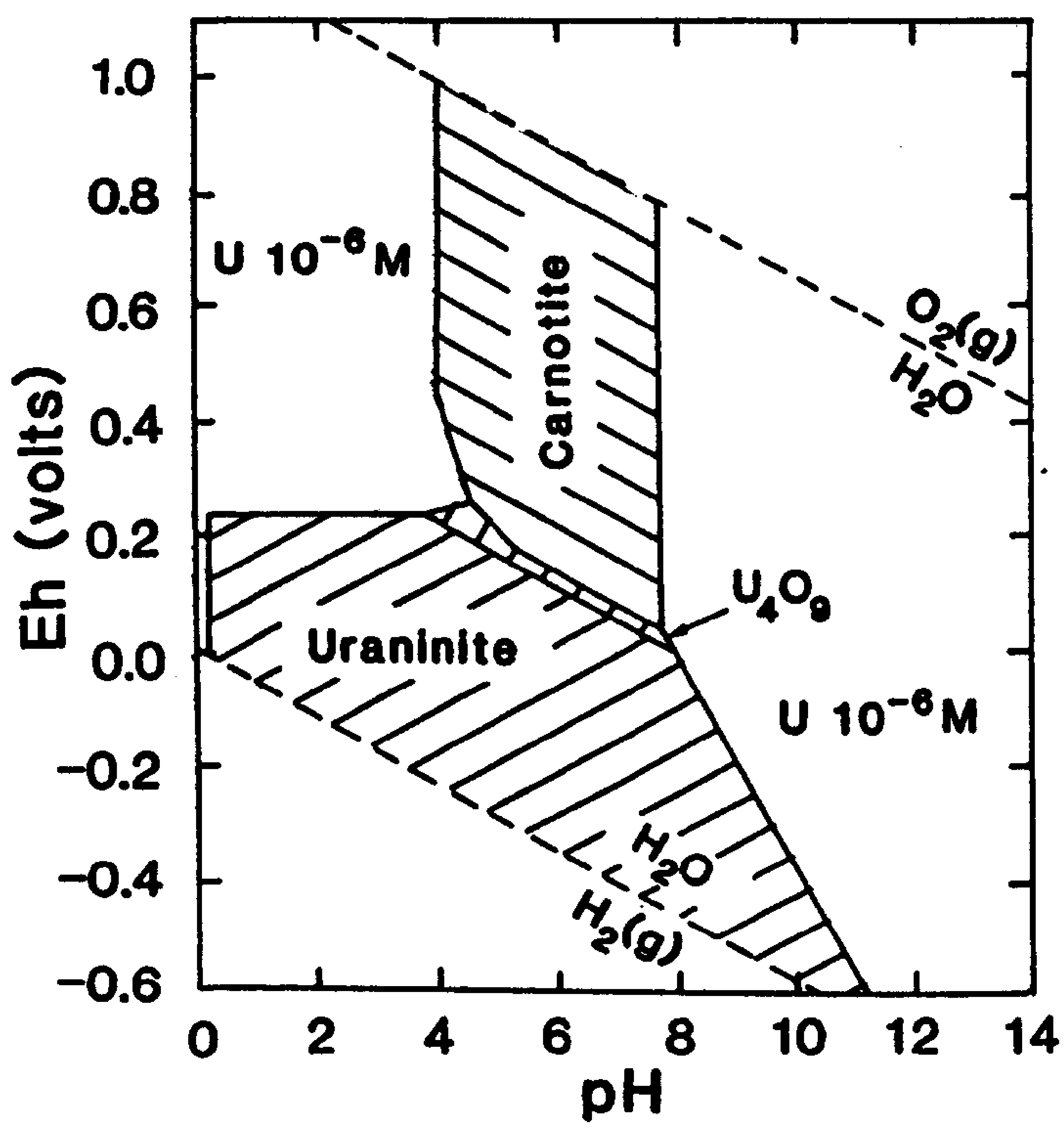


Figure 8.19. Eh-pH relations for uraninite and aqueous solution with vanadium present. Uranium can now be precipitated under oxidizing conditions too. (After Langmuir, 1978, Figure 19).

Origin of Roll-Front SSUV Deposits.

These are elongate and occur intermittently along an interface between oxidized and reduced sandstone (see Figure 8.20). They are thought to have originated from the downdip movement of oxidizing U-carrying groundwaters, or from remobilized U already contained within the sandstones. The U minerals were then precipitated when the waters encountered reducing conditions. These reduced facies are probably controlled by organic matter or sulphide species. Examples include the Tertiary and Cretaceous beds of Wyoming, South Dakota and southern Texas.

Origin of the Tabular SSUV Deposits.

Tabular bodies are discrete masses of U surrounded by reduced sandstone which are themselves scattered throughout oxidized sandstone. They are more closely related to large carbonaceous deposits of organic debris than roll-front deposits. There are two common theories as to their origin;

1) U was originally deposited together with sandstone under reducing conditions with later remobilization of secondary importance.

2) Later oxidizing solutions containing dissolved U precipitated this U on encountering pockets of sandstone rich in organic matter. Examples of tabular SSUVs include those deposits of the Jurassic and Triassic beds of Colorado, Utah, Arizona and New Mexico.

One of the most important questions concerning the origin of SSUVs is the nature of the reductant which causes reduction of U^{6+} to U^{4+} and hence precipitation of U deposits. Throughout much of the Colorado Plateau the carbonaceous material of tabular deposits is coalified wood (Edwards and Atkinson, 1986). However in the Grants Mineral District the U is associated with tabular layers of organic-rich material which is thought to be humate and is regarded as epigenetic (Adams et al, 1978).

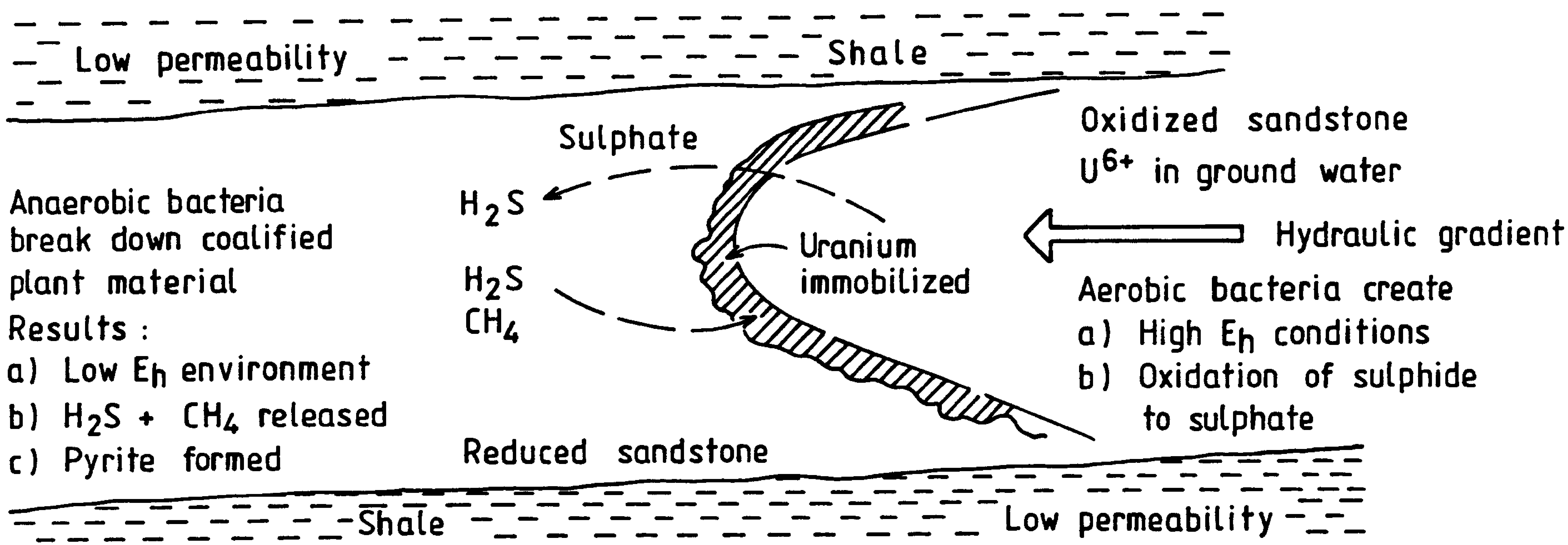


Figure 8.20. Idealized cross-section illustrating mechanism of formation of roll-front SSUV deposits. (After Edwards and Atkinson, 1986, Figure 9.12).

Another query concerns the two different forms of SSUV deposits. However it appears from the results that these two forms do not show a disparity in latitude so the climatic influence is unlikely to be one which dictates whether a roll-front or tabular SSUV deposit develops i.e. these are parochial features of the depository whose spatial differences are subject to the same climatic controls.

The major role of organic matter is to fix U by reduction of uranyl species in solution. Uraninite and stable organo-uranyl compounds are thus formed, with concommitant dehydrogenation of the organic matter. The influence of organic matter, regardless of its origin, is obviously great upon the development of SSUVs. This influence manifests itself in a number of different ways i.e. humic acids and humates, bacterial activity, particulate organic matter and larger deposits of carbonized wood. For tabulate deposits (Brookins, 1976) has shown that no uranyl species are thermodynamically stable in the presence of carbonaceous matter. Uranyl ions were thought to be transported in relatively reducing groundwaters rich in sulphate and with minor hydrogen sulphide. These ions vertically diffused into a more reducing zone or were vertically mixed with groundwater locally slightly discordant with the long axis of the tabular zone. A characteristic feature of pebbly sandstones in the U deposits of the Phu Wiang Basin, Thailand is the presence of carbonized and silicified plants and other organic remains (Gocht and Pluhar, 1981). The plant material occurs as debris, the grade of mineralisation evidently increased with the content of organic material.

Anaerobic bacteria combine with C and H in organic matter to extract O from sulphate ions in groundwater. A resulting waste product is H_2S which acts as a reductant. This is another role of the biomass in SSUV deposit development. For roll-front deposits the most important factor in the formation of SSUVs is the different roles of separate populations of bacteria on either side of the front (Rackley, 1976). Within the reduced zone anaerobic bacteria breakdown carbonaceous material to form CO_2 and H_2 which produce low Eh conditions and result in U precipitation. A separate species of bacteria (Desulfovibrio) utilizes CO_2 and inorganic sulphate to create methane and hydrogen sulphide. Hence organic matter acts as reductants for U. Within the oxidized zone aerobic bacteria create high Eh conditions within which uranyl species are the stable form of U and sulphide minerals are oxidized to sulphates. Soluble sulphate migrates into the reducing zone thus contributing to the continuing cycle of the geochemical cell.

In the Colorado Plateau two types of carbonaceous matter occur within the sandstone and conglomerate (Kimberley, 1978b). Those deposits associated with particulate organic matter which was co-sedimented with sand and gravel (e.g. Triassic, Chinle Formation) are relatively less voluminous than those associated with humates (e.g. the Jackpile deposit, near Laguna, New Mexico) which have precipitated from through-flowing ground water. Turner-Peterson (1985) also proposed that humate, as a pore-filling organic material closely associated with primary ore played an important role in the development of U deposits in the Grants mineral region of New Mexico. The basic premise was that humate originated as humic acids dissolved in the pore waters of greenish-grey lacustrine mudstones. During compaction waters carried the humic acids into adjacent sandstone beds where the humates were deposited. The close association of uranium with pore-filling organic matter and the high probability that localization and concentration of U was controlled entirely by the organic material place considerable significance on the origin of the organic matter for understanding the genesis of the SSUVs. A humic, as opposed to a hydrocarbon, origin for the organic matter is attractive because of the strong affinity of humic substances for U (Turner-Peterson, 1985).

Because humate is apparently one of the main ore controls for primary SSUVs an important question in modelling ore genesis is the specific source of the humate. Squyres (1980) proposed that it was derived from indigenous organic detritus (e.g. carbonized and silicified logs). Indeed most workers believe the humate was derived from within the Morrison Formation itself. However Granger (1968) proposed that erosion subsequent to the deposition of the Morrison Formation resulted in swampy areas which may have supplied the dissolved humic acids. The timing of such a humic-acid influx has been suggested as being:

1) syngenetic (Granger et al, 1961; Squyres, 1980) - although this fails to explain the observed infilling of secondary voids in detrital grains by humate,

2) closely following Morrison sedimentation (Moench and Schlee, 1967),

3) after deposition (Granger et al, 1961; Galloway, 1980). Granger (1968) proposed that the humic acids were introduced during deposition of the overlying Dakota Formation some 30 - 40 m.y. later,

4) both 2) and 3) with several intervening periods (Brookins, 1976).

A number of different climatic environments have been proposed for SSUV deposits. Wright (1979) maintained the host sandstones were deposited by streams in a semi-tropical humid region where the characteristic luxuriant growth of these areas was apt to be washed into the streams and buried. He proposed that the level of the water table was crucial to the preservation of the plant matter. A continuation of humid conditions would have kept the water table level high and preservation would be more likely. However destruction of plant matter results from a drop in the water table which may have been caused by climatic changes from humid to arid conditions or by a shift from sedimentation to erosion. However Wright (1979) considered that swamp and bog environments with abundant plant growth that may produce coal are not favourable to the development of SSUVs. In contrast to this Barthel (1974) proposed a swampy environment of SSUVs in the Lodève Basin, France. Haynes (1975) described SSUVs in the Lake Frome area of Australia which now experiences an arid climate with an average rainfall of 17cm. However he emphasized that although the rivers in this region are ephemeral, during floods they are still capable of transporting large volumes of debris. He suggested that such ephemeral rivers could achieve much more erosion and deposition than more evenly regulated streams in less arid areas.

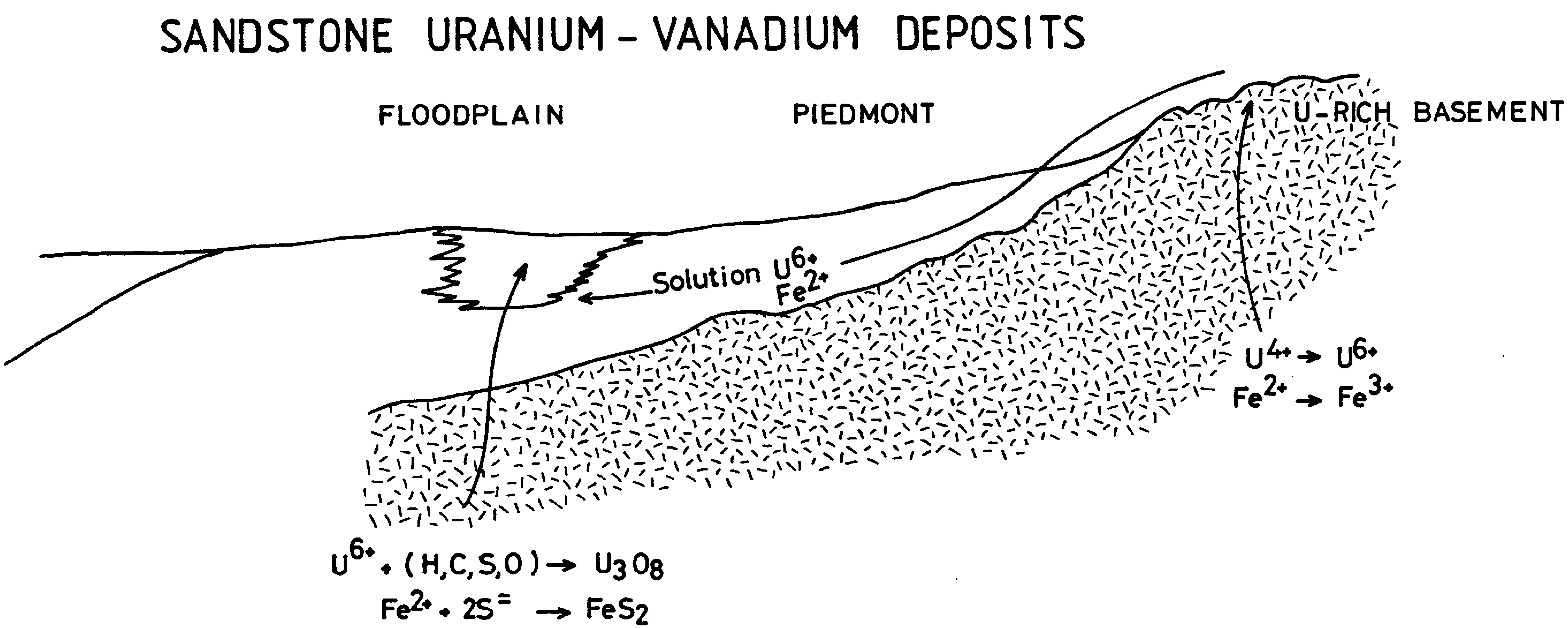
One of the main points raised by most workers (e.g. Barthel, 1974; Langford, 1974) is the importance of alternating wet/dry periods in the development of SSUVs. An entirely arid or very humid environment is not thought to be conducive to U precipitation. The rainfall must be sufficient for the development of a fluvial system and luxuriant plant growth, although not too great as to remove all the plant debris. Langford (1974) mentioned that on an aggrading continental surface organic material can be preserved by burial whereas on an erosional surface plant remains decay in place or are swept away. Intermittent rainfall, as in wet/dry zones, would increase channelization which is vital in providing suitable depositional environments for SSUVs and would also focus groundwater migration routes. The climate during deposition of U is also important as groundwater flow through the permeable sandstones is essential so a completely arid climate would not be favourable. During the wet period the fluvial channels were developed, sands deposited and organic debris laid down, hence preparing the conditions necessary for SSUV deposition. However at the onset of the dry period, oxidizing U-rich waters invaded the reducing environment and U was deposited. McLaurin (1979) proposed that a ubiquitous reducing environment during humid and tropical climates would preclude the solution and transport of U. Using this as a premise he predicted that the most favourable latitudes in which SSUVs would occur would be from 30° to 40° latitude where climatic conditions would be more temperate to subtropical.

The source of U is considered to be weathering of the adjacent hinterland in the majority of cases. The most common sources include tuffs and volcanics (Gocht and Pluhar, 1981), granites and keratophyric basement (Lukacs and Florjancic, 1974). A climate of high temperatures and relatively low rainfall would promote the weathering of such minerals as zircon, monazite, glass and hence release U, so an origin in the colder, high-latitude regions is unlikely.

All the constraints set out above on climatic conditions favourable for SSUV deposition are reflected in the distribution of SSUVs (see Chapter Six, section 6.3.5). There is an obvious suppression in the number of SSUVs found in the hot, arid zones whereas peaks in the distribution occur in the equatorial and temperate warm, humid rain belts. There is a peak around the equator in which deposits apparently formed in a continually tropical, wet zone. Perhaps the chemical weathering essential to release U, the development of fluvial channels and luxurious plant growth were the dominant controls on formation. The lack of deposits in a dry zone may be compensated for by efficient drainage in some tropical areas.

Rawson (1975) suggested that the conditions found in sabkhas could cause the deposition of U minerals in the carbonate material which progrades onto decaying algal mats etc. He proposed that the U-bearing groundwater would be drawn upwards through the zone beneath the sabkha. The U would be deposited when the reduced environment of decaying algal matter would be reached e.g. Todilto Formation of Jurassic age in the Grants region, New Mexico where organic layers in limestone have been replaced by uraninite. Such a theory may explain the few deposits which do apparently occur in the lower latitudes where a more arid climatic regime exists.

The evolution of land plants in the Silurian may have had two favourable effects on SSUV deposition, according to Wright (1979). Firstly it became possible for significant volumes of plant matter to be incorporated into fluvial depositional systems. Secondly the land plants tended to slow down the effects of rapid run-off and erosion and hence more time was available



CLIMATIC ENHANCEMENTS

1. High T, relatively low rainfall promotes weathering of Zircon, Monazite, Glass etc.
2. Intermittent rainfall increases channelisation & focuses groundwater migration routes
3. Wet periods promote abundant vegetation cover, development of fluvial system & deposition of permeable channel sands.

Figure 8.21. Idealized model of the formation of SSUV deposits showing source, transport and precipitation of uranium.

for chemical reactions during intense tropical weathering. Certainly roll-front and tabular SSUVs do appear to be more common in Phanerozoic than Precambrian times, although such a distinction may be due to selective preservation of younger deposits. It may also be due to the failure to recognize these deposit types in older rocks. Stanworth and Badham (1982) emphasized that some uranium deposits from the East Arm, Great Slave Lake, Canada were secondary diagenetic in origin and were not detrital accumulations as had previously been thought.

In general the main model of SSUV genesis (see Figure 8.21) is that U is leached from either granite or tuffs and transported along permeable palaeochannels in sandstones by oxygenated meteoric water. Uranium concentration is related to either adsorption by carbonaceous material or rapid changes in redox conditions. The main factors include uplift of the source area with the eventual exposure of uraniferous rocks, a climate which favours chemical weathering and luxuriant vegetation cover, and the preservation of permeable sediments with an appropriate content of organic matter, pyrite and clay minerals.

8.2.3 Placer Deposit Types.

Placers may be defined as surficial mineral deposits formed by mechanical concentration of mineral particles from weathered debris. The mechanical agent is usually alluvial but it may also be marine, aeolian, colluvial (creep) or glacial. As a general rule placer formation is favoured by prolonged sediment reworking causing heavy mineral concentration while transporting other sediment downstream. Hence exploration for eluvial, colluvial and fluvial deposits should, in part, be concentrated in looking for evidence of a prolonged weathering regime and a dynamic transport system with favourable conditions for deposition (Macdonald, 1983). Placers might therefore be expected to be associated with surfaces of fluvial degradation in drainage basins and in alluvial fan sequences of depositional basins (Schumm, 1977). Placers may also be found in areas where multiple periods of reworking have occurred due to sea level fluctuations which may increase the chance of enrichment (Henley and Adams, 1979).

8.2.3.1 Placer Gold Deposits (PLAU).

According to Henley and Adams (1979) the formation of PLAU deposits involves two complex stages. The first stage is the hydrothermal concentration of trace Au in the crust into epigenetic deposits (e.g. quartz veins) and does not seem to have been influenced by climatic conditions. The weathering, erosion and mechanical concentration of the Au into alluvial deposits constitutes the second stage and is clearly related to climatic processes. This includes the secondary mechanical concentration or resorting of earlier auriferous gravels e.g. California. For all examples the basic control upon PLAU formation is that the supply of Au depends on the supply of sediment from which it can be concentrated. In the north Westland area of New Zealand the richest placers are to be found downstream of terminal moraines formed during Pleistocene glaciations. The auriferous gravels of California were first accumulated on a deep weathered and lateritized peneplain as glass sand and clay deposits. Reworking of these gravels during Pleistocene valley glaciations and their associated fluvioglacial transport resulted in deposition in terrace formations (Henley and Adams, 1979). In both the above examples climatic conditions have obviously played an important role in the initial accumulation of the Au supply.

Although an original sediment supply is vital for PLAU formation, Schumm (1977) emphasized that fluvial reworking of eroded sediment to form placers is more likely during a period of low sediment supply. Hence very high rates of erosion are not particularly favourable for placer development as the large volume of sediment produced reduces the opportunity for mineral concentration. Such an alteration of periods of rapid sedimentation with those of lower erosion rates during which sediment is reworked could be produced by changes in tectonic movements of the source and depositional areas (Sutherland, 1985). Indeed, Henley and Adams (1979) noted an association of PLAU deposits with Mesozoic and Cenozoic mountain ranges particularly around the Pacific Ocean. These areas are typified by high relief, considerable tectonic and igneous activity and variable rates of erosion and deposition.

An indirect effect of water volume on fluvial placer formation is upon the distribution of Au within a palaeodrainage system e.g. the Carbon Leader Placer, Witwatersrand, South Africa (Buck and Minter, 1985). The central area of the system was characterized by greater fluvial current velocities than the marginal zones, hence were coarser-grained, and higher concentrations of Au occurred. The margins had weaker fluvial processes and so comprised finer-grained facies with poorer concentrations of PLAU.

8.2.3.2 Placer Diamond Deposits (PLDI).

As the primary source of diamonds in economic quantities are the igneous intrusions known as kimberlites (Dawson, 1980) it is to be expected that PLDIs have a strong geographical association with these sources. Kimberlites are usually found within cratonic areas and there is a clear correspondence of PLDIs with these regions (Sutherland, 1985). Therefore PLDIs occur in areas with slow and extended up- or downwarping with little erosion. Obviously the degree of erosion on cratons depends upon the time period under consideration (Sutherland, 1985) and also the prevailing climatic conditions. An examination of the distribution of PLDIs with regard to cratonic areas also shows a notable preference for the lower latitudes within these areas e.g. there is a marked lack of PLDI deposits in the Canadian and Scandanavian

Shields suggesting an absence of suitable source rocks, whereas the majority of deposits occur in the South American and South African lower latitude cratonic regions. Berbert et al (1981) described PLDI from western Minas Gerais and mentioned the dominance of intense chemical weathering in the area with extreme weathering surfaces. However even though PLDIs occur in many continental clastics in Brazil from the end of the Lower Proterozoic to Recent times the source remains elusive (Badham, pers.comm.).

The effect of changing magnitude and frequency of discharge variations in humid tropical river systems in response to major climatic shifts has been stressed by Baker (1978). Hall et al (1985) have applied this to explain the formation of Birim (West African) PLDI deposits in terms of palaeoclimatic changes during the late Quaternary suggesting that Quaternary environmental changes produced associations between diamond grades and chronostratigraphic units comprising floodplain gravels.

As with PLAU deposits water volume has an effect upon the morphology of PLDI deposits. A notable feature of the Birim Placer, West Africa is the association of medium- to coarse-grain size diamonds with coarse-grained e.g. pebble and cobble, gravels (Hall et al, 1985). This suggests that diamond concentration is enhanced by trapping and protective storage in the coarser gravels which have larger spaces between the clasts into which larger diamonds can penetrate (Prokopchuk, 1969). Fluvial transport also has an influence upon deposit formation as diamond size distribution is progressively modified with increasing travel distance from the source. The quality of the diamonds also improves as inferior types are destroyed.

It would seem that PLDI distribution is affected by a combination of the distribution of source rocks in cratonic areas on the globe and climatic conditions i.e. humid tropical and subtropical zones where fluvial transport and intense chemical weathering dominate.

9.3.3 Placer Tin (PLSN), Placer Oxide (PLOX) and Placer 'Other' (PLOT) Deposits.

Placer Tin Deposits.

Cassiterite (SnO_2), the most important tin ore mineral, is hard, heavy and highly resistant to weathering (see Chapter Three, Table 3.6). It therefore tends to concentrate naturally in superficial deposits derived from tin-bearing granites (in which it occurs as a primary mineral) and its associated metamorphic deposits e.g. the southeast Asian tin belt. Apart from the obvious pre-requisite of a high initial concentration of tin from a tin-bearing source rock, the other factors which have led to large deposits include deep, rapid, tropical weathering which has released large quantities of primary tin and the preservation of placers resulting from low terrain by low velocity streams (Hails, 1976). Cassiterite may be concentrated by chemical weathering which selectively removes feldspar, and to a lesser degree quartz, downslope. Hails (1976) considered that such tropical weathering was the most important process involved in the formation of the southeast Asian tin belt. The PLSN deposits of the Bangka and Billiton Islands are thought to have formed in a different way. It has been proposed (Adam, 1933; Aleva, 1985) that these deposits were the normal residue found on weathered surfaces within the humid tropics so they were more residual in nature with relatively little influence of fluvial transport. They were not considered to be the result of chemical weathering and subsequent erosion and mechanical transport e.g. some deposits in the Malaysian Peninsula and Thailand (Aleva et al, 1973).

Placer Oxide Deposits.

Climate is indirectly responsible for many PLOX deposits e.g. ilmenite in late Pleistocene beach deposits on the west coast of New Zealand which were deposited during interglacial high stands (NZ, DSIR, 1969). The occurrence of monazite in some deposits has been attributed to its relative stability in the weathering profile (Hails, 1976). It is resistant to mechanical abrasion so concentrates at the site of weathering and in streams e.g. in North and South Carolina, USA. Monazite tends to be disproportionately concentrated in alluvial gravel in accordance with the hypothesis that heavy resistate accessory minerals are proportionately more abundant in gravel than in sand, silt or clay (Overstreet et al, 1968).

The concentration of titanium placers in beach deposits of South Carolina and Georgia is thought to be due mainly to the prevailing climatic conditions which are conducive to the development of thick, well-drained saprolites over granitic or gneissic source rocks (Puffer and Cousminer, 1982). By analogy the Ti-Fe PLOX deposits of the Cohansey Formation, New Jersey are also thought to have been formed under conditions found in South Carolina and Georgia at present. The development of saprolites is considered to be an important early stage in the origin of Ti-Fe oxide beach placers along the west coast of Australia and West Africa (Force, 1976). The relatively insoluble Ti-Fe oxides were concentrated in an easily eroded residue during saprolite development whilst more unstable minerals were weathered and removed. Silicates e.g. garnets, which may contaminate the PLOX deposit are also weathered during saprolite development (Dryden and Dryden, 1946). The beach sands from Charleston, South Carolina to Miami, Florida, USA contain 1% garnet compared with 26% for northeastern Atlantic beaches such as Long Island (Martens, 1935).

Placer 'Other' Deposits.

One of the major factors in the concentration of Be in PLOT deposits is the extreme insolubility of the most abundant Be minerals e.g. beryl, chrysoberyl and phenacite. However it is important in the formation of such PLOT deposits that the Be-bearing silicate minerals have not been transported for large distances from the source region. If such weathering and transportation does occur over a long distance and an extended period of time the Be-bearing minerals would decompose and the Be would be adsorbed onto Al-rich clays (Rupasinghe et al, 1984). Hence such PLOTs as emerald deposits are more likely to be found in humid tropical or subtropical areas with locally restricted transport (which may be due to dense vegetation cover).

Once again source rock is an important development in PLOX and PLOT development but climatic conditions and depositional environment are also other important influences. The necessity of saprolite development for PLSN, PLOX and PLOT deposits explains the lack of such deposits in the warm, arid zones as shown in the results (Chapter Six, section 6.4.3).

Sutherland (1985) emphasized that climate influenced many of the variables which are relevant to placer formation e.g. weathering, rates of erosion, nature of sediment supply and opportunities for sediment reworking. These factors were discussed in terms of broad morphogenetic zones and for this reason this discussion is summarized below.

Humid Tropical Regions: These are characterized by high temperatures, constant stream flow and a dense vegetation cover which inhibits the direct influence of many mechanical erosional processes. If semi-arid conditions with fluvial activity sufficient to erode the decomposed regolith and transport liberated heavy minerals alternated with humid conditions, increased vegetation cover would reduce sediment supply and encourage reworking of material from fluvial systems developed in semi-arid periods. The extensive and deep chemical weathering of the bedrock which is characteristic of humid tropical regions influences;

i) the separation of weather-resistant minerals,

ii) the formation of a dominantly clay-sized regolith,

iii) chemical denudation and mass loss of as much as 40% of the rock by solution prior to any mechanical erosion (Thomas, 1974).

Semi-Arid to Arid Regions: These zones experience very high run-off rates due to irregular but intense rainfall and limited interception by sparse vegetation cover. Rock weathering is therefore dominated by mechanical rather than chemical breakdown. Fluvial activity is effective in these areas because of the high sediment load carried in rivers creates a relatively high density fluid in which heavy particles can be more easily transported. However such fluvial activity is very ephemeral and reworking is less likely than in other environments.

Humid Temperate Regions: These are more common in the northern than in the southern hemisphere because of the relative distributions of land and sea. Such areas are characterized by deep arenaceous weathering profiles with very little alteration to clay minerals (<10%) and a perennial river flow which is strongly moderated by the vegetation cover. The arenaceous weathering profiles produce large volumes of bedload sized material which may rapidly dilute heavy minerals from source. Also there is a low overall sediment yield from these areas because of the vegetation cover (Strakhov, 1967) so release of placer minerals from the bedrock is restricted. In general these areas are not those with optimum characteristics for placer development (Macdonald, 1983).

Cold Non-Glacial Regions: Such areas are underlain by continuous or discontinuous permafrost, have a thin soil layer, limited vegetation cover and experience a seasonal snow melt. Hence there is a brief period of highly concentrated fluvial activity in late spring and early summer and weathering is dominated by mechanical processes. An important influence of permafrost is that it prohibits widespread "stripping" of pre-existing deposits or regoliths created during the Tertiary when release of primary minerals was more effective because of the milder climate. However if these earlier sediments have been reached by the periglacial fluvial system, rich placer deposits may develop e.g. Yukon River, Alaska (Boyle, 1979).

<u>Glacial Regions:</u> High grade placer deposits can occur within formerly glaciated areas. There are three points worthy of note;

i) Generally fluvioglacial sediments are deposited rapidly with little sorting, and are usually unable to rework sediments of established drainage patterns so heavy mineral concentration is not common. Some deposit examples include PLAU in bench deposits near Nome, Alaska (Cobb, 1973).

ii) In ice-shed areas and towards the margins of ice sheets, the ice and glacial sediments may play a protective rather than an erosive role. Hence deeply weathered bedrock or pre-existing placers may be preserved e.g. in the Caribou area of British Columbia (Boyle, 1979).

iii) Glaciers may disperse heavy mineral with little sorting over considerable distances producing large volumes of very low grade sediment that may be reworked by post-glacial processes e.g. PLAU deposits in the North Saskatchewan River, Canada (Boyle, 1979).

The importance of climate to the formation of placer deposits cannot be stressed too greatly as each of the five morphogenetic regions is capable of producing placer deposits, although some e.g. humid tropical zones, are more favourable than others e.g. humid temperate and cold non-glacial regions. Evans (1981) proposed that the normal processes of lateritization may be important in upgrading the Au content of ultrabasic rocks and therefore increase their suitability as a source for placer deposits. In addition Welch et al (1975) suggested that rapid uplift of lateritized terrain under relatively wet climatic conditions was an important pre-requisite to the accumulation of heavy mineral sands. Puffer and Cousminer (1982) utilized palynological evidence from a lignite band to show that at the time of formation of Ti-Fe oxide-rich mineral sands in the Lakehurst area, New Jersey the region experienced a subtropical climate. A humid climate is favourable for eluvial placer development as enrichment in placer minerals is partly caused by the removal of soluble minerals by groundwater and partly by the transport of the lighter minerals by running water and wind action. Climate also plays an important role in the formation of fluvial placers. Under dry conditions the valley system can be fed from the products of rock decomposition, whereas in a wet climate the feed to the valley is restricted to the products of the river system erosion, although land slides may contribute. In a hot/wet climate the land is protected by a cover of vegetation which will restrict the movement of soil and rock towards the drainage system. Such humid, subtropical conditions are ideally suited to saprolite development which has been shown to be very important in the formation of some placer types (Fairbridge, 1968). The solubility of silica is greatly increased by increased temperatures so saprolites are more likely to develop in warm than in cool climates. (A temperature increase from 10° to 30°C will increase the solubility of quartz by a factor of 1.9; Siever, 1962).

Changes in climatic conditions have also been cited as influential controls on placer accumulation (e.g. Hall et al, 1985; Sutherland, 1985). For example variations in sea level during Plio-Pleistocene times due to climatic changes resulted in several periods of marine transgression and regression throughout the globe. These fluctuations led to the reworking of extensive zones of clastic sediments and created placer beach sand deposits in many areas (Edwards and Atkinson, 1986).

8.2.4 'Other' Deposit Types.

8.2.4.1 Manganese Formations (MNFM).

These examples are confined to Phanerozoic sedimentary Mn deposits in which metals were mainly derived through the weathering of rocks. They occur in present day marine and lacustrine basins. There are two main varieties of

MNFM which are of particular interest to this research; those associated with clastics and those associated with carbonates.

Clastic-hosted MNFMs.

The major concentration of clastic-hosted MNFMs occurred in the Oligocene in the southern European region of the USSR: classic examples are to be found at Nikopol and Chiatura. Nikopol-type deposits are localized as beds of Mn oxides and carbonates near the base of the marine ore-bearing sequence. They are characteristically associated with silt-clay or, more rarely, sandstone. In the Mn deposits of Nikopol and Bol'shoi Tokmak the ore-bearing sequences occur transgressively on the crystalline basement. The ore horizon hosts a variable shallow-water faunal assemblage e.g. gastropods, crabs, sea-urchins, solitary corals. Surficial enrichment of ore beds, particularly of carbonate ores by weathering is common. This oxidation-enrichment zone was formed in the Oligocene before the accumulation of the overlying plastic green clay beds (Gryaznov and Danilov, 1976). The Mn deposits in Georgia, USSR (e.g. Chiatura) contain an average of 35% Mn with oolitic (and pisolitic) varieties common. Higher oxide ores are interlayered with coarser-grained clastic rocks that show a minimum content of organic carbon. Conversely carbonate ores are interlayered with very fine-grained detritus which have the richest organic carbon content. Edilashvili et al (1974) concluded that in the western part of the Chiatura deposit Mn oxide ore deposition occurred during a period of minimum terrigenous input on a submarine rise in shallow water.

Most of the studies of the Nikopol-type MNFM deposits concluded that they were formed from products of a weathered zone of older rocks in a stable tectonic condition (Roy, 1981). Varentsov and Rakhmanov (1980) considered the accumulation rates of the ore to be low, Mn supplied from the weathered zone as organo-metallic compounds and precipitated by oxidation. The water of the

Chiatura Basin was assumed to have a low salinity (Sokolova, 1964). However the presence of stenohaline bivalves, sea urchins and solitary corals in the ore horizons of the Nikopol Basin have been interpreted as being indicative of normal salinity. Both the Nikopol (Selin, 1964 cited in Gryaznov, 1970) and Chiatura (Sokolova, 1964) Basins are thought to have had a sub-tropical temperature regime. Hence Mn deposition at both sites occurred in a humid climate. However the Mangyshlak deposit is thought to have been deposited at the boundary of humid and arid zones and the ores north of the Urals were deposited in more temperate climates (Varentsov, 1964). Shterenberg et al (1965) concluded that ore deposition in all the Oligocene basins of the USSR coincided with a sharp change in climate from humid subtropical to cooler and drier temperate conditions. This is supposed to have led to separation and concentration of Mn with respect to Fe at source and/or in the depositional environment due to changes in the weathering process. The altered climatic conditions would also allow the prolific development of Mn-oxidizing organisms. Thus these huge MNFM deposits were formed by a coincidence of suitable climatic, biological and physiographical conditions (Roy, 1981).

Carbonate-hosted MNFMs.

These bodies may be either oxides or carbonates and are associated with limestone-dolomite formation and red carbonate-terrigenous formation (Roy, 1981) e.g. the Permo-Triassic MNFM deposits of Narguechoum, Morocco. The ore horizon is successively overlain by red arkosic sandstone, red clay and finally a Lower Jurassic massive, sandy dolomite. However dolomite intercalations in the ore horizon are absent but for a thin layer at the bottom of the ore bed. This deposit is therefore confined to a red, terrigenous formation apparently deposited in an arid climate and an evaporite facies (Roy, 1981).

MNFMs related to the transgressive limestone-dolomite sequence are found at Imini (Upper Cretaceous) and Bou Arfa (Lias) in Morocco. Two or three persistent Mn-oxide beds occur within the dolomite and at the contact of sandstone and dolomite horizons. The thick dolomite sequence was deposited in deeper parts of the basin, passing into near-shore coarse detrital continental-type red-beds. Once again ore deposition is considered to have taken place in an arid climate in evaporite facies (Varentsov, 1964).

The bedded Mn orebodies of the Permo-Carboniferous, Lower Lias and Upper Cretaceous formations of Morocco are mostly accepted to be of sedimentary origin, derived by the supply of Mn through continental erosion during marine transgression. Pre-Cambrian and Carboniferous Mn-rich hydrothermal vein deposits are thought to be the principal Mn sources (Roy, 1981). Deposition of these sedimentary ore bodies took place in an arid climate and associations with evaporites have been recorded. It has been suggested that Mn deposition occurred in shallow water coastal lagoons (Vincienne, 1956). In view of the low detrital component and the very high Mn/Fe ratio Stanton (1972) suggested a highly selective biological process operated in the shallow-water environment to produce these orebodies.

In the USSR carbonate-hosted MNFMs are found at Ulu Telyak in the western Urals. These deposits are considered to have formed in very similar conditions to those of Morocco. Mn deposition occurred during a marine transgression in shallow, lagoonal waters with a Mn source on the adjacent continent to the west (Roy, 1981).

At Groote Eylandt off the northern coast of Australia a dominantly pisolitic MNFM has formed under shallow marine conditions. The absence of Mn carbonate minerals is notable but Slee (1980) suggested that the present ores were derived from pre-existing carbonates. Beneath the ore beds is an unconformable manganiferous marl which is thought to have been eroded so producing Mn which was transported by rivers and precipitated at the shoreline. This fresh water to marine transition would result in a rise in pH from 6 to 8 and MnO_2 should be precipitated under these strongly oxidizing conditions (Maynard, 1983).

One group of MNFMs which have not been included in this research are those of the supergene type because too few examples were collected. However these deposits appear to be important to this research as they are found extensively in the weathering profiles of many countries, particularly in humid tropical and sub-tropical zones, so they appear to show a strong climatic control. It has been mentioned previously that the source of Mn for many MNFMs may have been weathering profiles of adjacent hinterland. The formation of these supergene deposits may illustrate how a pre-concentration of Mn can occur. Intense weathering in a tropical region with adequate drainage to give enrichment of Mn would obviously prove favourable to the subsequent formation of clastic- or carbonate-hosted MNFMs. These deposits are considered to have formed by in situ residual concentration of Mn in the country rock or through dissolution of the metal and its reprecipitation within the weathering profile. Sizable supergene Mn deposits occur in Precambrian rocks in Brazil, West Africa, South Africa and India. Relatively smaller deposits are to be found in Montana, USA; Urkut, Hungary; Janggun, Korea and Toyoguchi and Noda Tamagawa, Japan (Roy, 1981).

The following generalizations were amongst others given in Roy (1981) for supergene MNFMs from Bricker (1965), Thienhaus (1967), Weber (1973) and Lelong et al (1976). These points have been singled out because of their possible climatic connection;

1) Mn concentration is due to prolonged weathering in high-lying plateaus.

2) In the upper zones of the weathering profile, pisolites and nodules containing minor Mn associated with Fe-hydroxide and clay are found. The concentration of the majority of Mn as oxide crusts takes place in deeper zones. So the Mn-oxide ores are effectively separated from other weathering

products. However in the presence of stagnant water over impermeable weathered rocks the dispersion of MNFMs is hindered. Thus adequate drainage and a flowing water supply are vital.

3) The common agents of the supergene concentration process are meteoric water and air.

The process by which Mn oxides and hydroxides were formed from a protore at Moanda, Gabon which has undergone severe tropical weathering illustrates the development of supergene Mn deposits well (Weber,1973). Superficial oxygen- and CO_2-rich waters permeated through the protore and sulphuric acid was produced by oxidation of pyrite in the upper levels which yielded sulphates and CO_2 on reaction with carbonates. Sulpho-reducing bacteria were considered to be active in deep horizons in the presence of sulphates and organic matter with the production of CO_2 and H_2S. In such an environment the acidic water dissolved Mn present in the carbonates. With enrichment of bicarbonates and the increased pH in water, Mn-hydroxide with some carbonates were later precipitated at the bottom of the weathering zone.

The climatic controls on carbonate-hosted MNFMs appear to be similar to those of OOFEs i.e. low detrital input is favourable, together with a more arid climate in which carbonates i.e. limestones and dolomites can proliferate. However neither a completely arid, nor a cold, climate are conducive to MNFM deposition as intense chemical weathering under humid conditions with adequate drainage is needed to release Mn from the adjacent hinterland. Given the above the distribution found here is somewhat surprising in that some deposits occur in intermediate latitudes. However those are the clastic-hosted deposits of the USSR and it appears that more humid conditions are necessary for their development than for carbonate-hosted MNFMs as they do not have the close association with evaporites. As mentioned earlier it has been suggested that the formation of these deposits has been linked to climatic changes which encourage biological activity etc.

8.2.4.2 Laterite Deposits (LATO).

Laterites and bauxites are the extreme end residual products of special climatic conditions. They are formed under oxidizing conditions in areas of low or moderate topographic relief with a minimum of erosion (see Figure 8.22). Laterite consists essentially of hydrated iron oxides; bauxite of hydrated aluminium oxides, although aluminous laterites and ferruginous bauxites are found. The most common impurity in both is silica. Laterites (and bauxites) are widely distributed throughout the humid tropics and subtropics (e.g. Brazil, Jamaica, West Africa). They may also occur in temperate latitudes (e.g. the northern shores of the Mediterranean, Oregon, USA) but they are never found in regions which experience a prolonged cold winter (Greensmith, 1978). The climate should also be persistently warm. Hence the origin of laterite deposits is again dominated by chemical weathering and the amount of rainfall experienced by a region. Slightly different factors dictate whether laterite or bauxite deposits are formed. These include parent rock composition and variations in climatic conditions.

Aluminium-rich Laterites i.e. Bauxites.

Aluminium bauxites are residual soils resulting from extreme leaching in a humid tropical environment. For example the Trombetas bauxites of the Amazon Basin, Brazil occur in an area where rainfall is seasonal with an average of 2.5 metres with 70 to 80% occurring during the wet season. However temperature shows little seasonal variation because of the latitude and the diurnal range is only 10°C (Greig, 1977). Bauxites require high drainage rates with sufficient drainage to allow Na, K, Ca and Mg to be removed e.g. in the Trombetas region the clay which constitutes the uppermost layer of the weathering profile is permeable and drains rapidly even after heavy rainfall.

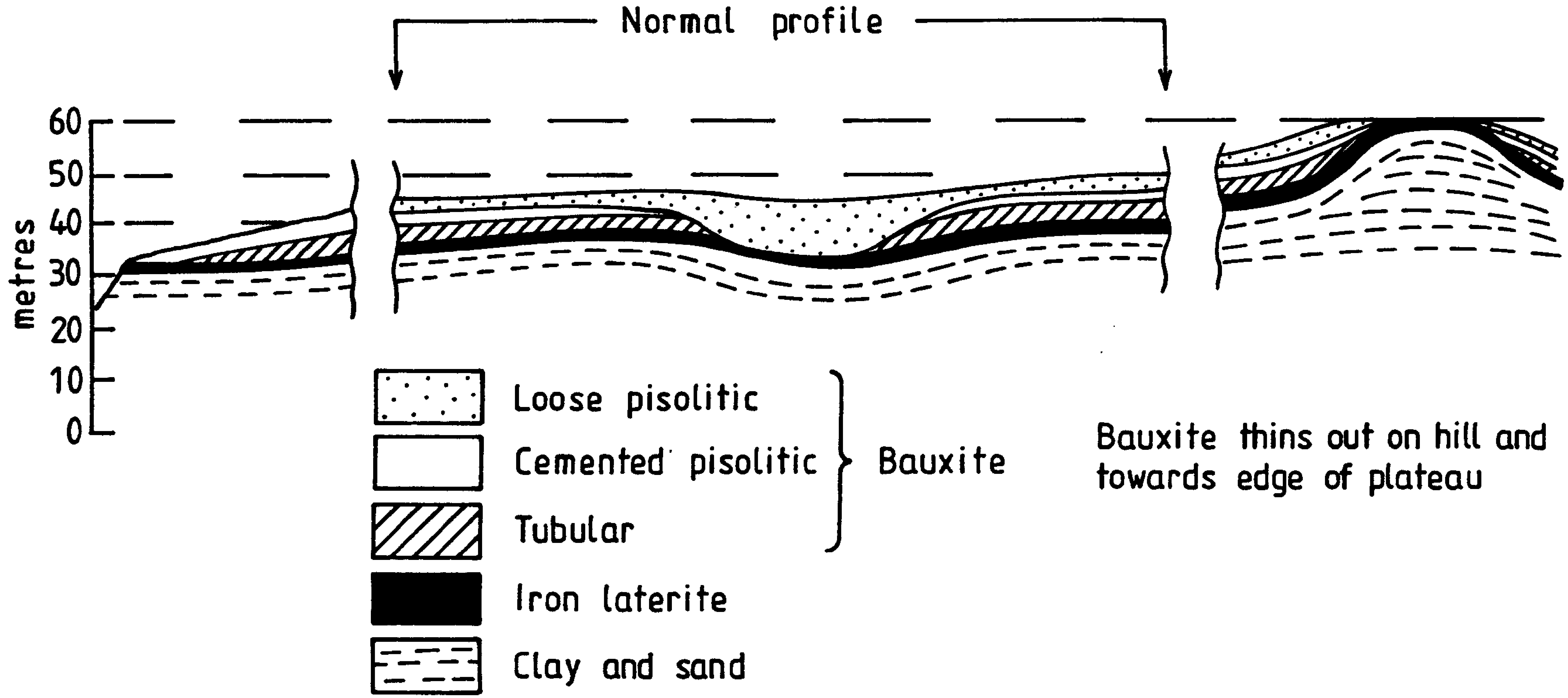


Figure 8.22. Schematic cross-section of laterite capping on a plateau of low to moderate topographic relief. (After Edwards and Atkinson, 1986, Figure 7.6a).

Bauxites are thus favoured by periods of intense chemical weathering but little erosion so that residual products may accumulate. The development of bauxites is also influenced by the mineralogy of the parent rock, although this is not quite as important a factor as with laterite deposits. If the silica is contained in aluminosilicates (e.g. feldspars or clays) it passes into solution relatively readily and hence is leached from the developing soil profile under humid conditions. In contrast it has been shown that quartz dissolves extremely slowly and so persists in tropical soils. Therefore bauxites tend to form over quartz-poor rocks such as syenites, basalts and limestones (Maynard, 1983).

To maintain the enrichment of Al relative to Fe a mechanism for separation is needed. A low-Fe bauxite may simply be due to a low-Fe parent rock. But gibbsite is more soluble than goethite so low-Fe bauxites may be formed from a high-Fe parent rock (e.g. basalt) by leaching of the Al and reprecipitation lower down in the profile. Indeed many basins have a cap of Fe-rich laterite over the bauxite zone (Maynard, 1983). Another mechanism of separation is by removal of Fe from lateritic soils under reducing conditions. The presence of Fe^{2+} minerals (e.g. siderite and chamosite) in some examples shows that such a reduction does occur (White, 1976). A potential source of reductant is the abundant plant debris characteristic of these climatic environments which occurs either within or immediately above the bauxitic layers.

Nickeliferous Laterites.

Nickeliferous laterite deposits comprise in situ lateritic weathering of a Ni-rich parent rock. Consideration of the relative abundances of Ni and Al (Table 3.3) shows that sedimentary Ni accumulations can only occur in association with pre-existing Ni concentration of ultrabasic igneous rocks. Ni enrichment in the laterite profile is therefore largely derived from olivine or serpentine. There are variations possible in the form of the weathering profile as it comprises four zones. The absence of certain of these zones is indicative of the prevailing climatic conditions, as described below;

<u>1) Profiles without an Intermediate Zone.</u> Weathering profiles without intermediate zones (i.e. nontronite or silica boxwork zones) are characteristic of the humid, tropical climate regimes and other areas with very high rainfall and a minimal dry season. They may result from extensive leaching so super-saturation conditions for smectite clays in the basal saprolite zone are not reached. Efficient drainage in a constantly wet profile would keep solutions sufficiently undersaturated to prevent the formation of smectite or quartz, except as a replacement for olivine.

<u>2) Profiles with an Intermediate Zone.</u> These typically occur in less humid, tropical climates with a marked dry season. Hence they result from relatively slow, inefficient leaching which leads to supersaturation conditions for smectite clays in the saprolite zone. The replacment of serpentine by smectite and quartz may occur if solutions remain in the weathering profile sufficiently long to achieve supersaturation in smectite or quartz. Such conditions would occur at a water table or during the dry season when solution movement downward is minimal. Obviously evaporative concentration would enhance this effect (Golightly, 1981). Thus nickeliferous laterites with nontronite or extensive silica boxwork zones may result from tropical wet-dry climates or elsewhere from retarded drainage (Golightly, 1981).

In the formation of bauxite and nickeliferous laterites perhaps the most essential climatic feature is a well-marked division of the year into wet and dry seasons. During the former extensive leaching of the rock occurs. During the dry season the solution containing the leached ions may be drawn to the surface by capillary action, evaporation occurs and the salts which remain may be washed away during the following wet season. Thus the zone from the lowest level to which the water table falls to the highest level it reaches is progressively depleted of the more easily leached elements e.g. Na, F, Ca and Mg. A solution containing such ions may have the correct pH to dissolve silica in preference to Al or Fe oxides. Thus the residuum consists mainly of these two oxides, the predominant one dependent upon the parent rock composition. An essential property of the rock is that it should maintain a porous structure during the leaching process, allowing free circulation of fluids.

The results show a broad spread of latitude of LATO formation, up to 50° latitude from the equator so the distribution covers the tropical, subtropical and temperate climatic regions. There are peaks in humid tropical areas (bauxites and nickeliferous laterites without an intermediate zone) and from 15° to 25° north and south i.e. the wet-dry tropical zone (nickeliferous laterites with an intermediate zone). Another hiatus occurs between 35° and 40° north i.e. the humid continental warm summer regime. This zone only occurs in the northern hemisphere as in the southern hemisphere water surface dominates over land area at these latitudes (Critchfield, 1983). The number of LATO deposits drastically reduces at latitudes greater than 45° from the palaeoequator because the temperatures are too low to maintain the high rates of chemical weathering and the precipitation is too low for the huge water volumes necessary for extensive leaching of weathered profiles. Hence the distribution of laterite deposits as shown in the results may be explained in terms of climatic influence upon their formation.

8.2.4.3 Phosphate Deposits (PHOS).

There is general agreement amongst workers that the formation of primary PHOS deposits is mainly due to upwelling of cold, oceanic currents (e.g. Gulbrandsen, 1969; Cook, 1976; Parrish et al, 1986). It should be first explained that upwelling simply raises water from below that photic zone (> 100m) to above it (<100m). In the photic zone, light penetrates and permits photosynthesis and phytoplankton remove dissolved P for metabolic use. The concentration of P in surface water is therefore about 10ppb. If off-shore wind blows this water away from the coastline it has to be replaced by water that upwells from depth and this typically has about 70ppb P or more. The solubility of phosphate in sea water increases with decreasing pH and temperature and so it is to be found concentrated in deep, cold sea water. When such waters are forced to upwell by divergence or other means along continental shelves, sea life flourishes and P is taken up. Also in this shallow environment the waters are warmed, resulting in the deposition of phosphate both inorganically and biochemically (Brown, 1974). Such a situation is found today along the southwest coast of southern Africa on the Agulhas Bank (Parker, 1971; Summerhayes, 1972, 1973; Birch, 1973). In areas of upwelling high pelagic biological activity occurs so the shelf and slope sediments are rich in organic carbon and biogenic carbonate. Phosphate-rich microenvironments are created in the sediment interstices by solutions derived from dead phyto- and zooplankton protoplasm. When these solutions then came into contact with a sediment with a high surface area/volume ratio which is susceptible to replacment (e.g. micrite) phosphatization would proceed (Dingle, 1974). Upwelling has been shown to be capable of fixing sufficient P to make a phosphorite (McKelvey et al, 1959). It is fixed in the biomass which sinks on death and undergoes sub-oxic diagenesis by bacteria and the contained P is released as the organic matter is oxidized. Eventually pore water phosphate concentrations rise enough to precipitate francolite directly or to favour replacement of carbonates and so phosphorite is formed.

Cold oceanic currents affect phosphate abundance of oceanic zones over considerable distances at the present time (Cook, 1976). Assuming the present principles of atmospheric circulation such currents presumably occurred in the past, although to a greater or lesser degree depending on the size of polar ice caps and oceanic/continent configuration as determined by palaeogeography. Sheldon (1964) found that the majority of important PHOS examples were deposited within 40° of the palaeoequator. Within this range at present most cold oceanic currents are associated with coastal upwelling (Cook, 1976). Freas and Eckstrom (1968) summarized the conditions conducive to upwelling, amongst them are the points given below which have been selected on the grounds of being latitudinally or climatically influenced. They found that upwelling was most likely to occur;

1) In the trade wind belts at low latitudes between 0° to 25°,

2) Along the north and west coasts in the northern hemisphere and south and west coasts in the southern hemisphere between 0° and 25°,

3) Along the coasts of land masses with dry climates e.g. Peruvian PHOS deposits with the Sechura Desert as hinterland and southwest Africa with the Namib Desert (Cook, 1976),

4) Along coasts in the belt of westerlies approximately between 20° and 40°.

To conclude, the most widely accepted theory on the origin of PHOS deposits has been given in Cook (1976) with regard to the formation of nodular and pelletal PHOS deposits and is summarized below.

1) An influx of nutrient-rich water, generally by upwelling, into a shallow-marine region (maximum water depth 500m) with a slow rate of terrigenous deposition occurs, often in a warm arid climate e.g. at Baja California, Mexico at present with an annual precipitation average of 12cm and a desertic vegetation of shrubs along the coast (D'Anglejan, 1967).

2) This is followed by the development of a prolific biota,

3) Then the formation of anoxic organic-rich bottom sediments occurs and the loss of C, N and H from dead organisms either before, or immediately after, burial,

4) Interstitial waters rich in P form below the sediment/water interface by leaching of phosphate from organic remains as a consequence of low pH and high alkalinity in the sediments,

5) Localized patches of apatite develop by phosphatization,

6) Reworking of sediments may occur.

The importance of climate to the development of PHOS deposits is very clear. An important condition permitting appreciable phosphatization is the rate of supply of detritus (Gulbrandsen, 1969; Dingle, 1974). The supply of detrital material must be low so that drowning of the incipient phosphorite enrichments by rapidly accumulating particulate detritus is prevented. Such conditions are met in arid regions with a low rate of detritus supply due to limited erosion and a lack of significant continental run-off and aeolian dust movement (Manheim et al, 1975). Brongersma-Sanders (1969) however noted that the process of upwelling and the associated presence of cold nearshore waters may be responsible for the aridity of the hinterland and the development of coastal deserts adjacent to PHOS deposits e.g. the Atacama and Namib Deserts.

It appears that a greater degree of phosphatization of sediments occurs in areas of very low phosphate concentration in surface waters with a slow rate of sedimentation, than in areas of very high phosphate concentration characterized by rapid sedimentation (Cook and Mayo, 1977). This emphasizes the main difficulty in postulating an estuarine source for the phosphate in PHOS deposits i.e. that there is a generally high rate of sedimentation in estuaries and in most cases this will produce nothing more than a very slightly phosphatic mud or sand. Hence an estuarine source for PHOS is only conceivable in circumstances where the great bulk of the river sediment is prevented from reaching the estuary (Cook, 1976). This clearly supports the theory of a marine source for the P in PHOS deposits.

The low sediment supply also assists in the development of a prolific biota in the upwelling oceanic waters. The influence of such a biomass upon the deposition of PHOS deposits is very great. One of the primary controls on phosphate solubility may be alkalinity, which is, in part, the result of the abundance of organic matter (Cook, 1976). The biota also act as concentrating mechanisms for phosphate as the prolific phytoplankton population releases phosphate at the sediment/water interface after sinking on death (Howard and Hough, 1979). In addition to supplying phosphate, the breakdown of organic matter should also lower pH, simultaneously promoting calcite undersaturation.

A number of PHOS deposits are associated with evaporites, a classic example being the Phosphoria Formation (McKelvey et al, 1959). Hite (1976) suggested that this is a direct association, with phosphate being precipitated as a result of interaction of P-rich continental brines with cold ocean waters. It has also been suggested that if periods of global aridity have occurred (due to broadening of trade wind belt) then they may have promoted periods of phosphogenesis. However Cook and McElhinny (1979) considered the association between PHOS and evaporites to be purely a plate tectonic one. They claimed that evaporites are the result of two processes; the movement of an area to a low latitude location and the initiation of rifting which may form a salt basin. They proposed that whether phosphate was synchronous with, immediately followed, or occurred much later than evaporite deposition was dependent primarily upon plate movements.

At the other extreme it has also been proposed that there is an association between PHOS deposits and glaciation. Data presented by Sheldon (1980, 1981) and Sheldon and Burnett (1981a and b) show one correlation between periods of glaciation and the distribution of west coast type phosphorites and another correlation between periods of no glaciation and a predominance of equatorial phosphorites. Cook and McElhinny (1979) considered that glacial periods may have been an important influence on phosphate deposition due to an intensification of upwelling caused by a greater temperature differential (Gardner, 1973) during such times. The deep ocean waters may also have had a greater concentration because of their lower temperatures during glacial periods. They also suggested that the onset of glaciation may have initiated oceanic overturn. On the other hand the major periods of phosphogenesis in the Mesozoic and early Cenozoic are totally devoid of any association with glacial episodes (Cook and McElhinny, 1979). Also it has been shown (Burnett, 1977) that Pleistocene phosphate off Peru and southern Chile have radiometric ages corresponding to interglacial periods when oceans were warmer, sea level higher and there was mechanical upgrading of PHOS deposits.

So phosphorites are thought to form in zones of upwelling in warm, arid climates for a number of reasons. Firstly the only efficient ways of removing P from sea water are biological or by adsorption to clays and/or oxides. Phosphorites rarely have evidence of clays or oxides (except phosphatic iron formations) but much evidence of associated organic matter. Secondly deep ocean water (>100m) is essential for PHOS development. If the use of surface water was to be invoked vast volumes would be necessary (perhaps by a factor of 100 times that postulated for deep waters) and there is no way of stripping even 1ppb from it. Thirdly since a biological mechanism for phosphate concentration seems to be essential, the fixation process must occur within the photic zone, the only area other than black smoker vents, where primary production of phosphate occurs (R. Miller, pers. comm.).

Not all upwelling is wind-driven. Some may be topographically-induced (as any sea floor obstruction in the path of a current may cause it to rise), although this may also be augmented by winds (Blanton et al, 1981). However most upwelling is wind-driven, the most effective being the Westerlies and Trades either side of the equator. A concentration of PHOS deposits in these zones would be expected and is observed today. PHOS deposit formation is most abundant in low-latitude locations with a preference for subequatorial (10° - 20°) locations rather than equatorial (0° - 10°) sites (Cook and McElhinny, 1979) although this is not so evident in the results given here (Chapter six, section 6.5.3). Cook and McElhinny (1979) also described a bimodal distribution with modes at 10° to 40°, particularly from major deposits in the Jurassic, Cambrian and, to a lesser degree, the Permian. They explained intermediate latitude PHOS deposits (>40°) as being formed in response to dynamic upwelling as currents were forced over topographic high. The bimodal distribution is not clear in these results, perhaps because the sample size was too small, but the peak around 25° to 40° is clear. This is probably due to the upwelling caused by the Westerlies in these latitude bands.

It is evident that for most PHOS deposits to form a coastal portion of a continent must drift into a low latitude location. However this is no guarantee that strong upwelling will be produced as the continent-ocean figuration is also of great importance. According to Cook and McElhinny (1979) a narrow east-west seaway in an arid climatic zone at low latitude (e.g. the Tethyan seaway) would probably produce PHOS deposits in response to strong dynamic upwelling caused by the strong westerly-directed flow through it. PHOS deposit development may also occur in broad north-south seaways in response to oceanic upwelling on the east side of the ocean.

In conclusion the distribution of PHOS deposits may be explained in terms of being climatically controlled in that they are deposited in response to upwelling, usually related to the Trade and Westerlies wind belts on either side of the equator. On a smaller scale, a warm arid climate is preferred as such conditions promote high productivity, insignficant continental run-off and low sediment supply to the sea.

Conclusion.

It has been shown that the separation and concentration of elements in the surface cycle is due, in part, to weathering and diagenetic processes e.g. the massive differentiation of many primary rock types into residual and solution products in lateritic weathering. Hence the role of chemical weathering in the breakdown of the Earth's crust is of great importance to the development of many mineral deposit types. The concentration of the elements released by such weathering processes into economic accumulations is largely dependent upon the involvement of the biomass in weathering and diagenesis. The influence of microbiological factors lies mainly in controlling the input of anions (e.g. HCO_3^-, HS^-, PO_4^{3-}, organic acid anions) into soil or sediment systems.

The results show that the majority of mineral types included in this study have an apparent latitudinal control upon their formation, with a preference for low- and mid- latitudes. However the range of latitude over which examples occur varies between mineral deposit types. For example the distributions of the LSBM, OOFE, SHBM, SDEX, SSUV, PLDI and PLSN deposits show a peak in the equatorial rainfall belt whereas the SSCU, PLAU, PLOT, PLOX, MNFM and PHOS deposits are apparently suppressed in numbers in this region. However the OOFE, SSUV, PLAU, PLDI, MNFM and PHOS deposits all show peaks in distribution in the warm temperate rainfall belt (about 40-55° north and south of the equator). Generally all deposit groups show a marked suppression in number above 50° latitude and in a number of cases no examples are found in higher latitudes. This suggests that chemical weathering (which decreases in intensity under cooler climatic conditions) does indeed play an important role in determining whether or not mineral deposits can develop.

However climate is only one of a number of possible enhancements that must occur for mineral deposits to form. Amongst these other coincidences are a source for the metals and palaeogeographic configurations (the latter favourably influence atmospheric and oceanic circulation patterns - see Chapter Four, section 4.5.2.2) as well as climatic conditions and weathering characteristics as a consequence of these patterns (see Chapter Eight, section 8.1). Each of these three aspects must be optimum for a deposit to form.

The formation of mineral deposits is therefore reliant upon the coincidence of several factors, one of which is climate. Nonetheless the influence of climate is sufficiently great to be reflected in the palaeo-distributions of deposits. For example some deposit types (e.g. OOFE and SSUV) require a high precipitation rate for their development. It promotes the release and transport of Fe in the first instance and provides the depositional environment in the second. Such a dependence is therefore reflected in the distribution of such deposits, a large number of which occur in the equatorial and temperate rainfall belts.

CHAPTER NINE

TEST CASES

The palaeolatitudinal distributions of some types of volcanogenic mineral deposits were determined in an effort to ascertain whether all mineral deposits show a latitudinal control upon their formation, regardless of origin. The results for the volcanogenic deposits may test the reliability of the data collection and processing as the same methods were used as for the sediment-hosted deposits.

Two types of volcanogenic mineral deposits were chosen as test subjects for this research as there is no obvious reason why their formation should be latitudinally/climatically controlled; porphyry copper (PORCU) and epithermal gold (EPIAU) deposits. A considerable amount of literature is also available for these deposit types so data collection was easier. Finally many of the examples are dated as Mesozoic and Cenozoic so the palaeomagnetic data from which the palaeolatitudes were determined should be more reliable than if the deposits were Palaeozoic in age.

9.1 Methods.

The same procedure as described in Chapter Five (Methods) for sediment-hosted mineral deposits was used to determine the palaeolatitudinal distribution of the PORCU and EPIAU mineral deposit examples.

9.2. Results.

Porphyry copper deposits show a range of distribution from 0° to 85° north and south of the palaeoequator, the majority plotting in the northern hemisphere as shown by the mean and median (Table 9.1). This is a reflection of a bias in the collection of data from a limited number of areas i.e. mainly USA, Canada and to a lesser extent Queensland, Australia. The majority of deposits occur in the mid to high latitudes i.e. 30° to 80° as shown in Figure 9.1. The plot of palaeolatitudes from the minimum ages of the deposits shows 65% in this latitude range with 66% of examples in the maximum-age derived plot. In both minimum and maximum histograms there is also a peak in the equatorial humid zone i.e.5°N to 10°S (24% of examples in both diagrams). However there is an apparent lack of deposits in the hot, arid climatic zones between 10° and 30° north and south of the palaeoequator.

Table 9.1: Mean, Median and Standard Deviation for PORCU and EPIAU Deposits.

MIN DEPOSIT TYPE		NO. OF EXAMPLES	MEAN (degrees) 90°N-90°S	MEDIAN (degrees)	STANDARD DEVIATION	COEFFICIENT OF VARIATION
PORCU	MIN	168	21.6	37.8	40.8	36.6
	MAX	168	18.0	36.6	43.2	40.0
EPIAU	MIN	59	18.9	28.7	27.0	24.8
	MAX	59	21.4	33.5	28.6	25.7

(All latitudes are in the northern hemisphere).

The palaeolatitudinal range of EPIAU deposits is 0° to 65° with the majority in mid latitudes i.e. 30° to 60° (minimum-age derived plot = 58%; maximum-age derived plot = 58%). The majority of deposits plot in the northern hemisphere as shown by the mean and median in Table 9.1 which again reflects the limited number of areas from which the examples were collected. Most

Figure 9.1: Porphyry Copper Deposits

Minimum Age Plot

a -	Malaysia & Phillipines	(50)
	Queensland, Australia	(200, 300)
b -	Alaska, USA	(350)
	Queensland, Australia	(250)
	Peru	(50)
c -	Peru	(0)
	Sulawesi, Indonesia	(50)
d -	New Brunswick & Quebec, Can	(350)
	Peru	(50)
e -	New Brunswick, Canada	(350)
	Newfoundland, Canada	(350)
	Chile & Peru	(50)
f -	Argentina	(0)
	Chile	(50)
	Kazakhstan, USSR	(300)
g -	Idaho, USA	(200)
	Argentina	(0)
	Chile	(0)
	Papua New Guinea	(50)
	Kazakhstan, USSR	(300)
h -	Arizona, USA	(50, 130)
	B.C., Canada	(200)
	Mexico	(50)
	New Mexico & Texas, USA	(50)
i -	Alaska, USA	(200)
	Arizona, USA	(50)
	B.C., Canada	(200)
j -	B.C., Canada	(200)
	Maine, USA	(450)
	Nevada, USA	(100)
	Queensland, Australia	(100)
	Utah, USA	(50)
	Washington, USA	(0)
k -	Alaska, USA	(0)
	Montana, USA	(50)
	Queensland, Australia	(400)
l -	Alaska, USA	(0)
	B.C., Canada	(50, 130)
	Vancouver I., Canada	(50)
	Washington, USA	(50)
m -	Alaska, USA	(0)
	B.C., Canada	(50, 100)
	Queensland, Australia	(130)
o -	Alaska, USA	(0)
	Queensland, Australia	(130)
	Vancouver I., Canada	(130)
p -	Alaska, USA	(50, 100)
	Yukon, Canada	(50)
q -	Alaska, USA	(50, 100)
		(100)

Maximum Age Plot

A -	Malaysia & Phillipines	(50)
	Queensland, Australia	(200, 300)
B -	Alaska, USA	(350)
	Queensland, Australia	(250)
	Peru	(50)
C -	Peru	(0)
	Sulawesi, Indonesia	(50)
D -	Quebec, Canada	(350)
	Peru	(50)
E -	New Brunswick, Canada	(350)
	Chile & Peru	(50)
F -	Argentina	(0)
	Chile	(50)
	Kazakhstan, USSR	(300)
G -	Idaho, USA	(200)
	Argentina & Chile	(0)
	Papua New Guinea	(50)
	Kazakhstan, USSR	(300)
H -	Alaska, USA	(250)
	Arizona, USA	(50, 130)
	B.C., Canada	(200)
	Mexico,	(50)
	New Brunswick, Canada	(400)
	Newfoundland, Canada	(400)
	New Mexico, USA	(50)
	Texas, USA	(50)
I -	Arizona, USA	(50)
	B.C., Canada	(200)
J -	B.C., Canada	(200)
	Maine, USA	(450)
	Nevada, USA	(100)
	Utah, USA	(50)
	Washington, USA	(0)
K -	Montana, USA	(50)
	Queensland, Australia	(400)
	Washington, USA	(50)
L -	Alaska, USA	(0)
	B.C., Canada	(50, 130)
	Vancouver I., Canada	(50)
	Washington, USA	(50)
M -	B.C., Canada	(50, 100)
	Queensland, Australia	(130)
N -	Alaska, USA	(50, 130)
	Queensland, Australia	(130)
	Vancouver I., Canada	(130)
O -	Alaska, USA	(50, 100, 130)
	Yukon, Canada	(50)
P -	Alaska, USA	(50,100)
Q -	Alaska, USA	(100)

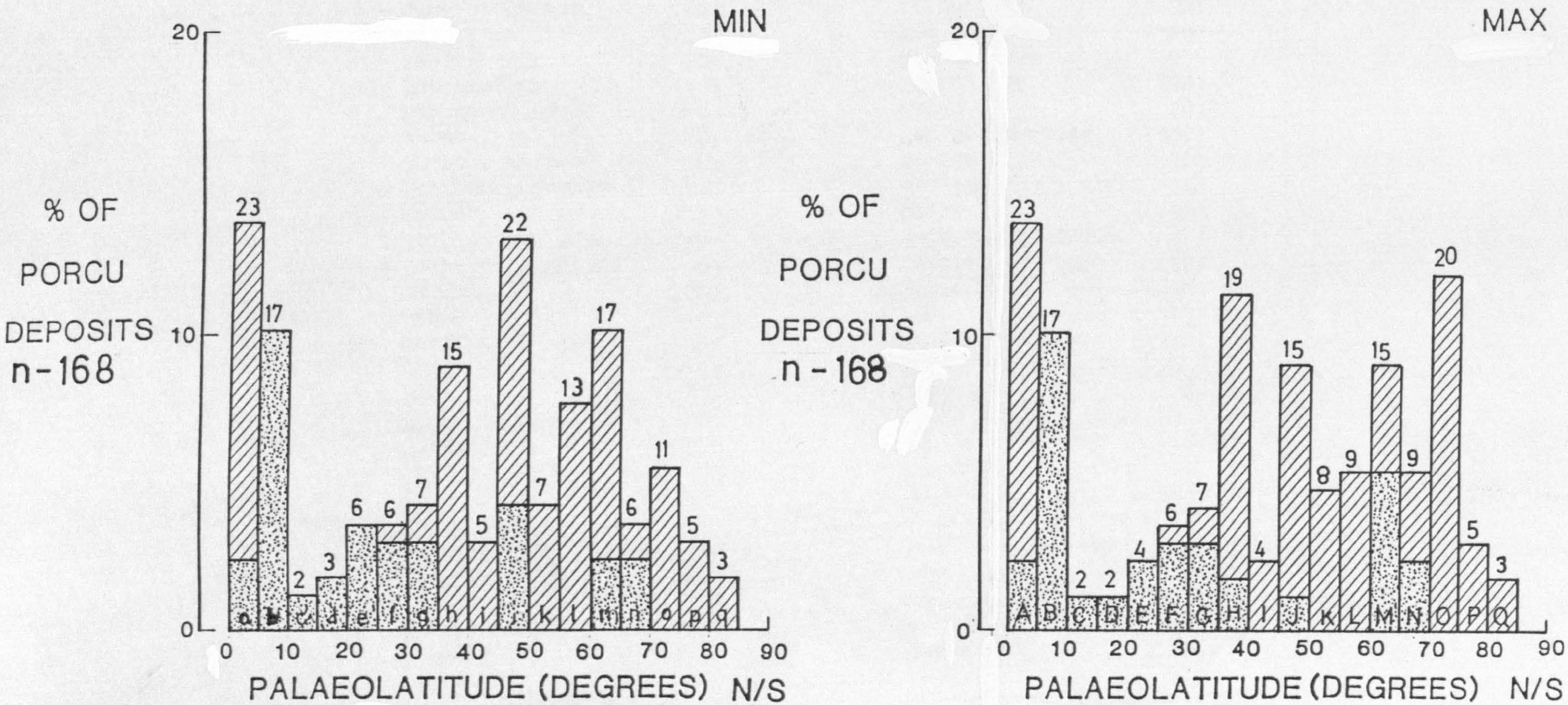


Figure 9.I. Palaeolatitude vs. Frequency for PORCU deposits from Tarling reconstructions.

For figure legend see Figure 6.2(i).

Figure 9.2: Epithermal Gold Deposits

Minimum Age Plot

a -	Queensland	(250)
	New South Wales	(250)
b -	Peru & El Salvador	(0)
c -	New South Wales	(350)
	Mexico, Fiji, Luzon	(0)
	Phillipines	(0)
	Peru	(130)
d -	Taiwan	(0)
	Mexico	(50)
	Chile	(130)
e -	Mexico	(50)
	Chile	(130)
f -	California, USA	(0)
	Japan	(0)
	Chile	(130)
g -	Colorado & Nevada	(0)
	New Zealand & Spain	(0)
	Chile	(130)
h -	British Columbia	(200)
	Idaho & Nevada, USA	(0)
	Japan	(0)
	New South Wales	(400)
i -	Montana, USA	(0)
	Nevada, USA	(0, 50)
	Argentina	(130)
j -	Cinola, Canada	(0)

Maximum Age Plot

A -	Queensland	(300)
B -	Queensland	(250)
	New South Wales	(250)
C -	Peru & El Salvador	(0)
D -	Mexico, Fiji, Luzon	(0)
	Phillipines	(0)
	Peru	(130)
E -	Taiwan	(0)
	Mexico	(50)
	Chile	(130)
F -	Mexico	(50)
	Chile	(130)
G -	British Columbia	(250)
	Japan	(0)
	Chile	(130)
H -	California, USA	(50)
	Colorado & Nevada	(0)
	Chile	(130)
	New Zealand & Spain	(0)
I -	Nevada, USA	(0)
	New South Wales	(400)
	Japan	(0)
J -	Idaho, USA	(50)
	Nevada, USA	(50, 100)
	New South Wales	(450)
	Argentina	(130)
K -	Montana, USA	(50)
L -	Cinola, Canada	(50)

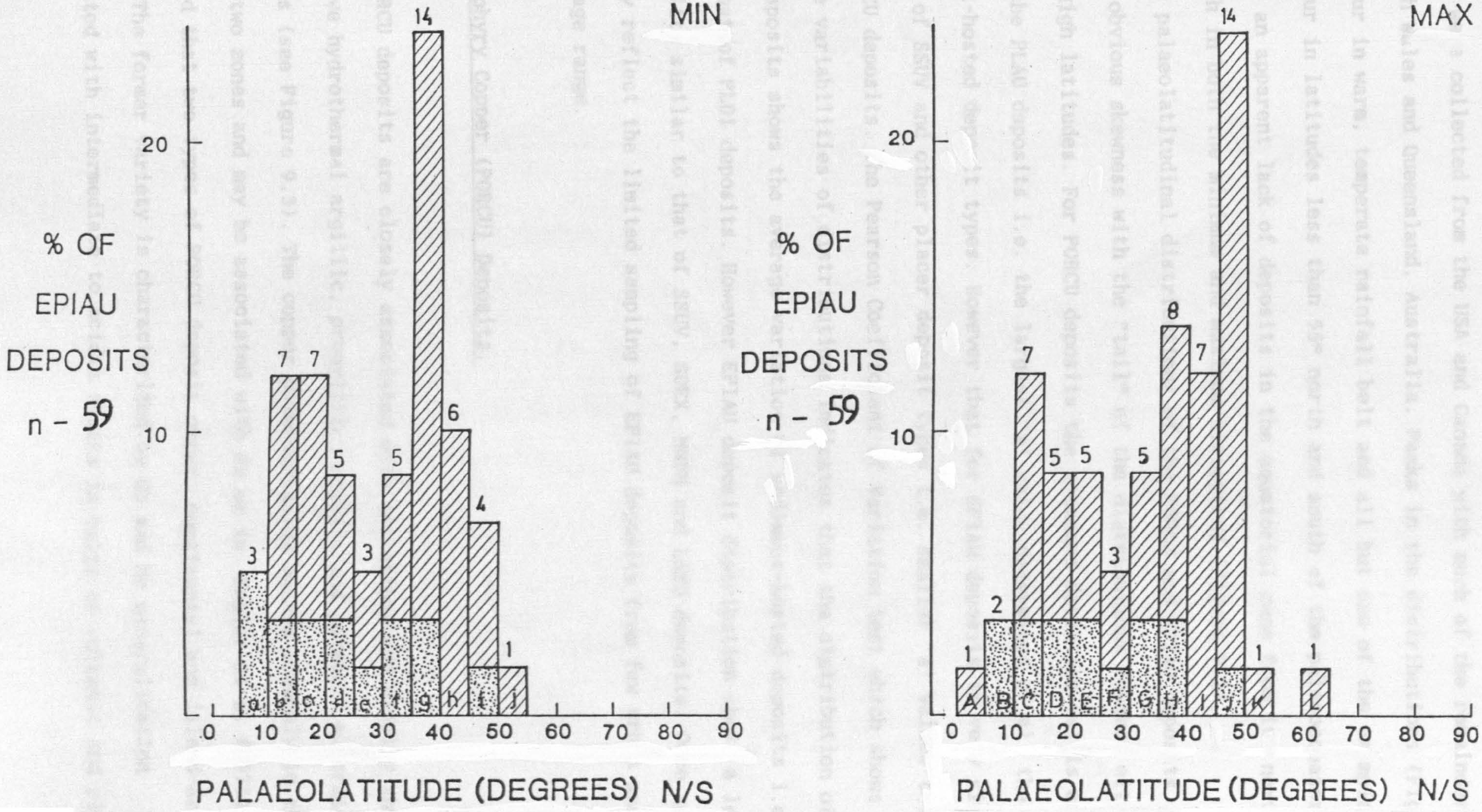


Figure 9.2. Palaeolatitude vs. Frequency for EPIAU deposits from Tarling reconstructions.
For figure legend see Figure 6.2(i).

examples were collected from the USA and Canada with much of the remainder from New South Wales and Queensland, Australia. Peaks in the distributions (Figure 9.2) occur in warm, temperate rainfall belt and all but one of the examples used occur in latitudes less than 55° north and south of the palaeoequator. There is an apparent lack of deposits in the equatorial zone from 10° north to 10° south in both the minimum and maximum histograms (Figure 9.2).

The palaeolatitudinal distributions of the PORCU and EPIAU deposits do not show an obvious skewness with the "tail" of the distributions towards either low or high latitudes. For PORCU deposits the standard deviation (s) is as high as for the PLAU deposits i.e. the largest 's' value determined of all the sediment-hosted deposit types. However that for EPIAU deposits is very similar to that of SSUV and other placer deposit types i.e. smaller 's' values than for the PORCU deposits. The Pearson Coefficient of Variation test which shows the relative variabilities of distributions indicates that the distribution of PORCU deposits shows the average variation of sediment-hosted deposits i.e. about that of PLDI deposits. However EPIAU deposit distribution shows a lower variability similar to that of SSUV, SDEX, MNFM and LATO deposits. Once again this may reflect the limited sampling of EPIAU deposits from few areas and a narrow age range.

9.3 Porphyry Copper (PORCU) Deposits.

PORCU deposits are closely associated with intrusions and usually show extensive hydrothermal argillic, propylitic, phyllic and potassic alteration aureoles (see Figure 9.3). The copper mineralisation occurs commonly in the latter two zones and may be associated with Mo or Au. Kesler et al (1975) proposed that two types of PORCU deposit occur; continental and island-arc types. The former variety is characterized by Cu and Mo mineralisation associated with intermediate to acidic stocks in belts of volcanic and plutonic

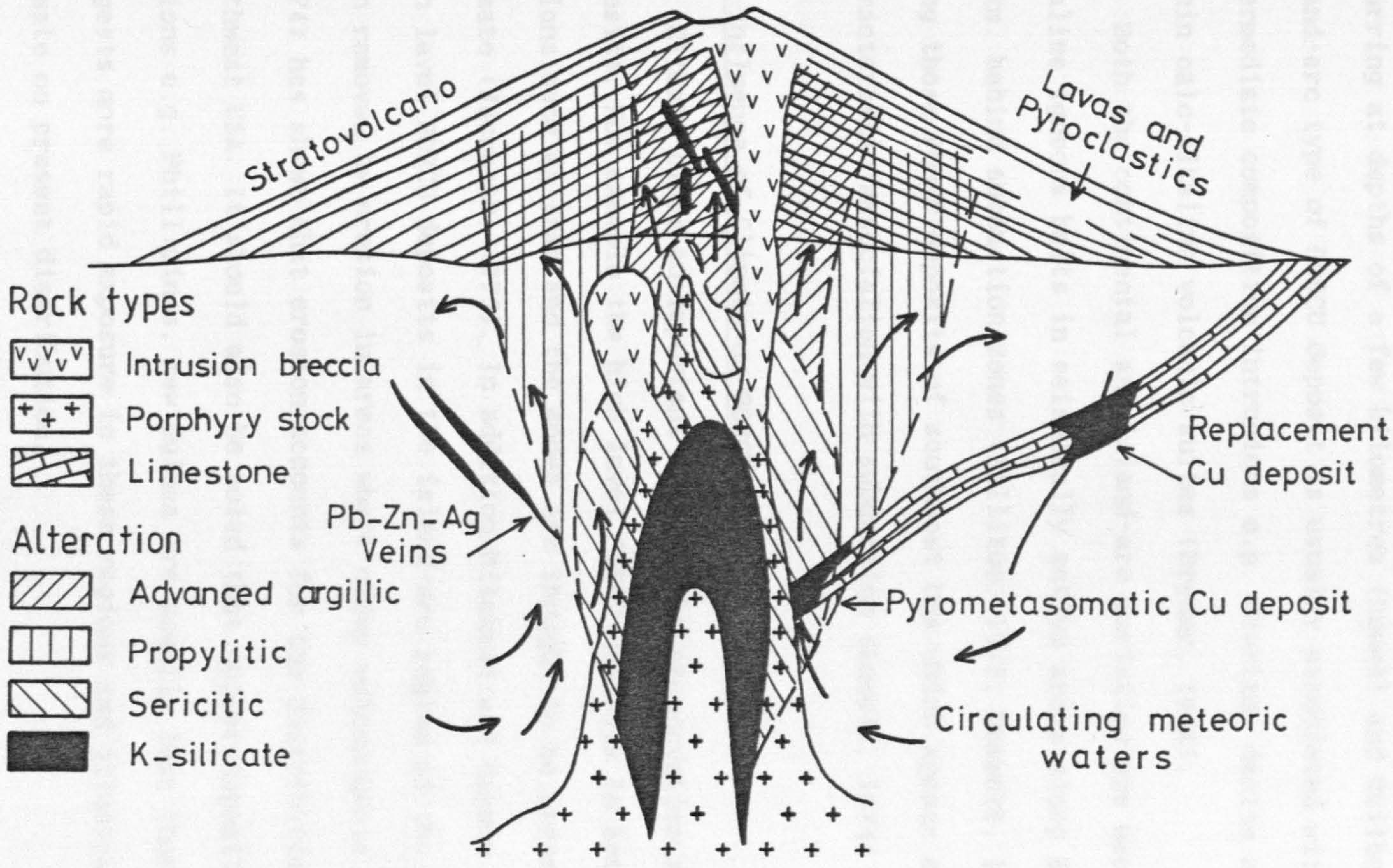


Figure 9.3. Idealized section through a porphyry copper complex. Note replacement and vein deposits and zones of alteration. (After Jacobsen, 1975, Figure 3).

rocks (Stringham, 1966). It is thought that the emplacements of the intrusive hosts, copper mineralisation and alteration are almost contemporaneous occurring at depths of a few kilometres (Lowell and Guilbert, 1970). The island-arc type of PORCU deposit is usually associated with subvolcanic, intermediate composition intrusions e.g. diorite, dacite and andesite porphyry within calc-alkaline volcanic suites (Bryner, 1968).

Both the continental and island-arc varieties are usually found in calc-alkaline igneous belts in seismically active areas along Benioff zones, some 200km. behind subduction zones (Sillitoe, 1972; Dummett, 1978). The exceptions being those PORCU deposits of southwest USA which appear not to show this characteristic association with subduction (Lowell, 1974; Noble, 1974).

The Influence of Climate on PORCU Distribution.

Climate plays an important role in the preservation and exposure of PORCU deposits. For example the high level PORCU deposits in active mountain forming regions such as Iran and the Andes are thought to be preserved by the arid climate (Jacobsen, 1975). In addition Mitchell and Garson (1972) suggested that high level PORCU deposits in the island-arc region of the west Pacific have been removed by erosion in areas where older volcanogenic deposits occur. Noble (1974) has shown that erosion accounts for the distribution of PORCUs in the southwest USA. It should also be noted that copper deposits in high rainfall regions e.g. Phillipines, New Guinea are usually less than 16 m.y. old which suggests more rapid exposure in these regions and illustrates that effect of climate on present distribution.

The dominant influence on the formation of PORCU deposits is generally thought to be the position of presently active and palaeo- Benioff zones. But the distribution of PORCUs shown here (Figure 9.1) suggests that they did not form in the arid latitudinal belts. However the lack of examples in these areas is probably due to a corresponding lack of land mass in these zones at the time of PORCU formation. It must be noted that some of the largest PORCU deposits known are located in particularly arid regions e.g. the Chuquicamata and El Teniente deposits of Chile. Hence the results given here suggest that climate did not influence PORCU formation.

9.4 Epithermal Gold (EPIAU) Deposits.

Two main types of epithermal volcanic-hosted precious metal deposits which form at low to moderate temperatures in near surface environments can be distinguished; the acid-sulphate type (e.g. Goldfield, Nevada) and the adularia-sericite type (e.g. Round Mountain, Nevada). The latter are considerably more abundant than the former (Heald et al, 1987). Detailed descriptions of the characteristics of both types can be found in Hayba et al (1985) and Heald et al (1987). Only a brief review is given here.

A third type of EPIAU deposit characterized by quartz-fluorite-carbonate-adularia-roscoelite alteration with Au commonly found as tellurides has been described (Bonham and Giles, 1983). These major telluride deposits (e.g. Cripple Creek, Colorado) are genetically related to alkalic igneous rocks, are low in sulphur and their unique mineral assemblage suggests that they constitute a distinct class of deposits. Hence examples of this type have not been included in this study.

The Acid-Sulphate-Type EPIAU Deposits.

These deposits are also referred to as high sulphur types (Bonham Jr., 1986). They are found associated with the margins of calderas, those in ring-fracture volcanic domes in particular (e.g. Summitville and Goldfield, Nevada) and this appears to be a genetic factor (Heald et al, 1987). The primary host rock is almost exclusively rhyodacite, commonly porphyritic (Ashley, 1982) and the ages of host and ore are very similar. Acid-sulphate type EPIAU deposits are characterized by argillic alteration and the mineralising fluids are thought to have been dominantly meteoric, possibly with a significant magmatic component, at temperatures of 200° to 300°C (Hayba et al, 1985). The salinity of these fluids is of greater variability than the mineralising fluids of the adularia-sericite type deposits, ranging from 1 to 24 wt.% NaCl e.g. 7 to 21 wt.% at Summitville and 5 to 18 wt.% NaCl at Goldfield (Bruha and Noble, 1983).

Acid-sulphate type EPIAU deposits form in the upper core of volcanic domes which are flooded by meteoric waters as they cool (see Figure 9.4A). The mineralising fluids are thought to have evolved from the interaction of meteoric waters with magmatic volatiles derived from the source magma of the dome. The less common occurrence of these deposits rather than the adularia-sericite type EPIAU deposits is probably related to genetic environment. They appear to require a physical proximity to the heat source, a restriction of the host rock composition to rhyodacite and a certain composition of magma source i.e. sufficiently oxidized to produce SO_2 as the magma degasses (Heald et al, 1987).

<u>The Adularia-Sericite-Type EPIAU Deposits.</u>

These deposits are also referred to as low sulphur types (Bonham Jr., 1986). They commonly occur along the margins of calderas (Hayba et al, 1985) and are hosted by silicic to intermediate volcanics i.e. andesitic to rhyolitic composition. These deposits are characterized by sericitic to argillic alteration and there are distinct differences in ages of the host rock and the ore i.e. >1 m.y. The mineralising fluids are thought to have been dominantly meteoric in nature, between 200° and 300°C (Hayba 1983) with variable salinity e.g. Eureka, Colorado 0 to 3.6 wt.% NaCl; Creede, Colorado 5 to 12 wt.% NaCl (Woods et al, 1982).

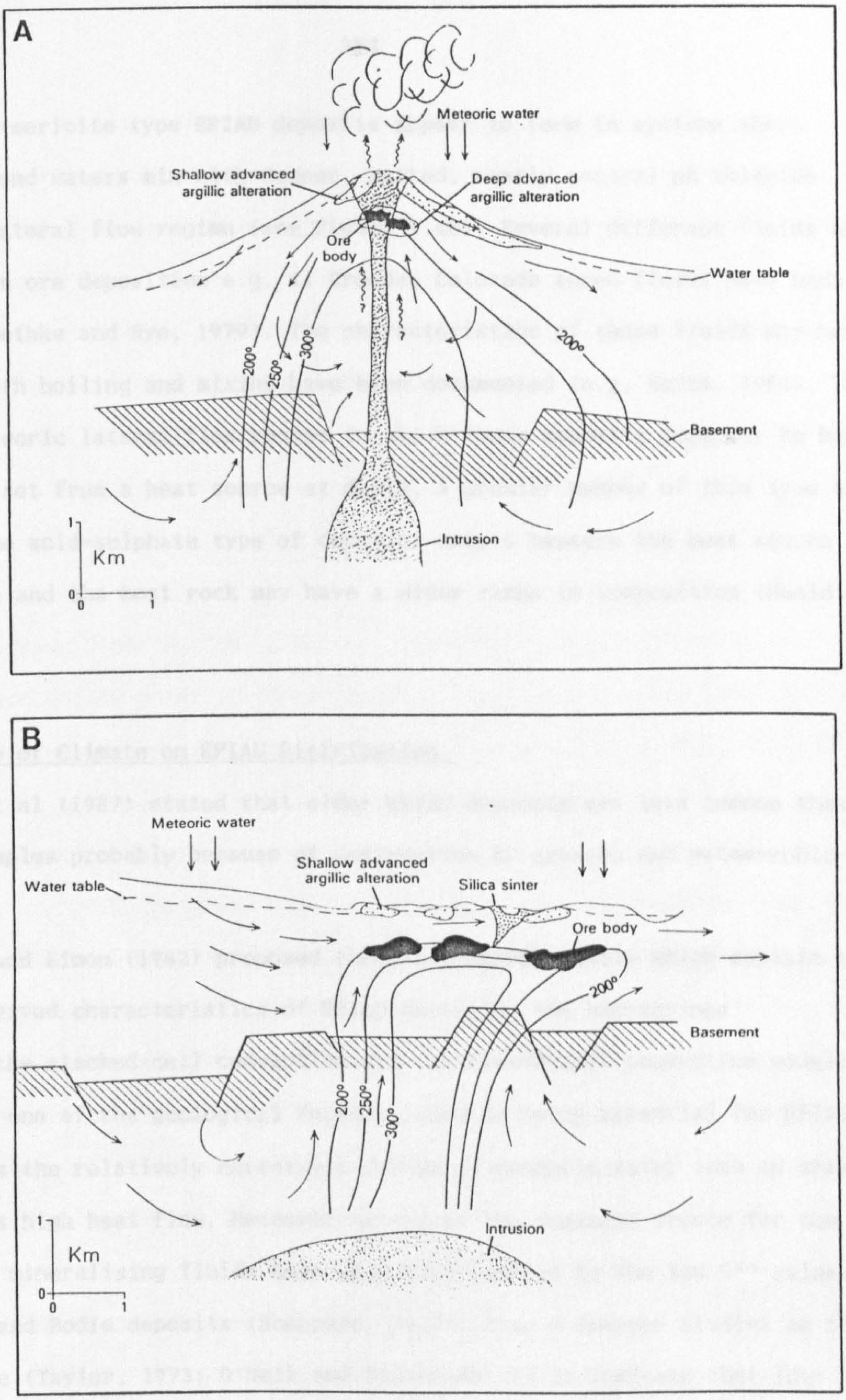


Figure 9.4. Idealized cross-sections of two types of EPIAU deposit genesis in a geothermal system. A) Acid-sulphate Type B) Adularia-sericite Type. (Modified after Henley and Ellis, 1983; Heald et al, 1987, Figure 4).

Adularia-sericite type EPIAU deposits appear to form in systems where surficial ground waters mix with deeper, heated, nearly neutral pH chloride brines in a lateral flow regime (see Figure 9.4B). Several different fluids may be involved in ore deposition e.g. at Creede, Colorado three fluids have been identified (Bethke and Rye, 1979). The characteristics of these fluids may have changed as both boiling and mixing have been documented (e.g. Hayba, 1984). The extensive meteoric lateral flow regime in which these deposits form may be high above or off-set from a heat source at depth. A greater number of this type as opposed to the acid-sulphate type of deposits exists because the heat source can be remote and the host rock may have a wider range in composition (Heald et al, 1987).

The Influence of Climate on EPIAU Distribution.

Heald et al (1987) stated that older EPIAU deposits are less common than Tertiary examples probably because of destruction by erosion and metamorphic overprinting.

Berger and Eimon (1982) proposed three end-member models which explain the range of observed characteristics of EPIAU deposits; the hot-springs deposition, the stacked-cell convection and the closed-cell convection models. In each case one of the geological factors cited as being essential for EPIAU formation was the relatively unrestricted flow of meteoric water into an area of continuous high heat flow. Meteoric waters as the dominant source for the hydrothermal mineralising fluids have also been implied by the low O^{18} values for Tonopah and Bodie deposits (Sheppard, 1977). Also O-isotope studies on the Comstock Lode (Taylor, 1973; O'Neil and Silberman, 1974) indicate that low-salinity meteoric solutions played a dominant role in the transportation and deposition of minerals. Giles and Nelson (1982) stated that economic concentrations of Au require initial high gold solubility, unrestricted recharge of meteoric water into a region of steady, high heat flow. They also considered that among primary processes of ore deposition is local ground water mixing. The importance of meteoric waters in formation may be reflected in the palaeolatitudinal distribution of EPIAU deposits i.e. a large proportion of examples occurs in warm temperate zones. However there are only a few examples in equatorial latitudes and a number of deposits occur in the arid latitude belt. So the distribution of EPIAUs given here does not suggest a strong climatic influence on their formation.

Conclusion.

The worldwide distribution of EPIAUs is similar to that of PORCUs and in a number of cases EPIAUs occur in close proximity to PORCUs e.g. Ely and Bingham, western USA, Taiwan and Mexico. However in other regions there is an absence of PORCU deposits e.g. Sumatra. Hutchinson (1983) considered this discrepancy to be related to erosion - deeper erosion than that in the vicinity of some EPIAUs is necessary to expose the PORCU deposits beneath. An economic factor should also be considered here as it influences the distribution of these deposits. In Sumatra a PORCU deposit with no gold is unlikely to be economic. Hence the apparent paucity of PORCUs in this region is due to the fact that such gold-poor deposits have not been evaluated and mined. Thompson et al (1986) also highlighted the relationship between PORCU and EPIAU deposits. They concluded that formation of epithermal gold-silver mineralisation e.g. the Temora deposit, within the high sulphur environments was related to the upper levels of PORCU systems in the Lachlan fold belt during the Middle Silurian. Heald et al (1987) considered that certain characteristics of acid-sulphate type deposits suggest that these deposits are more likely to be associated with PORCU systems than the adularia-sericite type. Such characteristics are the alteration and mineral assemblages, high copper contents and associated porphyritic rocks. However Henley and Ellis (1983) inferred that EPIAU ores in a calc-alkaline stratovolcano are more likely to be associated with porphyry ores at depth than are ores in silicic volcanic terrain.

If there is a genetic relationship between the two types then the differences in their palaeolatitudinal distributions as shown here need some explanation. There is only one EPIAU deposit above 55°N (Cinola, Canada; Figure 9.2) whereas PORCU deposits occur up to 85°N (see Figure 9.1). However many of the PORCUs are older than EPIAU examples so EPIAU deposits may have been present when the continents were at lower latitudes and have since been eroded from the geological record.

To conclude a comparison of the palaeolatitudinal distributions of PORCU and EPIAU deposits does not suggest that the two deposit types are genetically related. However the differences in their distributions may be explained as above. It is clear from a study of the models of PORCU genesis that one of the main controls upon their distribution may be the position of the palaeo- or present Benioff zone. The apparent lack of PORCUs in arid climatic zones may reflect the distribution of land mass during their development as the supergene process which occurs in arid regions aids PORCU development producing some of the richest and most economic deposits e.g. Chuquicamata deposit, Chile. As the distribution of neither PORCU nor EPIAU deposit types illustrates a strong climatic control their usefulness as "Test Cases" for this research is confirmed.

CHAPTER TEN

CONCLUSIONS AND COMMENTS

A number of basic assumptions concerning the Earth and its dynamics were made at the onset of the research and any subsequent conclusions were drawn in the knowledge of the caveats that follow. Firstly it was argued that the time-averaged magnetic field of the Earth had acted as a geocentric axial dipole throughout the Phanerozoic. This model is probably invalid during polarity transitions as the dipole component diminishes during such times. However these intervals have been assumed to have been sufficiently rare and short-lived as not to affect the results and conclusions. Other assumptions involved the constancy of the Earth's obliquity and its speed of rotation. It has been assumed that differences in the Earth's rotational speed between the present and the Ordovician/Silurian were not sufficient to require the abandonment of the six-cell zonal model of circulation as described in Chapter Four, section 4.4. Hence neither aspect has affected atmospheric and oceanic circulation patterns (and thence climatic regimes) too drastically despite the fact that they must have altered slightly during the Phanerozoic.

The relationship between mineral deposit types and climatic conditions that prevailed during the period of their formation, as inferred from palaeolatitude, is also based upon a number of assumptions. Firstly any possible differences in the equator-to-pole temperature gradient since the Ordovician/Silurian would not have been sufficient to invalidate the use of principles of modern atmospheric circulation. Secondly the heat budget of the Earth and its distribution through circulation patterns have been assumed to be approximately constant throughout the past 550 m.y. However it does not necessarily follow that climate zones (as confined to particular latitude bands) have also remained constant during that period. Nonetheless it still seems probable that climatic latitudinal zoning existed although their precise latitudinal extent may have varied.

It is concluded that the Uniformitarian approach to the problem of palaeoclimatology is valid in that some authors have successfully predicted regions of upwelling (Parrish and Curtis, 1982); the distribution of evaporites and coals (Parrish et al, 1982b) and petroleum source rocks (Barron, 1985) using this method. These successes obviously lend credence to the conclusion that uniformitarianism with regard to palaeoclimatology is, in part, acceptable. The use of geological data however is limited as they are frequently insufficient to interpret climatic conditions fully. This shortage may result from poor preservation, alteration or dissolution of materials or it may arise from the fact that many climatic parameters (e.g. atmospheric pressure) leave no discernible mark on the rocks. Once the assumptions concerning the constancy of the Earth's atmospheric circulation model throughout the Phanerozoic have been accepted the relationship between palaeolatitude and palaeoclimatology has also been established.

Within this study the site palaeolatitude is of major concern as it largely dominates the climate, sedimentation and erosion. It must therefore affect the distribution of sediment-hosted mineral deposits and any errors in the palaeomagnetic studies will result in errors for the palaeolatitudinal determinations. The palaeomagnetic data selected by Tarling were subjected to the procedures outlined in Chapter Two with the knowledge of the potential errors inherent in these studies as described (Chapter Seven). It is understood that the reliability of the palaeogeographic data decreases with increasing age i.e. the Palaeozoic reconstructions are not as reliable as those for the Mesozoic and Cenozoic. However in light of the other assumptions that have had to be made, it was deemed that the level of reliability was sufficiently high to base research on palaeolatitudes determined from the palaeogeographic reconstructions. The two sets of palaeolatitudes used in this study (i.e. those derived from Tarling and BP palaeogeographies) were generally in close agreement which suggests that a certain level of reliability in palaeomagnetic data selection, processing and interpretation has been achieved.

The results given in Chapter Six show there is a clear palaeolatitudinal control upon many types of sediment-hosted mineral deposit. However the palaeolatitudinal range of the various deposit types varies considerably. Those deposit groups which show a concentration (at least in part) in the equatorial rainfall belt include LSBM, OOFE, SHBM, SDEX, SSUV, PLDI and PLSN deposits. Conversely the SSCU, PLAU, PLOT, PLOX, MNFM and PHOS groups are suppressed in numbers in this region. The warm arid climatic zone ranging from 15° to 40° shows a concentration of SSCU, PLOT, MNFM and PHOS deposits but very few examples of SSPB, SHBM, SSUV, PLSN and PLOX groups. Finally OOFE, SSUV, PLAU, PLDI, MNFM and PHOS deposits all show peaks in their distribution in the warm, temperate rainfall belt (about 40° to 55° north and south of the equator).

Those deposits which have a preference for the low latitude zone in particular are the stratiform, sediment-hosted (Cu, Pb, Zn) types representative of the LSBM, SSCU, SSPB, SHBM and SDEX groups. The SSUV group shows a predeliction for the lower-mid latitudes in addition to the equatorial belt.

One trend common to all the mineral deposit types examined here is a limit to the lower and mid latitudes i.e. very few deposits occur in latitudes higher than 60° north and south of the equator. This suggests that the higher temperatures, greater degree of chemical weathering and availability of organic matter in lower as opposed to higher latitudes favour mineral accumulation and deposit development.

It is evident that the rate of mineral formation within certain latitudes is not directly related to the percentage of continental mass in the same latitude zone (see Tables 6.3 and 6.4). However a comparison of these relative distributions in the northern and southern hemispheres for each geological period show that they are very similar. Therefore the distributions of mineral deposits and continental mass within each hemisphere are uneven. These findings further substantiate the case for a palaeolatitudinal control upon the formation of some mineral deposit types.

Any conclusions made concerning the following mineral deposit groups must be notably tentative as the sample number is small (i.e. $n < 30$; OOFE, SSPB, SSCU, SHBM, PLSN, PLOX, PLOT and MNFM). However the general trends shown by their palaeolatitudinal distribution may still be of interest.

The separation of primary and secondary mineral deposits is very difficult. A primary sediment-hosted deposit is the product of sedimentary processes at, near, or above the sediment-water interface. The deposits are themselves sediments so they can show all sedimentary features. The main controls on this type of deposit are the physical and chemical conditions of sedimentation - hence palaeoclimatology and palaeogeography may be of particular interest. Secondary sediment-hosted mineral deposits are those which are not obviously a sediment nor have they a clear igneous source for the mineralisation. They appear to be more closely controlled by geotectonic processes, particularly those which affect the temperature and routes of migrating fluids. So geotectonics and environment may be particularly important constraints on secondary ore formation. Hence the development of the two main varieties of mineral deposits (i.e. syngenetic/primary and epigenetic/secondary) may be influenced by climate in different ways. With regard to the former, syn-ore climate appears to be of greatest importance, whereas both syn-ore and pre-ore climates would be influential in the deposition of epigenetic deposits. In the latter case, the products of climate in sediments may control later processes (e.g. permeability and zones of reduction produced by the presence of organic matter).

In the introduction it was proposed that the classification of mineral deposits might be elucidated by an examination of the different palaeolatitudinal ranges of the distributions of different deposit groups - if such a discrepancy were observed. It can now be concluded that the classification, as given in Chapter Three, was fundamentally important to the discussion as different groups do show different palaeolatitudinal ranges. However the more detailed aspects of the classification which were outlined (such as the various types of PHOS deposits) were not important as these individual groups could not be segregated within the results. With regard to the sediment-hosted stratiform (Cu, Pb, Zn) deposits it was prudent to classify them as separate deposit types as supported by the results. They do all show a low latitude control but Pb-Zn rich deposits are concentrated in rainfall belts (e.g. LSBM, SHBM, SDEX) whilst Cu-rich types (e.g. SSCU) show a concentration in the more arid zones. These results suggest that a genetic constraint exists upon the distribution of these deposit types. It may be that the Pb-Zn types preferentially occur where microbiological activity is at an optimum and the clay mineral content of the host rocks is high. In contrast the development of Cu-rich deposit may be more reliant upon the presence of oxides and evaporites. There are obvious exploration implications of these conclusions. For example SSCU deposits would be more likely to be found in lithologies indicative of warm arid climatic conditions than those deposited in a very wet environment. Hence it would be prudent to determine the palaeolatitude of the area which is to be explored at the time when the mineralisation occurred in order to ascertain the likelihood of finding a deposit.

In Chapter Eight an attempt was made to show the potential importance of climate on mineral deposit development. Climate affects the degree of weathering, the amount of available organic matter for reduction, evaporites for a sulphur source and dictates the volume and chemistry of potential metal-bearing transporting fluids. However climate is only one of many possible enhancements (i.e. initial metal source, metal enrichment, suitable depositional environment) that must occur in order to produce the conditions necessary for ore formation. Hence if the initial solution was not sufficiently metal-rich or the depositional environment was not sufficiently rich in organic matter, then no metal accumulations could occur.

In particular the benefits of the prevailing climatic conditions in low to mid latitudes are of importance to ore formation. Arid zones are characterized by the occurrence of evaporites which may provide chloride brines for transport of metals by complexing and a sulphur source for reduction to sulphide. They also have a low clastic input to depositional sites which precludes dilution of the metal supply. The rainfall belts of low to mid latitudes assist the development of other deposit types i.e. those which require channel sandstones as hosts for mineralisation, a large volume of ground water for initial release of metals and their transport to the depositional site and a large amount of organic debris. In both zones chemical weathering is dominant over physical weathering which aids the release of metals from their source rocks.

One feature of some mineral deposits is a close association with so-called climate-sensitive lithologies. However interpretation of this association must be made with some caution. Recent red beds are known to form under humid tropical conditions in tectonically unstable regions in addition to arid conditions. Also modern evaporites are thought to be primarily found in low-latitude, warm climatic zones with evaporation in excess of precipitation. However aridity is not necessarily confined to high temperature regions, it may also occur under temperate or cool conditions. The effect of local palaeogeography in the production of an unexpected distribution of such lithologies has been described (Chapter Four, section 4.5.2.2). From these examples it is apparent that the presence of climate-sensitive lithologies in association with mineral deposits does not conclusively indicate the climatic conditions at the time of formation. However the presence of red beds or evaporites can confirm conditions inferred by the palaeolatitudes.

In Chapter Four (section 4.3) it was mentioned that the latitudinal extent of rainfall belts may have varied due to the presence or absence of ice sheets on the globe. However this phenomenon is not reflected in the results by a reduction in the latitudinal extent of mineral deposits towards the equator. An examination of the distribution of five deposit types which show a marked palaeolatitudinal control (Figures 6.16 a and b) only reveals a reduction in latitudinal extent of examples during the Permo-Triassic period. This may reflect the influence of the supercontinent Pangaea upon atmospheric and oceanic circulation and hence palaeoclimate (see Chapter Eight, section

8.2.2.3). However such a distribution is the opposite to that which would be expected i.e. a broadening of the latitude range of warm climates. Gulf streams would have carried warm waters to high latitudes on the Tethyan margin whilst currents on the western side of the continent would have produced the opposite effect. Unfortunately no evidence of such extreme east-west climatic asymmetry, which may be produced by such a huge north-south land barrier, can be observed in Figures 6.1 a and b or Figures 6.16 a and b.

The palaeolatitudinal distribution of two volcanic-hosted mineral deposit types (i.e. PORCU and EPIAU) were determined to ascertain whether all mineral deposits show a latitudinal control upon their formation, regardless of origin. The results showed that both distributions did not reflect a bias for certain latitudinal belts. Hence it was concluded that climate did not greatly influence the development of both PORCUs and EPIAUs. Such a conclusion supported the use of these deposit types to test the reliability of the data collection and processing methods used for the sediment-hosted deposit types.

Early on in this project an attempt was made to determine the palaeolatitudinal distributions of deposits of Proterozoic age i.e. PAPL, FEFM, SSCU, SHBM and SHUR deposit types. However the results were too inconsistent to have any value to this discussion due to the poor reliability and selection of palaeomagnetic data and so they have not been presented here. Despite this, future research into the distribution of these types is strongly recommended. They may be able to contribute to the discussion on the reliability of Proterozoic reconstructions, the quality of the palaeomagnetic data and the problem of whether to accept the Uniformitarian Principle or not - more specifically, the debate as to the origin of these deposits includes discussion on the possibility of a Proterozoic anoxic/oxic atmosphere at the time of formation.

Many ore deposits are found in regions where microplates dominate such as Indonesia and Mexico. This suggests that the high tectonic energy conditions of these areas facilitate mineral deposit development. High heat flow characteristic of such regions would also assist in the warming of metal-bearing solutions. The tectonic environment generally associated with mineral deposit formation is one of rifting and transgression over shallow continental margins; the first criterion may be provided by microplate activity. A greater degree of subduction may also provide the primary release of metals for later concentration as deposits.

One implication of the palaeolatitudinal control on the formation of mineral deposits is that it may be used to solve certain geological problems. In particular the correct age of mineralisation of a deposit may be discerned if a palaeolatitude determined for one age is more appropriate (i.e. within the range of the majority of palaeolatitudes for that deposit type) than that for another age. However care must be taken to classify the deposit correctly and so compare with the relevant palaeolatitudinal distribution. Such an application of the theory has been discussed in Chapter Seven, section 7.1.3 (a) with reference to the dating of LSBM deposits of southeast Missouri.

Another implication may be in the resolution of palaeogeographic reconstruction problems such as the position of Thailand during the Mesozoic. However it was not possible to use the palaeolatitudinal control on the formation of SDEXs to solve this problem because of lack of data as explained in Chapter Seven, section 7.1.3 (a). Such a use of the theory is difficult to apply at this stage and may be more successfully applied when the palaeolatitude influence on deposit distribution has been more clearly defined.

One use of the control in the understanding of mineral deposit formation can be illustrated with reference to SSCU deposits. One school of thought consider these deposits to be mainly syngenetic or diagenetic in origin - a hypothesis supported by the results given here. The epigenetic school of thought is refuted by the data presented here.

The significance of these results is that they may assist in the evaluation of genetic models of mineral deposit formation and the discovery of any genetic relationships between mineral deposit types. In addition such palaeolatitudinal controls on the distribution of some mineral deposits may be useful in the assessment of potential sites for exploration.

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APPENDIX ONE

Limestone Base Metal Deposits

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS		AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.	B.P. PALAT
327	Hallouf Deposit, Djebel, Tunisia	36.2	9.9	11 - 5	0 50 Afr	36.2 21.4	 22.0
331	Gombe Deposit, Benue Trough, Nigeria	10.3	10.2	85 - 70	50 Afr 100 Afr	-4.3 -9.4	-3.8 -4.2
332	Akwana Deposit,Benue Trough, Nigeria	7.8	9.3	85 - 70	50 Afr 100 Afr	-6.7 -11.3	-6.0 -6.3
333	Abakaliki Deposit, Benue Trough, Nigeria	6.3	8.1	85 - 70	50 Afr 100 Afr	-8.1 -12.2	-7.2 -7.4
286	Marion, Kentucky, USA	37.3	-88.1	97 - 88	100 N.Am	35.4	38.7
287	Rosiclare, Illinois, USA	37.5	-88.3	97 - 88	100 N.Am	35.6	38.9
305	Boccheggiano & Gavorrano Mines, Tuscany, Italy	43.3	11.1	5 & 200	0 200 Eur	43.3 38.1	 27.7
328	Bou Beker-Touissit District, Morocco	34.5	1.8	5 & 200	0 200 Afr	34.5 26.8	 22.0
329	Bedianne Deposit, Touissit-Bou Beker Area, Morocco	34.7	-2.4	24 & 200	0 200 Afr	34.7 28.4	 23.5
330	Oued Mekta, Touissit-Bou Beker Area, Morocco	34.5	-1.8	24 & 200	0 200 Afr	34.5 28.0	 23.1
312	Chorozow, Poland	50.3	18.9	213	200 Eur	46.6	42.2
306	Sedimochislenitsi Ore Area, Bulgaria	43.0	23.4	213	200 Eur	41.0	35.7
307	Chechlo Deposit, Poland	51.3	15.5	213	200 Eur	46.7	42.2

Limestone Base Metal Deposits cont.....

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.	B.P. PALAT
308	Wiktor Emanual, Poland	50.5 19.3	213	200 Eur	46.9	42.6
309	Zawiercie, Poland	50.7 19.5	213	200 Eur 250 CEur	46.9 9.2	42.6
313	Trzebionka & Matylda Mines, nr. Chrzanow, Austria	50.2 19.3	213	200 Eur 250 CEur	46.6 8.9	42.2
314	Bleiberg-Kreuth Deposit, Gailtalek Alps, Austria	46.6 13.7	231	250 SEur	4.2	
315	Mezica Deposit, Northern Karawankes, Yugoslavia	46.5 14.9	231	250 CEur	4.3	
316	Raibl (Cave del Predil) Dep, Udine, Italy	46.5 13.6	231	250 SEur	4.0	
317	Salafossa Deposit, Belluno Province,Italy	46.6 12.7	231	250 SEur	3.9	
318	Pila Deposit, Tribec Mts, Czechoslovakia	48.5 18.6	238	200 Eur 250 CEur	44.9 7.1	40.6
319	Ardovo Deposit,Juhoslovensky Kras Mts, Czechoslovakia	48.4 20.3	238	250 CEur	7.5	
321	Maluzina Deposit, Nizke Tatry Mts, Czechoslovakia	48.9 20.0	238	250 CEur	7.8	
323	NW of Polkowice & Lubin, Poland	51.5 16.2	286 - 248	250 CEur 300 CEur	9.3 5.4	
324	Bolzano Basin, Italy	46.6 11.2	286 - 248	250 SEur 300 SEur	3.6 -0.4	

Limestone Base Metal Deposits cont.....

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.
325	Bozen-Trient Province, Italy	46.7 11.3	286 - 258	250 Seur 300 Seur	3.7 -0.3
290	Picher, Oklahoma, USA	37.0 -94.9	320	300 Laur	-9.5
294	Oronogo, Missouri, USA	37.2 -94.5	320	300 Laur	-9.7
292	Baxter Springs, Kansas, USA	37.0 -94.0	320	300 Laur	-0.4
293	Cranby, Missouri, USA	36.9 -94.3	320	300 Laur	-0.3
289	Commerce, Oklahoma, USA	36.9 -94.9	320	300 Laur	0.0
291	Galena, Kansas, USA	37.0 -94.7	320	300 Laur 350 Laur	0.0 -9.7
295	Joplin, Missouri, USA	37.1 -94.5	320	300 Laur 350 Laur	0.0 -9.8
297	Pine Point, NWT, Canada	60.9-114.2	320 - 310	300 Laur 350 Laur	25.3 10.7
326	Mirgalimsai Deposit, Kazakhstan, USSR	44.4 52.2	352	350 Eur	-8.9
296	Magnet Cove Deposit, Nova Scotia, Canada	45.0 -65.0	360 - 320	300 Laur 350 Laur	-4.4 -21.7
298	Robb Lake Deposit, BC,Canada	56.9-124.8	408 - 360	350 Laur	15.5
299	Galena, Illinois, USA	42.4 -90.4	408	400 Laur	-26.0
300	Platteville, Wisconsin, USA	42.7 -90.5	408	400 Laur	-25.8

Oolitic Ironstone Deposits

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.	B.P. PALAT
574	Fisherman's Bay, SA, Australia	-33.6 138.0	1 - 0	0	-33.6	
580	Cucuta Deposit, Colombia	7.9 -72.5	11 - 5	0	7.9	
581	Lagunillas Deposit, Venezuala	10.1 -71.3	24 - 11	0	10.1	
575	Bida District, Middle Niger Valley, Nigeria	9.1 5.7	73 - 65	50 Afr	-5.0	-3.9
576	Lokoja District, Middle Niger Valley, Nigeria	7.2 6.7	73 - 65	50 Afr	-7.0	-6.0
577	Illo District, Middle Niger Valley, Nigeria	11.3 4.2	73 - 65	50 Afr	-2.6	-1.3
582	Paz de Rio Deposit, Colombia	6.2 -72.7	42 - 38	50 S.Am	2.7	7.9
583	Sabanalarga Deposit, Colombia	10.6 -74.9	42 - 38	50 S.Am	7.4	12.7
597	Wadi Fatima, Saudi Arabia	21.5 39.6	54 - 38	50 Arab	3.8	1.5
596	South Kelsey, Lincs, England	53.5 -0.4	131 - 125	100 Eur 130 Eur	37.1 48.4	45.4 39.4
571	Sturminster, Newton, Dorset, England	50.9 -2.3	159 - 156	130 Eur	45.9	36.8
572	Westbury, Wiltshire, England	51.3 -2.2	159 - 156	130 Eur	46.3	37.2
573	Red Down, Highworth, Wiltshire, England	51.6 -1.7	159 - 156	130 Eur	46.6	37.5

Oolitic Ironstone Deposits cont......

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.	B.P. PALAT
589	Boulby, nr. Brotton, Yorkshire, England	54.6 -0.9	206 - 194	200 Eur	46.1	41.0
590	nr. Grantham, Lincs, England	52.9 -0.6	206 - 194	200 Eur	44.5	39.1
591	nr. Banbury, Oxon, England	52.1 -1.3	206 - 194	200 Eur	43.6	38.0
592	nr. Northampton, Northants, England	52.2 -0.9	181 - 175	200 Eur	43.8	38.3
593	Corby, England	52.5 -0.7	181 - 175	200 Eur	44.1	38.6
595	nr. Towcester, Northants, England	52.1 -1.0	181 - 175	200 Eur	43.7	38.1
594	nr. Scunthorpe, Lincs, England	53.6 -0.6	206 - 200	200 Eur	45.2	40.0
569	Birmingham Red-Ore District, Alabama, USA	33.5 -86.8	420	400 Laur	-33.1	
570	Bell Island, Conception Bay, Newfoundland, Canada	47.7 -53.0	505 - 468	450 Laur	-47.3	
584	San Bernardo Deposit, Leon Province, Spain	42.4 -6.1	468 - 488	450 Seur	-37.0	
585	Vivaldi Deposit, Leon Province, Spain	42.5 -6.6	468 - 448	450 Seur	-37.1	
586	Vivero Deposit, Spain	43.6 -7.6	468 - 458	450 Seur	-36.4	
588	Villalba, Lugo Area, Spain	43.3 -7.8	468 - 458	450 Seur	-36.7	

Sandstone Copper Deposits

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.	B.P. PALAT
185	Pisakeri Mine, Bolivia	-17.1 -68.3	20 - 10	0 50 S.Am	-17.1 -20.9	 -16.0
186	Veta Verde, Bolivia	-17.4 -68.4	20 - 10	0 50 S.Am	-17.4 -21.2	 -16.3
187	Chacarilla, Bolivia	-17.5 -68.2	20 - 10	0 50 S.Am	-17.5 -21.4	 -16.5
188	Corocoro, Bolivia	-17.2 -68.5	20 - 10	0 50 S.Am	-17.2 -21.0	 -16.1
176	Cachoeiras Deposit, Cuanza, Angola	-11.0 14.0	90	100 Afr	-30.4	-26.6
177	Ain-Sefra, Algeria	32.7 -0.6	90	100 Afr	15.2	20.7
178	J.Bou-Kechba Deposit,Algeria	34.0 -2.5	90	100 Afr	17.1	22.5
179	Mazzer Deposit, Morocco	33.5 -3.2	90	100 Afr	16.9	22.3
180	Merija Deposit, Morocco	34.0 -3.0	90	100 Afr	17.3	22.9
181	English Mine, Angola	-12.6 13.4	119 - 97	100 Afr	-31.6	-28.1
182	Cachoeiras de Binga, Angola	-11.0 13.8	119 - 97	100 Afr	-30.3	-26.5
183	Zenza Deposit, Angola	-9.3 14.2	144 - 119	100 Afr 130 Afr	-29.4 -21.7	-24.9 -25.9
184	Novo Redondo Deposit, Angola	-11.2 13.9	144 - 130	130 Afr	-23.1	-27.4

Sandstone Copper Deposits cont....

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.	B.P. PALAT
163	Eureka Mine, New Mexico, USA	36.2-106.7	231 - 213	200 N.Am 250 Laur	23.5 11.5	14.9
164	Nacimiento Mine, New Mexico, USA	36.0-106.8	231 - 213	200 N.Am 250 Laur	23.4 11.4	14.8
165	Blue Bird Prospect, New Mexico, USA	35.8-106.9	231 - 213	200 N.Am 250 Laur	23.3 11.2	14.7
166	San Miguel Mine, New Mexico USA	35.2-106.9	231 - 213	200 N.Am 250 Laur	22.6 10.9	13.9
167	Stauber Mine, New Mexico,USA	34.8-105.0	231 - 213	200 N.Am 250 Laur	22.0 9.6	13.4
168	Pintada Mine, New Mexico,USA	35.0-104.8	258 - 255	250 Laur	9.6	
169	Rayo District, New Mexico, USA	34.5-106.5	258	250 Laur	10.2	
171	Courtney Mine, New Mexico, USA	33.0-105.7	268	250 Laur	8.7	
170	Blue Star Mine & Cole Mine, New Mexico, USA	34.5-106.4	268	250 Laur	10.2	
156	Westville, Nova Scotia, Canada	45.6 -62.7	340 - 310	300 Laur 350 Laur	-4.3 -22.1	
157	Scotsburn, Nova Scotia, Canada	45.8 -62.9	340 - 310	300 Laur 350 Laur	-4.2 -22.0	
158	Memramcook, New Brunswick, Canada	46.0 -64.6	340 - 310	300 Laur 350 Laur	-3.5 -21.0	

Sandstone Copper Deposits cont.....

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.
159	Antigonish, Nova Scotia, Canada	45.6 -62.0	340 - 310	300 Laur 350 Laur	-4.5 -22.4
160	Dorchester, New Brunswick, Canada	45.9 -64.5	340 - 310	300 Laur 350 Laur	-3.6 -21.9
161	Albert, New Brunswick, Canada	45.7 -64.8	340 - 310	300 Laur 350 Laur	-3.8 -21.2

Sandstone Lead Deposits

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.	B.P. PALAT
4	Bou-Sellam Deposit, Morocco	33.5 -4.0	90	100 Afr	17.2	22.5
5	Kroussou Deposit, Morocco	-1.5 10.0	90	100 Afr	20.1	-15.6
6	Loeto Deposit, Angola	-11.7 13.8	130	130 Afr	-23.4	-28.0
7	Wallau Deposit, Oberpflaz Area, West Germany	50.9 8.5	225 - 219	200 Eur	44.7	39.4
8	Eschenbach Deposit, Oberpflaz Area, West Germany	49.7 11.7	225 - 219	200 Eur	44.3	39.2
9	Freihung Deposit, Oberpflaz Area, West Germany	49.5 11.9	225 - 219	200 Eur	44.2	39.0
1	Warnock Mine, High Rolls District, USA	33.0-105.7	268	250 Laur	8.7	
2	L'Argentiere Deposit, France	44.5 4.3	245 - 238	250 Ceur	0.3	
10	Mechernich, West Germany	50.6 6.6	248 - 243	250 Ceur	6.6	
11	Boumia, Morocco	32.6 -5.0	310	300 Gond	-18.8	

SHALE BASE METAL DEPOSITS

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS		AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.	B.P. PALAT
110	Slovenia, Yugoslavia	46.4	16.2	286 - 213	200 Eur 250 Ceur 300 Ceur	42.3 4.6 0.8	36.8
111	Medicine Mounds, Texas, USA	34.2	99.6	255	250 Laur	4.9	
112	Old Glory, Texas, USA	33.1	100.0	255	250 Laur	4.6	
113	Crowell, Texas, USA	34.0	99.7	255	250 Laur	4.8	
115	Mangum, Oklahoma, USA	34.9	99.5	255	250 Laur	5.1	
117	Eisleben, East Germany	51.4	11.5	258	250 Ceur	8.3	
118	Mansfeld, East Germany	51.6	11.5	258	250 Ceur	8.4	
119	Witzenhausen, West Germany	51.3	9.8	258	250 Ceur	7.8	
120	Walkenreid, West Germany	51.6	10.1	258	250 Ceur	8.2	
121	Linsburg, West Germany	52.6	9.3	258	250 Ceur	9.0	
122	Huggel, nr. Osnabruck, West Germany	52.3	8.0	258	250 Ceur	8.5	
123	Gronigen, Holland	53.2	6.6	258	250 Eur	9.1	
124	Durham, England	54.8	-1.6	258	250 Ceur	9.7	
125	Lubin, Poland	51.4	16.2	258	250 Ceur	9.2	

Shale Base Metal Deps cont....

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.
129	Chicote Grande, Bolivia	-17.2 -67.0	300	300 Gond	17.1
130	Toropaica, Bolivia	-20.3 -65.7	300	300 Gond	14.6
131	Cordillera Real (North and South), Bolivia	-19.0 -66.0	300	300 Gond	15.5
126	Frances Lake, Selwyn Basin, Canada	61.3 -129.0	458 - 438	450 Laur	-21.6

Sedimentary Exhalative Deposits

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.	B.P. PALAT
1981	Reocin Deposit, Santander Province, Spain	43.3 -4.1	119 - 113	100 Eur	27.5	32.6
1977	Sidi-Lahlen, S. of Taourit, Morocco	34.4 -2.9	213 - 188	200 Afr	28.3	23.4
1978	Beni Tadjit Area, Morocco	32.2 -3.5	213 - 188	200 Afr	26.6	21.4
1979	Talsint Area, Morocco	32.5 -3.5	213 - 188	200 Afr	26.8	21.7
1980	Mibladen-Aouli, Nr. Midlet, Morocco	32.7 -4.7	213 - 188	200 Afr	27.5	22.2
1982	Mae Sod, Thailand	16.7 98.5	248 - 144	130 Sund 200 Sund 250 Sund	13.5 -3.2 -24.9	5.6 37.0
1913	Red Dog, De Long Mountains, Alaska, USA	68.5 -64.0	300	300 Laur	17.7	
1910	Kansas City Area, Kansas,USA	39.0 -95.0	320 - 286	300 Laur	1.7	
1911	Chamberlain Creek Syncline, Arkansas, USA	34.5 -93.7	333 - 320	300 Laur 350 Laur	-2.5 -11.6	
1912	Fancy Hill District, Ouchita Mts, Arkansas, USA	34.5 -94.4	333 - 320	300 Laur 350 Laur	-2.1 -11.1	
1914	Rubiales Deposit, Spain	42.6 -6.7	360 - 286	300 Seur 350 Seur	-7.5 -28.1	
1915	Keel Deposit, Ireland	54.0 -10.1	360 - 320	300 Ceur 350 Ceur	3.5 -19.2	

Sedimentary Exhalative Deposits cont....

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.
916	Tynagh Deposit, Ireland	53.2 -8.4	360 - 320	300 Ceur 350 Ceur	2.8 -20.1
917	Silvermines Deposit, Ireland	52.8 -8.2	360 - 320	300 Ceur 350 Ceur	2.4 -20.5
918	Navan Deposit, Ireland	53.6 -6.7	360 - 320	300 Ceur 350 Ceur	3.4 -19.7
919	Ballinalack, West Meath, Ireland	53.6 -7.5	360 - 320	300 Ceur 350 Ceur	3.3 -19.7
920	Pontebba-Sappada, Austria-Italy	46.6 13.0	360 - 320	300 Seur 350 Ceur	0.1 -23.7
906	Tom & Jason Deposits, Yukon, Canada	63.2-130.2	374 - 360	350 Laur	18.4
966	Narharla Deposit, Kimberley, Australia	-17.2 124.7	374 - 360	350 Gond	-2.8
973	Dzhezkazgan, Kazkhstan, USSR	47.8 67.4	374 - 360	350 Sib	23.8
974	Atasu, Kazakhstan, USSR	48.7 71.6	374 - 360	350 Sib	26.7
924	Chaudfontaine Deposit, Belgium	50.6 5.7	374 - 367	350 Ceur	-22.2
925	Booishot Drill Hole, Belgium	51.0 4.8	374 - 367	350 Ceur	-21.9
926	Heibaart 1 Drill Hole, Belgium	51.4 4.7	374 - 367	350 Ceur	-21.5

Sedimentary Exhalative Deposits cont....

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS		AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.
1975	Achisai Deposit, Kazakhstan, USSR	43.2	68.9	380 - 352	350 Sib 400 Sib	23.5 21.0
1929	Meggen Deposit, West Germany	51.1	8.1	386 - 374	350 Ceur 400 Ceur	-21.4 -35.2
1921	Anterselva Deposit, Italy	46.6	10.7	408 - 360	350 Seur 400 Seur	-24.0 -34.6
1930	Rammelsberg Deposit, West Germany	51.9	10.4	387 - 380	400 Ceur	-33.9
1937	Sarrouyes Deposit, Central Pyrenees, France	42.7	0.5	394 - 387	400 Ceur	-45.0
1939	Pene det Pouri, Central Pyrenees, France	42.9	0.1	394 - 387	400 Ceur	-44.9
1940	Arrens Deposit, Pyrenees, France	42.9	-0.2	401 - 394	400Ceur 450 Ceur	-45.0 -24.8
1934	Frohnleiten-Peggau District, France	47.2	15.3	408 - 387	400 Seur	-32.4
1935	Schrems Deposit, Styria, Austria	48.8	15.1	408 - 387	400 Seur	-31.2
1931	Bodennec, Britanny, France	48.6	-3.7	408 - 387	400 Ceur	-40.2
1932	La Porte aux Moines, Britanny France	48.4	-2.9	408 - 387	400 Ceur	-40.3

Sedimentary Exhalative Deposits cont.....

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.
933	Nerbiou, Hautes Pyrenees, France	43.0 -0.1	408 - 387	400 Ceur	-44.9
941	Saube Deposit, Pyrenees, France	42.7 1.1	408 - 401	400 Ceur 450 Ceur	-44.9 -24.2
969	Phu Mai Tong District, Thailand	16.8 101.0	420 - 380	400 Sund	11.1
907	Vulcan Prospect, Logan Mts., Canada	62.3-128.2	421 - 387	400 Laur	1.7
946	Reichertleiten Alun, Scharrn Alm, Austria	47.6 12.5	438 - 408	400 Seur 450 Seur	-33.1 -25.2
948	Estaing Deposit, Pyrenees, France	43.0 -0.2	448 - 438	450 Ceur	-24.7
950	Cheze Deposit, Pyrenees, France	42.9 0.0	448 - 438	450 Ceur	-24.6
951	Artigues Deposit, Pyrenees, France	42.8 0.6	448 - 438	450 Ceur	-24.3
952	Liat Deposit, Pyrenees, France	42.8 0.8	448 - 438	450 Ceur	-24.1
956	Bulard Deposit, Pyrenees, France	42.8 1.0	448 - 438	450 Ceur	-24.0
957	Bosost Deposit, Spain	42.8 0.7	448	450 Seur	-34.2
905	Logan Mts., Yukon, Canada	62.5-129.3	458 - 438	450 Laur	-9.4

Sedimentary Exhalative Deposits cont...

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE-LAT.
963	Crabioules Deposit, Pyrenees, France	42.7 0.5	458 - 448	450 Ceur	-24.4
976	La Helvecia, Argentina	-29.5 -68.8	468 - 458	450 Gond	7.1
904	Summit Lake, Yukon, Canada	62.6 -129.6	505 - 438	450 Laur	-9.2
970	Bo Noi District, Meklong Highlands, Thailand	15.3 98.7	505 - 468	450 Sund	9.6
971	Bo Yai District, Meklong Highlands, Thailand	14.3 99.1	505 - 468	450 Sund	10.6
972	Bawdwin District, Burma	23.1 97.3	520 - 478	450 Sund	3.2
945	La Troya, Asturin Province, Spain	42.0 -6.5	590 - 408	400 Seur 450 Ceur	-43.4 -29.3

Sandstone U-V Deposits

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.	B.P. PALAT
272	Kalka Deposit, Himachal Pradesh, India	30.8 76.9	5 - 0	0 50 Ind	30.8 1.4	-6.9
273	Ramshahr-Kalka-Morni Region, India	30.6 76.7	5 - 0	0 50 Ind	30.7 1.3	-7.0
274	Jammu Region, India	32.7 74.9	5 - 0	0 50 Ind	32.7 3.7	-4.6
275	Himachal Region, India	31.7 77.2	5 - 0	0 50 Ind	31.7 2.2	-6.0
276	Haryana Region, India	28.8 76.2	5 - 0	0 50 Ind	28.8 -0.4	-8.7
258	Yotsugi Deposits, Honshu, Japan	35.1 134.1	10	0	35.1	
259	Nakatsugo Deposits, Honshu, Japan	34.9 133.9	10	0	35.0	
271	Baghal Chur Deposit, Sulaiman Range, Pakistan	30.3 70.4	14 - 1	0	30.3	
265	Sulaiman Range, Pakistan	29.9 70.3	14 - 1	0	29.9	
267	Sulaiman Range, Pakistan	29.7 70.1	14 - 1	0	29.7	
268	Sulaiman Range, Pakistan	29.4 70.0	14 - 1	0	29.6	
261	Tsukiyoshi Deposit, Tono Min Japan	39.3 141.5	14 - 11	0	39.3	

Sandstone U-V Deposits cont....

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE-LAT.	B.P. PALAT
262	Misano Deposit, Gifu Prefecture, Japan	39.5 141.7	14 - 11	0	39.5	
263	Jorinji Deposit, Gifu Prefecture, Japan	39.4 141.3	14 - 11	0	39.4	
248	Beverley Deposits, South Australia, Aust.	-30.3 139.7	24 - 5	0	-30.3	
196	Pumpkin Buttes, Wyoming, USA	43.9-105.8	38 - 10	0 50 N.Am	43.9 47.3	 63.5
197	Monument Hill District, Wyoming, USA	43.3-105.3	38 - 10	0 50 N.Am	43.3 46.6	 62.4
198	Box Creek District, Wyoming, USA	43.1-105.1	38 - 10	0 50 N.Am	43.1 47.4	 63.6
199	Ross District, Wyoming, USA	43.5-105.9	38 - 10	0 50 N.Am	43.5 47.0	 62.9
249	Curnamona Channel, Frome Lake Area, Australia	-31.7 139.6	65 - 25	50 Aust	-54.3	-63.7
257	Phu Wiang Deposits, Khorat Plateau, Thailand	16.7 102.2	125	130 Sund	13.8	6.6
208	Radium King Mine, Red Canyon Utah, USA	37.5-110.3	144 - 65	50 N.Am 100 N.Am 130 N.Am	42.5 43.7 41.7	55.3 46.3 46.6
210	Atomic No.1 Mine, White Canyon, Utah, USA	37.8-110.3	144 - 65	50 N.Am 100 N.Am 130 N.Am	42.8 44.0 42.0	55.7 46.7 47.0

Sandstone U-V Deposits cont.....

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.	B.P. PALAT
212	Rusty Mine, White Canyon, Utah, USA	37.5-110.1	144 -65	50 N.Am	42.6	55.5
				100 N.Am	43.8	46.5
				130 N.Am	41.8	46.8
215	Happy Jack Mine, White Canyon Utah, USA	37.7-110.1	144 - 65	50 N.Am	42.7	55.6
				100 N.Am	43.9	46.6
				130 N.Am	45.4	46.9
216	Monument No.2 Mine, Arizona, USA	36.9-109.8	144 - 65	50 N.Am	41.8	54.3
				100 N.Am	43.0	45.4
				130 N.Am	41.1	45.7
217	Big Four Mine, Monument Valley, Arizona, USA	36.9-110.2	144 - 65	50 N.Am	41.9	54.5
				100 N.Am	43.2	45.6
				130 N.Am	44.6	45.9
219	Moonlight Mine, Monument Valley, Arizona, USA	37.0-110.3	144 - 65	50 N.Am	42.0	54.5
				100 N.Am	43.3	45.6
				130 N.Am	44.7	46.0
221	Taylor-Reid Mine, Monument Valley, Utah, USA	37.2-110.3	144 - 65	50 N.Am	42.1	54.7
				100 N.Am	43.4	45.8
				130 N.Am	41.3	46.1
224	Moki Mine, Needles County, Utah, USA	38.2-109.7	144 - 65	50 N.Am	43.0	56.1
				100 N.Am	44.1	46.9
				130 N.Am	42.3	47.2
225	Glade-Abe Mine, Elk Ridge, Utah, USA	38.0-109.8	144 - 65	50 N.Am	42.7	55.1
				100 N.Am	43.9	46.6
				130 N.Am	45.1	46.9
226	East Payday Mine, Elk Ridge, Utah, USA	37.8-109.8	144 -65	50 N.Am	42.7	55.6
				100 N.Am	43.8	46.8
				130 N.Am	42.0	46.5

Sandstone U-V Deposits cont.....

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.	B.P. PALAT
227	Lucky Lady Mine, Elk Ridge, Utah, USA	37.6-109.8	144 -65	50 N.Am	42.6	55.5
				100 N.Am	43.7	46.7
				130 N.Am	41.9	46.4
201	Black Hills, Wyoming, USA	44.3-105.2	144 -125	100 N.Am	47.8	53.5
				130 N.Am	47.6	52.7
202	Grants Region, New Mexico, USA	35.2-108.0	150 - 130	130 N.Am	39.1	42.8
204	Laguna District,Grants Dist. New Mexico, USA	35.0-107.6	150 - 130	130 N.Am	38.9	42.4
205	Church Rock,Grants District, New Mexico, USA	35.5-108.7	150 - 130	130 N.Am	39.5	43.5
206	Smith Lake, Grants District, New Mexico, USA	35.4-108.1	150 - 130	130 N.Am	39.3	43.1
207	Ambrosia Lake, Grants Dist., New Mexico, USA	35.3-107.7	150 - 130	130 N.Am	39.2	42.9
239	Mecsek Mountains, Hungary	46.2 18.2	248	250 Ceur	4.9	
278	Chirmatekri, India	23.3 82.4	248	250 Gond	-28.3	
228	Rifle-Garfield Deposit, Colorado, USA	39.5-107.8	248 - 144	130 N.Am	43.3	47.9
				200 N.Am	26.9	18.5
				250 Laur	14.5	
235	Nr. Fieberbrun, Tyrol, Austria	47.5 12.5	248 - 213	200 Eur	41.7	36.8
				250 Seur	4.8	

Sandstone U-V Deposits cont.....

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.
240	Zirovski Vrh Deposit, Yugoslavia	46.1 14.4	253	250 Ceur	3.9
238	Val Rendena Mine, Lombard Deposits, Italy	46.0 11.0	258 - 248	250 Seur	3.0
241	Murtschenalp, Verrucano, Switzerland	47.1 9.1	260	250 Ceur	3.7
279	Figueira Area, Parana Basin, Brazil	-23.7 -50.7	265 - 255	250 Gond	-1.5
242	Obermoschel, West Germany	49.7 7.8	268	250 Ceur	6.0
236	Laguepie-Monesties District, France	44.1 2.0	270	250 Ceur 300 Ceur	-0.4 -4.8
244	St. Affrique,Massif Central, France	44.0 2.9	270	250 Ceur 300 Ceur	-0.4 -4.7
245	Brive District, Massif Central, France	45.1 1.5	270	250 Ceur	0.5
246	Rodex District, Massif Central, France	44.3 2.6	270	250 Ceur	-0.1
252	Fraserburg, South Africa	-31.8 21.5	280 - 250	250 Gond 300 Gond	-64.7 -62.4
253	Beaufort West, South Africa	-32.3 22.6	280 - 250	250 Gond 300 Gond	-64.4 -62.9

Sandstone U-V Deposits cont....

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS		AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.
243	Lodeve Basin,Massif Central, France	43.7	3.3	286 - 270	250 Ceur 300 Ceur	-0.6 -4.9
247	Northern Black Forest, East Germany	54.0	34.7	320 - 290	300 Ceur	13.4
250	Mount Eclipse, NT, Australia	-23.0	133.0	360 - 286	300 Gond 350 Gond	-11.1 4.5

Placer Gold Deposits

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.
340	Quebrada Ceibo Grand, Belize Mexico	16.6 -89.2	1 - 0	0	16.6
341	Yu-Mini & Tia-Ndiato, Oaxaca Mexico	16.8 -97.5	1 - 0	0	16.8
342	Bacubirito, Sinaloa, Mexico	25.8-107.9	1 - 0	0	25.8
343	Viznaga, Mexico	31.7-116.1	1 - 0	0	31.7
344	Rio Bobos, Guatemala	15.4 -88.7	1 - 0	0	15.4
345	La Canoa, Guatemala	14.9 -90.4	1 - 0	0	14.9
346	Rio Guayape, Honduras	14.7 -86.0	1 - 0	0	14.7
347	Lower Buller R., West Coast, SI, New Zealand	-41.8 172.0	1 - 0	0	-41.8
348	Hokitika Deposit, West Coast SI, New Zealand	-42.7 171.0	1 - 0	0	-42.7
349	Westport Deposit, West Coast SI, New Zealand	-41.8 171.6	1 - 0	0	-41.8
350	Charleston Deposit, West Coast, SI, New Zealand	-41.9 171.4	1 - 0	0	-41.9
351	Kumara, West Coast, SI, New Zealand	-42.6 171.2	1 - 0	0	-42.6
352	Lower Grey R. Valley, SI, New Zealand	-42.6 171.4	1 - 0	0	-42.6

Placer Gold Deposits cont.....

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.
353	Mont d'Or Claim, Ross, SI, New Zealand	-42.9 170.8	1 - 0	0	-42.9
354	Port Pegasus, Stewart Island New Zealand	-47.1 167.7	1 - 0	0	-47.1
355	Tin Range, Stewart Island, New Zealand	-46.7 167.8	1 - 0	0	-46.7
356	Orepuki Area, Southland, SI, New Zealand	-46.4 168.3	1 - 0	0	-46.4
372	Buller District, West Coast, SI, New Zealand	-41.7 172.1	1 - 0	0	-41.7
1002	Nr. Wau, Papua New Guinea	-7.4 146.7	1 - 0	0	-7.4
357	Wairau Valley, Marlborough, SI, New Zealand	-41.4 173.4	2 - 0	0	-41.4
358	Golden Block Lodes, Anatori R., SI, New Zealand	-40.5 172.6	2 - 0	0	-40.5
359	Takaka,Leslie & Karamea R.'s SI, New Zealand	-41.2 172.5	2 - 0	0	-41.2
360	Buller R., Maruia Tributary, SI, New Zealand	-41.7 172.4	2 - 0	0	-41.7
361	Marlborough Sounds District, SI, New Zealand	-41.1 173.9	2	0	-41.1
362	Preservation Inlet,Southland SI, New Zealand	-46.1 168.0	2	0	-46.1

Placer Gold Deposits cont......

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.	B.P. PALAT
335	Quartz Creek, Klondike, Yukon, Canada	63.8-139.1	3 - 1	0	63.8	
336	Bonanza Creek, Klondike, Yukon, Canada	63.9-139.4	3 - 1	0	63.9	
337	Hunker Creek, Klondike, Yukon, Canada	64.0-139.1	3 - 1	0	64.0	
338	Sulphur Creek, Klondike, Yukon, Canada	63.7-138.8	3 - 1	0	63.7	
339	Dominion Creek, Klondike, Yukon, Canada	63.5-138.8	3 - 1	0	63.5	
363	Hauraki Goldfield, S.Aukland, NI, New Zealand	-37.1 175.8	14 - 2	0	-37.1	
366	Ballarat Deposit, Victoria, Australia	-37.6 144.0	26 - 1	0	-37.6	
367	Bendigo Deposit, Victoria, Australia	-36.8 144.3	26 - 5	0	-36.8	
365	La Chiripa (Doradito), Nayarit, Mexico	22.2-105.2	38 - 2	0 50 NZ	22.2 26.6	34.3
364	Grey Mouth District, SI, New Zealand	-42.5 171.2	60 - 0	0 50 NZ	-42.5 -49.8	-51.6
368	Alexandra, Otago, SI, New Zealand	-45.2 169.4	65 - 1	0 50 NZ	-45.2 -52.5	-50.1

Placer Gold Deposits cont......

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.	B.P. PALAT
369	Ida Valley, Otago, SI, New Zealand	-45.0 169.9	65 - 1	0 50 NZ	-45.0 -52.1	 -49.8
370	Bannock Burn, Otago, SI, New Zealand	-45.1 169.2	65 - 1	0 50 NZ	-45.1 -52.6	 -49.9
371	Waikaia District, Otago, SI, New Zealand	-45.7 168.9	65 - 1	0 50 NZ	-45.7 -53.1	 -50.8
373	Glenore Deposit, Otago, SI, New Zealand	-46.1 169.9	88 - 65	50 NZ 100 NZ	-52.8 -59.3	-51.3
375	Naseby Deposit, Otago, SI, New Zealand	-45.0 170.2	144 - 97	100 NZ 130 NZ	-58.5 -49.3	
376	Fox R., SI, New Zealand	-42.0 171.3	144 - 125	130 NZ	-46.5	
377	Ahaura, West Coast, SI, New Zealand	-42.5 171.2	144 - 125	130 NZ	-46.9	
378	Koiterangi, West Coast, SI, New Zealand	-42.7 171.1	144 - 125	130 NZ	-47.1	
1021	Fairbanks District, Alaska, USA	65.0-147.7	0	0	65.0	
1022	Lena R., Siberia, USSR	64.5 127.0	1	0	64.5	
1023	Amur R., Siberia, USSR	53.5 122.5	1	0	53.5	
1024	Magdalena R., Colombia	8.5 -74.0	0	0	8.5	
1025	Yuba R., California, USA	39.2-121.7	0	0	39.2	

Placer Diamonds

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.
445	Cocalinho, Goias, Brazil	-14.7 -51.0	1 - 0	0	-14.7
446	Sono R., Goias, Brazil	-10.0 -47.5	1 - 0	0	-10.0
447	Near Craolandia, Goias, Brazil	-7.9 -47.1	1 - 0	0	-7.9
448	Cacu, Claro R., Goias,Brazil	-18.6 -51.1	1 - 0	0	-18.6
449	Verissimo R., Brazil	-19.7 -48.3	1 - 0	0	-19.7
450	Imperatriz, Tocantins R., Brazil	-5.5 -47.5	1 - 0	0	-5.5
451	Piui, Minas Gerais, Brazil	-20.5 -46.0	1 - 0	0	-20.5
452	Tibagi, Parana, Brazil	-24.5 -50.5	1 - 0	0	-24.5
527	Benton Harbor, Michigan, USA	47.1 -86.4	5 - 1	0	47.1
528	Kenosha, Wisconsin, USA	42.6 -87.8	5 - 1	0	42.6
529	Lorain, Ohio, USA	41.5 -82.2	5 - 1	0	41.5
530	Gary, Indiana, USA	41.6 -87.3	5 - 1	0	41.6
460	Hottentot Bay, Namibia	-26.0 14.8	20	0	-26.0
461	Orange Mouth, Namibia	-28.6 16.4	20	0	-28.6
462	Luderitz, Namibia	-26.6 15.2	20	0	-26.6

Placer Diamond Deposits cont....

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.	B.P. PALAT
459	Bingara, NSW, Australia	-29.8 150.6	53	50 Aust	-46.4	-53.1
463	West Oubangui Deposits, Sangha R., Central Afr. Rep.	4.3 15.6	71	50 Afr	-10.9	-10.9
464	West Oubangui Deposits, Mambere R.,Central Afr. Rep.	4.3 16.7	71	50 Afr	-11.0	-11.1
465	West Oubangui Deposits, Lobaye R., Central Afr. Rep.	3.8 17.8	71	50 Afr	-11.5	-11.9
466	East Oubangui Deposits, Bria Region,Central Afr.Rep.	6.5 22.0	71	50 Afr	-9.1	-10.0
467	East Oubangui Deposits, Mouka Region,Central Afr.Rep	7.3 21.9	71	50 Afr	-8.3	-9.2
468	East Oubangui Deposits, Ouadda Region, C. Afr. Rep.	8.1 22.3	71	50 Afr	-7.6	-8.4
469	Bakwanga District, Katanga Province, Zaire	-6.2 23.6	72	50 Afr	-21.9	-23.3
510	Gilbues, Piaui, Brazil	-9.8 -45.4	121 - 77	100 S.Am 130 S.Am	-16.9 -9.9	-14.5 -12.7
511	Coromandel, Minas Gerais, Brazil	-18.4 -47.2	121 - 77	100 S.Am 130 S.Am	-25.0 -18.7	-23.6 -21.5
513	Patos de Minas, Minas Gerais Brazil	-18.6 -46.5	121 - 77	100 S.Am 130 S.Am	-25.3 -18.8	-23.7 -21.7
514	Estrela do Sul, Minas Gerais Brazil	-18.7 -47.7	121 - 77	50 S.Am 100 S.Am 130 S.Am	-25.4 -25.2 -19.0	-21.3 -23.9 -21.8

Placer Diamonds deposits cont....

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.	B.P. PALAT
470	Hopetown, Orange R., S. Afr.	-29.6 24.1	90	100 Afr	-51.2	-52.4
471	Christiana, Vaal R., S. Afr.	-29.9 25.2	90	100 Afr	-50.2	-50.5
472	Bloemhof, Vaal R., S. Afr.	-27.6 25.6	90	100 Afr	-50.0	-50.3
473	Wedburg, Vaal R., S. Afr.	-28.3 24.7	90	100 Afr	-50.3	-50.9
476	Alexander Bay, Orange R., S. Afr.	-28.7 26.5	90	100 Afr	-47.1	-47.9
477	Kimberley, Vaal R., S. Afr.	-28.7 24.8	111	100 Afr	-50.7	-51.5
478	Klerksdorp, Vaal R., S. Afr.	-26.9 26.6	111	100 Afr	-49.8	-49.6
479	Migdol, Upper Harts R., S. Afr.	-26.9 25.4	111	100 Afr	-49.3	-49.2
480	Riet R., South Africa	-28.9 24.2	111	100 Afr	-50.6	-51.5
515	Caruachi Deposit, Venezuela	8.1 -62.9	85	100 S.Am	3.5	3.6
516	San Pedro de la Brocas, Bolivar Province, Venezuela	7.0 -62.9	85	100 S.Am	2.5	2.5
517	Cuchivero R., Venezuela	6.9 -65.7	85	100 S.Am	2.8	2.5
518	Coura R., Venezuela	6.2 -64.7	85	100 S.Am	2.0	1.8
520	Ventuari R., Venezuela	5.0 -66.3	85	100 S.Am	1.0	0.5

Placer Diamonds Deposits cont...

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.	B.P. PALAT
521	Suapure R., Venezuela	6.6 -67.0	85	100 S.Am	2.7	2.2
454	Phenix City, Alabama, USA	32.5 -85.0	150	130 N.Am	32.7	30.3
455	Charlotte, North Carolina, USA	35.3 -80.8	150	130 N.Am	34.7	31.5
456	Columbia, South Carolina,USA	34.0 -81.0	150	130 N.Am	33.5	30.4
458	Richmond, Virginia, USA	37.6 -77.4	150	130 N.Am	36.4	32.9
483	Tshikapa Region, West Kasai Province, Zaire	-5.7 20.8	136	130 Afr	-22.1	-25.9
484	Tshikapa Region, Bandundu Province, Zaire	-5.8 19.2	136	130 Afr	-21.3	-25.2
485	Bougande Area, Baghsalogo Region, West Africa	13.0 -0.1	137	130 Afr	4.8	1.5
486	Dunkwa Region, Ghana	6.0 -1.7	137	130 Afr	-0.2	-3.6
487	Aboissa District,Ivory Coast	5.4 -3.2	137	130 Afr	0.1	-3.3
488	Cavally Deposit, Ivory Coast	6.2 -8.2	137	130 Afr	3.6	0.0
489	Tienko District, Ivory Coast	10.1 -6.9	137	130 Afr	6.1	2.6
490	Bomi Hills, Liberia	6.9 -10.9	137	130 Afr	5.7	2.0
491	Katata Region, Liberia	6.6 -10.3	137	130 Afr	5.1	1.5

Placer Diamonds Deposits cont....

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.	B.P. PALAT
492	Sanniquelli District,Liberia	7.4 -8.7	137	130 Afr	4.9	1.3
493	Sansanto Deposits, Kenieba Region, Mali	13.0 -11.4	137	130 Afr	11.0	7.5
494	Zeerust Area, Lichtenburg District, S. Afr	-25.5 26.1	137	130 Afr	-41.4	-50.1
495	Lichtenburg District,Western Transvaal, S. Afr	-26.1 26.2	147	130 Afr	-41.9	-50.9
507	Barito R., Kalimantan,Borneo	-2.7 114.9	150	130 Sund	-4.2	-7.7
508	Landak R., Kalimantan,Borneo	0.0 109.4	150	130 Sund	-2.1	-7.0
509	Kapuas R., Kalimantan,Borneo	-2.2 114.3	150	130 Sund	-3.7	-7.5
512	Monte Carmelo, Minas Gerais, Brazil	-18.7 -47.5	121 - 77	100 S.Am 130 S.Am	-25.2 -19.0	-23.7 -21.8
496	Ngamo Area, Zimbabwe	-19.1 27.5	300	300 Gond	-70.6	
497	Gwelo District, Zimbabwe	-19.4 29.9	300	300 Gond	-72.8	
498	Willoughby's Spur, Zimbabwe	-19.6 29.7	300	300 Gond	-72.7	
522	Rio Abaete District, Brazil	-19.2 -45.4	350	350 Gond	-9.1	
523	Grao Mongol District, Minas Gerais, Brazil	-16.5 -42.8	350	350 Gond	-11.5	
524	Bagagem District, Minas Gerais, Brazil	-14.6 -48.1	350	350 Gond	-6.4	

Placer Diamonds Deposits cont......

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.
525	Diamantina Deposits, Minas Gerais, Brazil	-18.3 -43.6	350	350 Gond	-10.8
526	Aripuana Deposits, Mato Grosso, Brazil	-9.2 -60.7	350	350 Gond	6.3
501	Kleinzee, Southern Namaqualand, S. Afr	-29.7 17.0	400	400 Gond	-58.0
499	Nababiep, Buffels R., Namaqualand, S. Afr	-29.6 17.8	65 & 400	50 Afr 400 Gond	-49.4 -57.6
502	Zwartlintjies R., Southern Namaqualand, S. Afr	-30.3 17.4	65 & 400	50 Afr 400 Gond	-50.3 -57.4
503	Spoeg R., Southern Namaqualand, S. Afr	-30.5 17.4	65 & 400	50 Afr 400 Gond	-50.5 -57.1
504	Groen R., Southern Namaqualand, S. Afr	-31.5 17.8	65 & 400	50 Afr 400 Gond	-52.0 -56.1
505	Bitter R., Southern Namaqualand, S. Afr	-30.9 17.6	65 & 400	50 Afr 400 Gond	-51.1 -56.7
506	Olifants R., Southern Namaqualand, S. Afr	-31.7 18.1	65 & 400	50 Afr 400 Gond	-52.4 -55.8

Placer Tin Deposits

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.
379	Sapioris, Durango, Mexico	25.2-104.9	1 - 0	0	25.2
380	Mountain Pine Ridge,Belize, Mexico	17.1 -88.9	1 - 0	0	17.1
381	West Coast, SI, New Zealand	-43.5 169.5	1 - 0	0	-43.5
382	West Coast, SI, New Zealand	-43.7 169.9	1 - 0	0	-43.7
384	West Coast, SI, New Zealand	-43.0 170.6	1 - 0	0	-43.0
385	West Coast, SI, New Zealand	-42.8 170.8	1 - 0	0	-42.9
386	West Coast, SI, New Zealand	-42.5 171.2	1 - 0	0	-42.5
387	West Coast, SI, New Zealand	-42.5 172.1	1 - 0	0	-42.5
388	West Coast, SI, New Zealand	-42.4 171.7	1 - 0	0	-42.4
390	West Coast, SI, New Zealand	-42.3 171.4	1 - 0	0	-42.3
393	West Coast, SI, New Zealand	-41.9 171.7	1 - 0	0	-41.9
394	West Coast, SI, New Zealand	-41.7 172.0	1 - 0	0	-41.7
1007	Gibsonvale, NSW, Australia	-33.7 146.7	1 - 0	0	-33.7
1009	Mount Wills, Victoria, Australia	-36.6 147.5	1 - 0	0	-36.6
395	nr. Port Pegasus,Stewart I., New Zealand	-47.1 167.8	2 - 0	0	-47.1

Placer Tin cont......

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.
396	Mount Garnet, nr. Herberton, Queensland, Australia	-17.4 145.9	10 - 0	0	-17.4
397	Blue Tier Area, Tasmania	-41.4 147.1	10 - 0	0	-41.4
1010	Tandjung Pandan, Billiton, Indonesia	-2.7 107.6	0	0	-2.7
1011	Tg. Modong, Billiton, Indonesia	-2.7 108.0	0	0	-2.7
1012	Toboali, Bangka I.,Indonesia	-3.0 106.4	0	0	-3.0
1013	Belinju, Bangka I.,Indonesia	-1.6 105.8	0	0	-1.6
1014	Selangor, Malaya, Malaysia	3.3 101.5	0	0	3.3
1019	Ipoh, Malaysia	4.6 101.1	0	0	4.6
1020	Telok Anson, Malaysia	4.0 101.2	0	0	4.0

Placer Oxide Deposits

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.
398	Coyutlan, Colima, Mexico	18.1-104.0	1 - 0	0	18.1
399	Manzanillo, Colima, Mexico	19.0-104.3	1 - 0	0	19.0
400	Cuilapan, Oaxaca, Mexico	17.0 -96.8	1 - 0	0	17.0
401	Laguna de los Micos,Honduras	15.8 -87.6	1 - 0	0	15.8
403	Waiuku State Forest, NI, New Zealand	-37.2 174.7	1 - 0	0	-37.2
404	Karamea Beaches,West Coast, SI, New Zealand	-41.2 172.1	1 - 0	0	-41.2
405	West Port, West Coast, SI, New Zealand	-41.8 171.6	1 - 0	0	-41.8
406	Fiordland, SI, New Zealand	-45.0 167.5	1 - 0	0	-45.0
407	Stewart Island, New Zealand	-46.7 168.0	1 - 0	0	-46.7
408	Wanganui, NI, New Zealand	-39.9 175.0	1 - 0	0	-39.9
409	Taranaki, New Plymouth, NI, New Zealand	-38.0 174.8	1 - 0	0	-38.0
410	Orepuki, Southland, SI, New Zealand	-46.4 168.2	1 - 0	0	-46.4
411	Orepuki, Southland, SI, New Zealand	-46.2 167.6	1 - 0	0	-46.2

Placer Oxide Deposits cont....

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.
412	Raglan, NI, New Zealand	-37.8 174.9	1 - 0	0	-37.8
413	West Port Beaches,West Coast SI, New Zealand	-41.8 171.6	1 - 0	0	-41.8
1003	Tuggerah, NSW, Australia	-33.3 151.4	1 - 0	0	-33.3
1004	Bonny Hills, Australia	-31.5 152.4	1 - 0	0	-31.5
1005	Moreton I.,Queensland, Aust.	-27.3 153.0	1 - 0	0	-27.3
1008	Eneabba, West Australia,Aust	-30.0 115.0	1 - 0	0	-30.0
414	Muriwai-Whatipu, NI, New Zealand	-38.7 177.9	2 - 0	0	-38.7
415	Manukau Peninsula, NI, New Zealand	-37.0 174.5	2 - 0	0	-37.0
416	Kawhia-Aotea, NI,New Zealand	-37.8 174.6	2 - 0	0	-38.0
402	Lakehurst Area, NJ, USA	40.0 -74.3	5 - 0	0	40.0
418	Taharoa-Marakopa, NI, New Zealand	-38.4 174.6	5 - 0	0	-38.4
419	Gral. Zepeda, Mexico	25.4-101.5	144 - 88	100 C.Am	29.6
1015	Pulmoddai area, Sri Lanka	9.2 80.9	0	0	9.2
1017	Richard's Bay, Natal, S. Af.	-28.8 32.1	0	0	-28.8

Placer Other Deposits

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.
420	Cuiaba, Mato Grosso, Brazil	-15.5 -56.1	1 - 0	0	-15.5
421	Coxipo R., Mato Grosso, Brazil	-15.3 -56.0	1 - 0	0	-15.5
422	Jauru R., Mato Grosso,Brazil	-18.5 -54.3	1 - 0	0	-18.5
424	Ouro Preto, Minas Gerais, Brazil	-20.9 -43.5	1 - 0	0	-20.9
425	Cubarao, Santa Catarina, Brazil	-28.5 -49.0	1 - 0	0	-28.5
426	Teofilo Otoni, Minas Gerais, Brazil	-17.9 -41.3	1 - 0	0	-17.9
427	Pedra Azul, Minas Gerais, Brazil	-16.0 -41.3	1 - 0	0	-16.0
428	Campo Belo, Minas Gerais, Brazil	-20.9 -45.3	1 - 0	0	-20.9
429	Diamantina, Minas Gerais, Brazil	-18.3 -43.6	1 - 0	0	-18.3
430	Governador Valaderes, Minas Gerais, Brazil	-18.8 -41.9	1 - 0	0	-18.8
431	Muzo, Colombia	5.6 -74.1	1 - 0	0	5.6
432	Trueno, Nacimiento, Sonora, Mexico	28.2-109.7	1 - 0	0	28.2

Placer Other Deposits cont....

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.
433	Yamba Lake, NWT, Canada	65.2-111.4	1 - 0	0	65.2
434	Mt. Arthur Tableland,Nelson, SI, New Zealand	-41.2 173.0	1 - 0	0	-41.2
435	Glenorchy District, Otago, SI, New Zealand	-44.9 168.4	1 - 0	0	-44.9
436	Dun Mountain, Lee R.,Nelson, SI, New Zealand	-41.4 173.3	1 - 0	0	-41.4
437	Macraes Flat, Otago, SI, New Zealand	-45.4 170.4	1 - 0	0	-45.4
438	Nr. Hokitika, West Coast,SI, New Zealand	-42.7 171.0	1 - 0	0	-42.7
439	Nr. Reefton, West Coast,SI, New Zealand	-42.1 171.8	1 - 0	0	-42.1
440	Orepuki District, Southland, SI, New Zealand	-46.3 167.7	1 - 0	0	-46.3
441	Mangles Valley, Murchison, SI, New Zealand	-41.7 172.4	1 - 0	0	-41.7
442	Mangaorongo R., NI, New Zealand	-38.3 174.7	1 - 0	0	-38.3
443	Owharoa, W. of Waihi, NI, New Zealand	-37.4 175.8	1 - 0	0	-37.4

Placer Other Deposits cont....

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.	B.P. PALAT
1006	Capel, W. Aust., Australia	-33.6 115.4	1 - 0	0	-33.6	
444	Golden Bay, Nelson, SI, New Zealand	-40.6 172.8	65	50 NZ	-47.6	-49.7
1016	Pulmoddai area, Sri Lanka	9.2 80.9	0	0	9.2	
1018	Richard's Bay, Natal, S. Af.	-28.8 32.1	0	0	-28.8	

Manganese Deposits

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.	B.P. PALAT
677	Autlan, Jalisco, Mexico	19.9-104.3	38 - 2	0 50 C.Am	19.9 24.1	 31.4
690	Tchiatura Deposits, USSR	42.3 43.3	38 - 24	0 50 Eur	42.3 33.9	 35.3
691	Nikopol Deposits, USSR	47.6 34.4	38 - 24	0 50 Eur	47.6 38.7	 41.2
692	Bolshe Tomak Deposit, USSR	47.2 35.7	38 - 24	0 50 Eur	47.2 38.4	 40.9
693	Varna District, USSR	43.2 27.9	38 - 24	0 50 Eur	43.2 34.2	 35.7
679	Groote Eylandt, Australia	-14.0 136.0	96 - 94	100 Aust	-47.4	-41.9
694	Imini Mine, Morocco	30.7 -6.9	144 - 97	100 Afr 130 Afr	15.9 22.9	20.7 20.1
680	Molango, Hidalgo, Mexico	20.8 -98.7	243 - 144	130 C.Am 200 C.Am 250 Laur	23.5 21.5 -4.5	23.5 2.6
695	Urkut District, Hungary	47.1 17.7	213 - 188	200 Eur	43.3	38.1
696	Epleny Mine, Hungary	47.2 17.9	213 - 188	200 Eur	43.5	38.4
697	Molango Mine, Mexico	20.8 -98.7	213 - 188	200 C.Am	21.5	2.6
698	Leiping District, China	23.5 109.5	286 - 258	250Chi/J 300Chi/J	-12.8 -17.2	
699	Tsunyi District, China	27.5 106.5	286 - 258	250Chi/J 300Chi/J	-11.9 -17.5	

Laterite Deposits

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.
48	Samar Island, Phillipines	12.0 125.0	2 - 0	0	12.0
49	Rennell Island, Solomon Is.,	-11.7 160.2	2 - 0	0	-11.7
50	Wagina Islands, Solomon Is.,	-7.4 157.7	2 - 0	0	-7.4
51	Niue Island, Tonga Group	-19.0-169.9	2 - 0	0	-19.0
52	Lau Islands, Fiji	-17.2-179.0	2 - 0	0	-17.2
53	Mare Island, Loyalty Group	-21.5 168.0	2 - 0	0	-21.5
54	Lifou Island, Loyalty Group	-21.0 167.0	2 - 0	0	-21.0
64	Cayman Grand Island, Cayman Islands	19.3 -81.2	5 - 2	0	19.3
59	La Vega, Dominion Republic	19.2 -70.5	11 - 2	0	19.2
60	St. Marc, Haiti	19.1 -72.7	11 - 2	0	19.1
61	New Port,Manchester Plateau, Jamaica	17.9 -77.3	11 - 2	0	17.9
62	Bog Walk, St.Catherine Plateau,Jamaica	18.1 -77.0	11 - 2	0	18.1
63	Clarendon Plateau, Jamaica	18.0 -77.4	11 - 2	0	18.0
57	Cockpit County, Jamaica	18.3 -77.7	24 - 0	0	18.3

Laterite Deposits cont....

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.	B.P. PALAT
55	Lang Son, Vietnam	21.8 106.7	24 - 2	0	21.8	
56	Cao Bang, Vietnam	22.7 106.3	24 - 2	0	22.7	
58	Brown's Town, St.Ann Plateau Jamaica	18.5 -77.4	24 - 2	0	18.5	
78	Montufar, Guatemala	15.4 -89.1	38 - 2	0 50 C.Am	15.4 15.8	22.8
79	Niquegua (Eximbal), Guatemala	15.5 -89.4	38 - 2	0 50 C.Am	15.5 16.0	23.0
90	Paraiso, Mexico	24.7-104.0	38 - 2	0 50 C.Am	24.7 28.7	36.8
91	Vaquerias (El Sabinal), Hidalgo, Mexico	20.3 -98.6	38 - 2	0 50 C.Am	20.3 23.0	30.5
92	Riddle, Oregon, USA	42.9-123.4	50 - 0	0 50 N.Am	42.9 50.9	67.7
93	Moa, Cuba	20.7 -74.9	50 - 0	0 50 C.Am	20.7 17.1	24.5
94	Livingston, Guatemala	15.8 -88.7	50 - 0	0 50 C.Am	15.8 16.1	23.1
95	Kosovo, Yugoslavia	42.6 21.1	50 - 0	0 50 Eur	42.6 33.5	34.8

Laterite Deposits cont....

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.	B.P. PALAT
96	Marmara, Greece	38.8 22.1	50 - 0	0 50 Eur	38.8 29.7	19.3
97	Kharkov, Ukraine, USSR	50.0 36.2	50 - 0	0 50 Eur	50.0 41.2	44.5
98	Orsk, Urals, USSR	51.2 58.6	50 - 0	0 50 Eur	51.2 43.8	48.1
99	Greenvale, Queensland, Australia	-18.9 145.1	50 - 0	0 50 Aust	-18.9 -41.2	-43.8
100	Buhiga, Burundi	-3.0 30.1	50 - 0	0 50 Afr	-3.0 -19.0	-21.1
101	Nr. Mankono, Ivory Coast	8.0 -6.1	50 - 0	0 50 Afr	8.0 -4.2	-1.7
102	Thio, New Caledonia	-21.6 166.2	50 - 0	0 50 Sund	-21.6 -52.5	
103	Sulawesi, Indonesia	-2.8 121.5	50 - 0	0 50 Sund	-2.8 -11.7	
104	Gebe Island, Halmahera, Indonesia	-0.1 129.5	50 - 0	0 50 Sund	-0.1 -14.4	
106	San Fernando de Atabapo, Venezuala	4.0 -67.7	50 - 0	0 50 S.Am	4.0 -0.2	4.9
107	Tunja, Colombia	5.5 -73.4	50 - 0	0 50 S.Am	5.5 2.2	7.4

Laterite Deposits cont....

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.	B.P. PALAT
108	Barro Alto, Brazil	-15.1 -48.9	50 - 0	0 50 S.Am	-15.1 -21.7	 -17.3
109	Liberdade, Minas Gerais, Brazil	-22.0 -44.4	50 - 0	0 50 S.Am	-22.0 -29.1	 -25.4
69	Gramsh, Albania	40.9 20.2	50 - 42	50 Eur	31.8	22.0
68	Jammu, Kashmir, India	32.7 74.9	65 - 45	50 Ind	3.7	-4.6
66	Arkalyk, Kazakhstan, USSR	50.3 66.8	65 - 50	50 Sib	43.7	48.0
67	Sargodha, Pakistan	32.1 72.7	65 - 54	50 Ind	3.6	-4.8
65	Tatarsk, Novosibirsk, USSR	55.2 76.0	65 - 55	50 Sib	49.5	56.5
70	Ebro Massif, Catalonia, Spain	44.7 9.2	65 - 55	50 Eur	35.7	31.1
72	Salzburg, Tirol, Austria	47.9 13.0	91 - 88	100 Eur	30.5	36.4
71	Ariege, Pyrenees, France	43.1 -1.0	98 - 88	100 Eur	27.0	30.8
80	Spinazzola, Apulia, Italy	41.0 16.1	98 - 88	100 Eur	23.5	22.8
89	Turgat Area, USSR	49.6 63.4	113 - 88	100 Sib	34.5	37.6
86	Naurzumsk, Kazakhstan, USSR	51.5 64.5	113 - 91	100 Sib	36.5	40.0

Laterite Deposits cont...

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.	B.P. PALAT
87	Chadobets, USSR	58.7 98.9	113 - 91	100 Sib	49.9	55.7
88	Keul, Angara Area, USSR	58.4 102.7	113 - 91	100 Sib	50.5	56.3
85	Megara, Greece	38.0 23.3	113 - 98	100 Eur	20.2	24.1
77	Salair, Altay-Sayan Region, USSR	54.2 85.9	119 - 97	100 Sib	42.9	46.6
81	Orsk, Urals, USSR	51.2 58.6	119 - 97	100 Eur	35.4	39.1
82	Chelyabinsk, Urals, USSR	55.2 61.4	119 - 97	100 Sib	39.7	44.4
83	Alapayevsk, Urals, USSR	57.9 61.7	119 - 97	100 Sib	42.3	48.1
84	Uralsk, Kazakhstan, USSR	51.3 51.3	119 - 97	100 Eur	34.7	38.9
75	St. Paul de Fenouillet, Pyrenees, France	42.8 2.5	119 - 98	100 Eur	26.3	32.0
76	Nurri, Sardinia	39.7 9.2	119 - 98	100 Eur	22.6	30.1
74	Morella, Maestrazgo, Catalonia, Spain	40.6 -0.1	119 - 113	100 Eur	24.4	28.1
45	Chimkent, Kazakhstan, USSR	42.3 69.1	138 - 131	130 Sib	36.7	24.7
47	Kandahar, Afghanistan	31.0 65.5	144 - 97	100 Ind 130 Ind	-34.5 -63.8	17.6 12.8

Laterite Deposits cont...

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.	B.P. PALAT
73	Kansaysk Area, Tadzhikistan, USSR	40.5 69.7	144 - 97	100 Sib 130 Sib	26.6 35.0	27.7 22.9
46	Vlasenica, Yugoslavia	44.2 19.0	144 - 98	100 Eur 130 Eur	26.5 38.1	31.4 25.7
44	Padurea Craiului, Rumania	44.3 23.8	144 - 125	130 Eur	38.1	25.4
40	Rovinj Vrsar, Istria, Yugoslavia	45.1 13.7	150 - 144	130 Eur	39.2	20.7
41	Crimea, USSR	45.0 33.7	156 - 150	130 Eur	38.6	25.6
42	Atlanti, Helicon Range, Greece	38.6 23.0	163 - 156	130 Eur	32.4	19.4
43	Euboia Island, Greece	38.8 23.5	163 - 156	130 Eur	32.6	19.7
38	Vanoise, France	45.4 6.8	188 - 163	130 Eur 200 Eur	39.9 39.1	28.7 32.7
39	Campellpore, Pakistan	33.8 72.4	188 - 163	130 Ind 200 Ind	-61.8 -31.8	-35.4 -16.2
37	Skopelos Island, Pelagonian Zone, Greece	39.2 23.7	213 - 188	200 Eur	37.5	22.7
35	Kerman, Central Plateau, Iran	30.3 57.1	231 - 213	200 Arab 250 Gond	5.9 -28.0	13.5
36	Yazd, Central Plateau, Iran	31.9 54.4	231 - 213	200 Arab 250 Gond	8.0 -26.5	14.7

Laterite Deposits cont....

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.	B.P. PALAT
34	Albanian Alps	42.4 19.7	238 - 225	200 Eur 250 Ceur	36.3 1.8	21.4
33	Volos, Pelagonian Zone, Greece	39.4 23.7	248 - 213	200 Eur 250 Seur	37.7 0.3	22.9
30	Menderes Massif, Turkey	39.8 26.8	258 - 243	250 Turk	1.8	
22	Chura Gali, NW Frontier, Pakistan	34.3 73.4	258 - 243	250 Gond	-21.3	
23	Bukan, Iran	36.5 46.2	258 - 243	250 Gond	-21.6	
28	Samos, Aegean Islands	37.7 26.9	258 - 243	250 Seur	-0.1	
29	Naxos, Aegean Islands	37.1 25.4	258 - 243	250 Seur	-1.2	
25	Lang Son, Vietnam	21.8 106.7	258 - 248	250 Sund	-15.8	
26	Cao Bang, Vietnam	22.7 106.3	258 - 248	250 Sund	-15.5	
27	Sisophon, Kampuchea	13.6 103.0	258 - 248	250 Sund	-24.1	
31	Alanya,Western Taurus,Turkey	36.5 32.0	258 - 248	250 Turk	0.9	
32	Adama, Eastern Taurus,Turkey	38.6 28.3	258 - 248	250 Turk	1.2	
21	Vienna, Gasconade County, Missouri, USA	38.2 -92.0	333 - 300	300 Laur 350 Laur	-0.4 -11.0	

Laterite Deposits cont...

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE-LAT.
20	Kuei-Yang, Kweichow Province, China	26.6 106.7	333 - 305	300Chi/J 350Chi/J	-17.8 -15.6
19	Tobyssk, Timan Range, USSR	63.2 53.1	352 - 333	350 Eur	27.0
18	Kulkuduk, Bukantau, USSR	42.5 63.3	360 - 300	300 Sib 350 Sib	24.5 -5.0
16	Kunming, Yunnan Province, China	25.1 102.7	360 - 320	300Chi/J 350Chi/j	-21.7 -19.4
17	Hsu-Ch'ang, Gun District, China	34.0 113.8	360 - 320	300Chi/J 350Chi/J	-8.8 -8.1
15	Chitral, Hindukush Range, Pakistan	35.9 72.0	387 - 374	350 Gond 400 Gond	-38.0 -13.3
14	Tselinograd, Kazakhstan, USSR	51.2 71.5	468 - 438	450 Sib	20.3

Phosphate Deposits

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.
773	No.209 Bore (135m),Off-shore Peru	-7.7 -81.2	1 - 0	0	-7.7
774	No.212 Bore (300m),Off-shore Peru	-9.2 -78.8	1 - 0	0	-9.2
775	No.221 Bore(1000m,Off-shore) Peru	-15.1 -75.8	1 - 0	0	-15.1
766	Off-shore, Baja California, Mexico	26.4-113.4	2 - 0	0	26.4
767	Off-shore, Baja California, Mexico	25.8-112.6	2 - 0	0	25.8
768	Boca de las Animas, Baja California, Mexico	25.7-112.1	2 - 0	0	25.7
769	Boca de Solidad, Baja California, Mexico	25.2-112.2	2 - 0	0	25.2
770	Off-shore, Baja California, Mexico	25.6-113.1	2 - 0	0	25.6
771	Off-shore, Baja California, Mexico	25.1-112.7	2 - 0	0	25.1
772	Agulhas Bank, Cape of Good Hope, S. Africa	-35.8 22.5	2 - 0	0	-35.8
728	Ipswich, Suffolk, England	52.1 1.2	5 - 2	0	52.1
729	Notojima Island, Japan	37.2 137.0	11 - 5	0	37.2

Phosphate Deposits cont.....

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.
730	Usetsu Area, Noto Peninsula, Japan	37.4 137.2	11 - 5	0	37.3
747	Off-shore, S. Africa	-30.8 15.5	24 - 2	0	-30.8
748	Off-shore, nr. Saldanha, S. Africa	-32.8 17.5	24 - 2	0	-32.8
749	Off-shore, nr. Cape Town, S. Africa	-34.3 18.0	24 - 2	0	-34.3
750	Off-shore, nr. Agulhas Bank, S. Africa	-35.5 19.0	24 - 2	0	-35.5
751	Off-shore, E. Agulhas Bank S. Africa	-35.7 21.5	24 - 2	0	-35.7
752	Off-shore,W. of Cape Recife, S. Africa	-34.3 23.0	24 - 2	0	-34.3
753	Off-shore,W. of Cape Recife, S. Africa	-34.3 24.0	24 - 2	0	-34.3
776	Willemstad, Table Mountain, Curacao	12.1 -68.7	24 - 2	0	12.1
735	Ocala Hard Rock, Florida,USA	29.2 -82.1	24 - 5	0	29.2
736	Savannah, Georgia, USA	32.1 -81.1	24 - 5	0	32.1
737	St. Augustine, Florida, USA	29.9 -81.3	24 - 5	0	29.9

Phosphate Deposits cont....

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAEO -LAT.
738	Waldo, Florida, USA	29.8 -82.2	24 - 5	0	29.8
739	Macclenny, Florida, USA	30.3 -82.1	24 - 5	0	30.3
740	Thelma, Georgia, USA	30.8 -82.8	24 - 5	0	30.8
741	Melbourne, Florida, USA	28.1 -80.6	24 - 5	0	28.1
742	Fort Ogden, Florida, USA	27.1 -81.9	24 - 5	0	27.1
743	Mulberry, Florida, USA	27.9 -82.0	24 - 5	0	27.9
744	Beaufort, Newington, Georgia, USA	32.6 -81.5	24 - 5	0	32.6
745	Frying Pan, Off-shore, North Carolina, USA	33.8 -77.7	24 - 5	0	33.8
746	Aurora, Pungo R., North Carolina, USA	35.3 -76.8	24 - 5	0	35.3
755	Langebannweg, S. Africa	-32.9 18.1	24 - 5	0	-32.9
756	Bomgat, nr. Saldanha, Hoedjiespunt, S. Africa	-33.0 17.9	24 - 5	0	-33.0
757	Hondeklipbaai, Namaqualand, S. Africa	-30.3 17.3	24 - 5	0	-30.3
758	Ysterplatt, E. of Cape Town, S. Africa	-33.9 18.5	24 - 5	0	-33.9

Phosphate Deposits cont....

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.	B.P. PALAT
759	Minchales, Sechura, Peru	5.9 -80.5	24 - 5	0	5.9	
765	Lee Creek's Mine, North Carolina, USA	35.3 -77.1	22 - 19	0	35.3	
777	S.Hilario (North), Mexico	24.5-111.0	38 - 2	0 50 C.Am	24.5 30.3	 36.3
778	S.Hilario (South), Mexico	24.3-111.2	38 - 2	0 50 C.Am	24.3 30.1	 36.1
779	Western Desert, (Northern Limit), Iraq	33.2 39.0	52 - 50	50 Arab	15.5	13.2
780	Western Desert, (Southern Limit), Iraq	32.3 39.4	52 - 50	50 Arab	14.6	12.3
781	Ga'ara, Iraq	33.7 40.4	73 - 65	50 Arab	15.9	13.5
782	Rutba Deposit, Iraq	33.0 40.3	73 - 65	50 Arab	15.2	12.9
761	Taplow, London, England	51.5 -0.1	88 - 65	50 Eur 100 Eur	42.8 35.1	46.7 42.8
760	Arnager, Southern Bornholm, Denmark	55.0 14.8	97 - 91	100 Eur	37.5	45.1
762	Woburn, Beds, England	52.0 -0.6	119 - 91	100 Eur	35.7	43.5
764	Nr. Cambridge, Cambs,England	52.2 0.1	119 - 91	100 Eur	35.8	43.7
783	Chapopote 1 & 2, Nuevo Leon, Mexico	25.7 -99.7	144 - 88	100 C.Am 130 C.Am	29.2 28.5	29.2 28.9

Phosphate Deposits cont.....

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.	B.P. PALAT
784	Mitra & Arteaga, Nuevo Leon, Mexico	24.8-100.8	243 - 144	130 C.Am 200 C.Am 250 Laur	27.8 25.9 -0.3	28.5 6.8
785	Sierra Gomez-Farias, Carboneras, Mexico	24.9-101.0	243 - 144	130 C.Am 200 C.Am 250 Laur	27.9 26.1 -0.1	28.6 6.9
786	Sierra la Catana, Mexico	25.2-101.3	243 - 144	130 C.Am 200 C.Am 250 Laur	28.2 26.5 0.4	29.0 7.2
787	Unknown, Idaho, USA	43.0-112.0	258 - 248	250 Laur	19.2	
788	Unknown, Oregon, USA	43.0-118.0	258 - 248	250 Laur	22.5	
789	Unknown, Oregon, USA	46.0-117.0	258 - 248	250 Laur	23.8	
790	Unknown, Western Australia Australia	-18.0 125.0	263 - 248	250 Gond	-24.3	
791	Unknown, B.C., Canada	52.0-118.0	263 - 253	250 Laur	27.9	
792	Unknown, Queensland, Australia	-24.0 150.0	263 - 253	250 Gond	-6.7	
793	Unknown, Kiangsi, China	29.0 116.0	286 - 248	250Chi/J 300Chi/J	-4.8 -9.4	
794	Unknown, British Columbia, Canada	57.0-133.0	286 - 248	250 Laur 300 Laur	37.9 32.2	
795	Unknown, Komi, USSR	64.0 58.0	286 - 248	250 Eur 300 Eur	32.0 28.9	

Phosphate Deposits cont.....

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.
796	Unknown, Islamabad, Pakistan	34.0 73.0	286 - 248	250 Gond	-21.7
				300 Gond	-33.1
797	Unknown, Western Australia, Australia	-27.0 116.0	286 - 263	250 Gond	-35.6
				300 Gond	-27.0
798	Unknown, Kampuchea	11.0 104.0	296 - 286	300 Sund	-27.6
799	Unknown, Vietnam	21.0 106.0	296 - 286	300 Sund	-21.2
800	Unknown, Kansas, USA	38.0 -95.0	315 - 286	300 Laur	0.9
801	Unknown, Alberta, Canada	51.0-115.0	315 - 296	300 Laur	20.1
802	Unknown, Texas, USA	32.0 -97.0	320 - 315	300 Laur	-2.6
803	Unknown, Idaho, USA	42.0-112.0	330 - 320	300 Laur	18.9
				350 Laur	-4.5
804	Unknown, Yukon Territory, Canada	68.0-132.0	360 - 320	300 Laur	35.1
				350 Laur	18.8
805	Unknown, Tennessee, USA	36.0 -88.0	360 - 320	300 Laur	-4.1
				350 Laur	-14.8
806	Unknown, Islamabad, Pakistan	34.5 73.0	360 - 352	350 Gond	-38.6
807	Unknown, Islamabad, Pakistan	34.0 73.0	374 - 360	350 Gond	-39.6
808	Unknown, Iran	32.0 56.0	374 - 360	350 Gond	-49.4

Phosphate Deposits cont....

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.
809	Unknown, Bushehr, Iran	28.0 52.0	374 - 360	350 Gond	-54.6
810	Unknown, Tehran, Iran	35.0 52.0	374 - 360	350 Gond	-48.3
811	Unknown, Azarbaijan, Iran	38.0 47.0	374 - 360	350 Gond	-46.8
812	Unknown, Pennsylvania, USA	40.0 -77.0	374 - 360	350 Laur	-19.4
813	Unknown, Kazakhstan, USSR	48.0 75.0	374 - 360	350 Sib	28.8
819	Unknown, Chelyabinsk, USSR	54.0 58.0	387 - 374	350 Eur 400 Eur	1.0 -5.8
814	Unknown, Cilicia, Turkey	37.0 35.0	408 - 360	350 Turk 400 Turk	-33.6 -24.3
815	Unknown, Guipuzcoa, Spain	43.0 -2.0	408 - 360	350 Seur 400 Seur	-28.1 -41.6
817	Unknown, Finnmark, Norway	69.0 25.0	408 - 360	350 Eur 400 Eur	2.0 -11.1
818	Unknown, Kirgiziya, USSR	42.0 76.0	408 - 360	350 Sib 400 Sib	28.2 25.8
820	Unknown, Krasnoyarskiy Kray, USSR	68.0 89.0	408 - 387	400 Sib	32.1
821	Unknown, Pennsylvania, USA	40.0 -77.0	408 - 387	400 Laur	-35.4

Phosphate Deposits cont....

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.
822	Unknown, Islamabad, Pakistan	34.0 73.0	408 - 387	400 Gond	-12.9
823	Unknown, Krasnoyarskiy Kray, USSR	67.0 100.0	408 - 387	400 Sib	36.4
824	Unknown, Chelyabinsk, USSR	54.0 58.0	428 - 414	400 Eur 450 Eur	-5.8 6.8
825	Unknown, Clywd, Wales	53.0 -3.0	438 - 408	400 Ceur 450 Ceur	-35.8 -21.6
826	Unknown, Kiangsi, China	29.0 116.0	438 - 408	400Chi/J 450Chi/J	23.6 -13.5
827	Unknown, Kazakhstan, USSR	43.0 65.0	438 - 408	400 Sib 450 Sib	18.2 20.8
828	Unknown, Amur Oblast, USSR	53.0 129.0	438 - 408	400 Sib 450 Sib	56.2 32.3
724	Llangyog Area, Clywd, Wales	52.8 -3.4	458 - 438	450 Eur	-27.8
725	Dubuque County, Iowa, USA	42.5 -90.8	448 - 438	450 Laur	-37.9
726	Dubuque County, Iowa, USA	42.5 -90.6	448 - 438	450 Laur	-38.3
727	Dubuque County, Iowa, USA	42.3 -90.4	448 - 438	450 Laur	-38.0
829	Unknown, NSW, Australia	-36.0 148.0	458 - 438	450 Gond	48.4
830	Unknown, Wisconsin, USA	43.0 -91.0	458 - 438	450 Laur	-37.5

Phosphate Deposits cont.....

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.
831	Unknown, Kristianstad,Sweden	56.0 14.0	478 - 438	450 Eur	-17.6
832	Unknown, NT, Australia	-23.0 130.0	478 - 458	450 Gond	31.1
833	Unknown, Kazakhstan, USSR	42.0 73.0	478 - 458	450 Sib	26.4
834	Unknown, Krasnoyarskiy Kray, USSR	67.0 92.0	478 - 458	450 Sib	15.8
835	Unknown, Evenkiyskiy Nats. Okrug, USSR	62.0 97.0	478 - 458	450 Sib	21.2
836	Unknown, Krasnoyarskiy Kray, USSR	58.0 107.0	478 - 458	450 Sib	26.7
837	Unknown, Yukutskaya, USSR	61.0 118.0	478 - 458	450 Sib	24.7
838	Unknown, Evenkiyskiy Nats. Okrug, USSR	63.0 92.0	478 - 458	450 Sib	19.2
839	Unknown, Irkutskaya Oblast, USSR	57.0 102.0	478 - 458	450 Sib	26.8
840	Unknown, Taymyrskiy, USSR	74.0 90.0	478 - 458	450 Sib	9.4
841	Unknown, Oklahoma, USA	35.0 -98.0	505 - 438	450 Laur	-36.6
842	Unknown, Szechwan, China	29.0 103.0	505 - 438	450Chi/J	-3.1
843	Unknown, Kentucky, USA	37.0 -86.0	505 - 438	450 Laur	-44.1
844	Unknown, Alabama, USA	33.0 -86.0	505 - 438	450 Laur	-46.4
845	Unknown, Nova Scotia, Canada	44.0 -66.0	505 - 438	450 Laur	-48.6

Phosphate Deposits cont....

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.
846	Unknown, Gwynedd, Wales	53.0 -4.0	505 - 478	450 Ceur	-22.1
847	Unknown, NSW, Australia	-15.0 130.0	505 - 478	450 Gond	28.6
848	Unknown, NT, Australia	-36.0 148.0	505 - 478	450 Gond	48.4
849	Unknown, Puerto Rico	18.0 -67.0	505 - 478	450 Laur	-68.9
850	Unknown, Leningrad, USSR	59.0 29.0	505 - 478	450 Eur	-9.0
851	Unknown, Kirgiziya, USSR	42.0 78.0	505 - 478	450 Sib	29.4

APPENDIX TWO

Porphyry Copper Deposits

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.
2091	Dry Creek, Alaska, USA	56.5-158.4	3.3	0	56.5
2089	Attu, Alaska, USA	52.9-172.8	5.9	0	52.9
2094	Mesatchee, Washington, USA	47.5-121.4	6.2	0	47.5
2090	Pyramid Mt., Alaska, USA	55.6-160.7	6.3	0	55.6
2096	North Fork, Washington, USA	47.6-121.6	9.9	0	47.6
2092	Earl, Washington, USA	46.3-122.1	16	0	46.3
2093	Middle Fork, Washington, USA	47.5-121.4	18	0	47.5
2095	Quartz Creek, Washington,USA	47.7-121.6	18	0	47.6
2097	Glacier Peak, Washington,USA	48.2-120.9	22	0	48.2
2032	McCoy, Washington, USA	46.4-121.8	24	0 50 N.Am	46.4 53.8
2012	Jimmy Lake, Alaska, USA	61.7-153.2	26	0 50 N.Am	61.7 74.9
2038	Ross Lake-Davis, Washington, USA	49.0-121.1	30	50 N.Am	56.0
2037	Vesper, Washington, USA	48.0-121.5	32	50 N.Am	55.2
2050	Bingham, Utah, USA	40.5-112.1	37	50 N.Am	45.8

Porphyry Copper Deposits cont....

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.
2051	Battle Mt., Utah, USA	40.7-117.1	39	50 N.Am	47.3
2031	Catface, Vancouver I.,Canada	49.2-126.0	48	50 N.Am	57.4
2035	Monument, Washington, USA	48.8-120.5	49	50 N.Am	55.7
2028	Berg, B.C., Canada	53.8-127.5	50	50 N.Am	62.0
2030	Granisle, B.C., Canada	55.0-126.3	50	50 N.Am	62.8
2025	Morrison, B.C., Canada	55.0-127.1	52	50 N.Am	63.0
2041	Morenci, Arizona, USA	33.1-109.4	55	50 N.Am	38.1
2040	Tyrone, New Mexico, USA	32.7-108.3	56	50 N.Am	37.6
2013	Jay Creek, Alaska, USA	62.2-153.7	57	0 50 N.Am	62.2 75.5
2042	Cananea, Sonora, Mexico	31.0-110.3	59	50 N.Am	36.3
2009	Dutton, Alaska, USA	60.7-153.9	59	0 50 N.Am	60.7 74.1
2043	Miami-Inspiration, Arizona, USA	33.4-110.9	60	50 N.Am	38.8
2046	Pima-Mission, Arizona, USA	32.9-109.8	60	50 N.Am	38.0

Porphyry Copper Deposits cont....

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.
2026	Maggie, B.C., Canada	50.7-121.3	61	50 N.Am	57.6
2047	Esperanza, Texas, USA	31.1-105.7	62	50 N.Am	35.2
2039	Santa Rita, New Mexico, USA	32.8-108.1	63	50 N.Am	37.5
2044	Ray, Arizona, USA	33.2-111.0	63	50 N.Am	38.6
2045	Silver Bell, Arizona, USA	32.4-111.5	63	50 N.Am	38.0
2052	Ajo, Arizona, USA	32.0-112.6	63	50 N.Am	37.9
2126	Copper Basin, Arizona, USA	34.5-112.9	64	50 N.Am	40.3
2001	Treasure Creek, Alaska, USA	62.9-149.3	65 - 2	0 50 N.Am	62.9 75.3
2002	Maclaren, Alaska, USA	63.2-146.7	65 - 2	0 50 N.Am	63.2 75.0
2003	Kaskawulsh, Alaska, USA	66.5-139.0	65 - 2	0 50 N.Am	66.5 75.8
2004	Rainy Hollow, Alaska, USA	59.6-136.6	65 - 2	0 50 N.Am	59.6 69.5
2005	Great Hog Basin, Alaska, USA	56.5-132.1	65 - 2	0 50 N.Am	56.5 65.6
2007	Unalaska, Alaska, USA	53.9-166.5	65 - 2	0 50 N.Am	53.9 69.1

Porphry Copper Deposits cont....

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAEO -LAT.
2008	Warner Bay, Alaska, USA	56.2-158.4	65 - 2	0 50 N.Am	56.2 70.5
2010	Kijik, Alaska, USA	60.3-154.4	65 - 2	0 50 N.Am	60.3 73.9
2011	Hayes Glacier, Alaska, USA	61.6-152.5	65 - 2	0 50 N.Am	61.6 74.7
2014	Kuskokwim, Alaska, USA	61.4-153.2	65 - 2	0 50 N.Am	61.5 74.8
2015	Ivanof, Alaska, USA	55.9-159.4	65 - 2	0 50 N.Am	55.9 70.4
2033	Fortune, Washington, USA	47.4-121.1	65 - 2	0 50 N.Am	47.4 54.6
2128	San Manuel, Arizona, USA	33.0-110.8	67	50 N.Am	38.4
2049	Butte, Montana, USA	46.0-112.5	69	50 N.Am	51.0
2027	Casino, Yukon, Canada	62.7-138.8	70	50 N.Am	72.7
2036	Mazama, Washington, USA	48.6-120.4	70	50 N.Am	55.5
2048	Bagdad, Arizona, USA	34.6-113.2	71	50 N.Am	40.5
2127	Ithica Peak, Arizona, USA	35.7-114.4	72	50 N.Am	41.9

Porphyry Copper Deposits cont....

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.
2062	Fish Lake, B.C., Canada	50.8-133.1	77	100 N.Am	63.4
2061	Purcell Mt., Alaska, USA	66.2-157.5	80	100 N.Am	83.0
2065	Huckleberry, B.C., Canada	53.8-127.2	80	100 N.Am	63.7
2059	Indian Mt., Alaska, USA	66.1-153.8	81.5	100 N.Am	81.6
2060	Zane Hill, Alaksa, USA	66.3-156.1	81.9	100 N.Am	82.5
2064	Ox Lake, B.C., Canada	53.7-127.0	83	100 N.Am	63.6
2055	Monte Cristo Creek, Alaska, USA	62.2-143.0	109	100 N.Am	75.5
2063	Bond Creek, Alaska, USA	62.0-142.8	109	100 N.Am	75.4
2053	Ely, Nevada, USA	39.2-114.9	111	100 N.Am	46.9
2054	Yerington, Nevada, USA	39.0-119.2	111	100 N.Am	48.3
2056	Horsfeld, Alaska, USA	62.0-141.2	111	100 N.Am	74.7
2057	Ptarmigan Creek, Alaska, USA	62.0-141.0	114	100 N.Am	74.7
2058	Baultoff, Alaska, USA	62.2-141.2	114	100 N.Am	74.8
2137	Sandy Creek, Queensland, Australia	-20.0 148.7	125 - 115	100 Aust 130 Aust	-46.6 -64.5
2138	Finley Creek, Queensland, Australia	-20.1 147.8	125 - 115	100 Aust 130 Aust	-47.1 -64.0

Porphyry Copper Deposits cont...

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.
2141	Roma Peak, Queensland, Australia	-20.3 148.2	125 - 115	100 Aust 130 Aust	-47.0 -64.4
2144	Andromache R., Queensland, Australia	-20.6 148.4	125 - 115	100 Aust 130 Aust	-47.2 -64.8
2145	Pentecost I., Queensland, Australia	-20.4 149.0	125 - 115	100 Aust 130 Aust	-46.7 -64.9
2129	Mount Vista, Queensland, Australia	-20.5 147.9	125	130 Aust	-64.5
2130	Mount Poole, Queensland, Australia	-20.7 147.9	125	130 Aust	-64.6
2131	Mount Leslie, Queensland, Australia	-20.9 147.9	125	130 Aust	-64.8
2132	Blenheim, Queensland, Australia	-21.1 148.2	125	130 Aust	-65.1
2133	Eungella, Queensland, Australia	-21.2 148.5	125	130 Aust	-66.1
2134	Mount Gotthart, Queensland, Australia	-21.4 148.3	125	130 Aust	-65.5
2135	Mount Hess, Queensland, Australia	-21.6 148.4	125	130 Aust	-66.0
2136	Mount Flora, Queensland, Australia	-22.0 148.5	125	130 Aust	-64.5

Porphyry Copper Deposits cont....

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.
2067	Brenda, B.C., Canada	49.9-120.0	140	130 N.Am	55.3
2006	Costello Creek, Alaska, USA	63.2-149.5	144- 65	50 N.Am 100 N.Am 130 N.Am	75.6 78.5 71.9
2017	Co, Alaska, USA	62.7-138.5	144 - 65	50 N.Am 100 N.Am 130 N.Am	72.6 74.2 70.2
2018	Cockfield, Alaska, USA	62.5-138.5	144 - 65	50 N.Am 100 N.Am 130 N.Am	72.4 74.1 70.1
2019	Dennis, Alaska, USA	63.4-142.4	144 - 65	50 N.Am 100 N.Am 130 N.Am	74.1 76.0 71.4
2020	Taurus, Alaska, USA	63.5-141.3	144 - 65	50 N.Am 100 N.Am 130 N.Am	74.0 75.7 71.3
2022	Mt. Nansen, Alaska, USA	62.0-137.1	144 - 65	50 N.Am 100 N.Am 130 N.Am	71.7 73.2 69.4
2023	West Cape, St.Lawrence Island, Alaska, USA	63.5-171.5	144 - 65	50 N.Am 100 N.Am 130 N.Am	78.9 84.9 73.2
2024	Granite Mt., Alaska, USA	65.3-161.4	144 -65	50 N.Am 100 N.Am 130 N.Am	79.7 83.9 74.8

Porphyry Copper Deposits cont....

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.
2066	Island Copper, Vancouver I, Canada	59.6-127.5	153	130 N.Am	65.8
2068	Bisbee, Arizona, USA	31.4-109.9	163	130 N.Am	35.7
2075	Duckling Creek, B.C., Canada	55.9-125.4	170	200 N.Am	45.2
2070	Liard, B.C., Canada	57.3-130.8	182	200 N.Am	47.9
2074	Galore Creek, B.C., Canada	57.1-131.4	182	200 N.Am	47.3
2076	Copper Mt., B.C., Canada	49.3-121.6	193	200 N.Am	38.3
2077	Afton, B.C., Canada	50.6-120.5	198	200 N.Am	39.4
2071	Guichon Batholith, B.C., Canada	50.5-121.0	200	200 N.Am	39.4
2073	Cuddy Mt., Idaho, USA	45.2-116.2	200	200 N.Am	33.5
2072	Gibraltar, B.C., USA	52.5-122.3	204	200 N.Am	41.5
2146	Waitara, Queensland, Australia	-21.8 148.8	210?	200 Aust	2.5
2147	Funnel Creek, Queensland, Australia	-21.7 149.1	210?	200 Aust	2.8
2148	Green Hill, Queensland, Australia	-21.7 149.4	210?	200 Aust	2.9
2149	Knight Island, Queensland, Australia	-21.4 149.7	210?	200 Aust	3.3

Porphyry Copper Deposits cont....

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.
2150	Yeppoon, Queensland, Australia	-23.0 150.8	220	200 Aust	2.5
2151	Native Dog, Queensland, Australia	-24.7 151.9	220	200 Aust	1.5
2152	Enoggera, Queensland, Australia	-27.4 152.8	230	250 Gond	-6.4
2153	Mount Crosby, Queensland, Australia	-27.6 152.7	230	250 Gond	-6.5
2154	Bald Mountain, Queensland, Australia	-28.3 152.0	230	250 Gond	-7.5
2155	Cania, Queensland, Australia	-24.6 151.0	235	250 Gond	-6.3
2158	Anduramba, Queensland, Australia	-27.1 152.1	240?	250 Gond	-6.8
2159	Moonmera, Queensland, Australia	-23.6 150.6	245	250 Gond	-6.1
2069	McClean Arm, Alaska, USA	54.8-130.2	248 - 213	200 N.Am 250 Laur	44.9 35.6
2161	Riverhead, Queensland, Australia	-24.0 150.9	250 - 235	250 Gond	-6.0
2164	Munholme Creek, Queensland, Australia	-24.5 151.1	250 - 235	250 Gond	-6.1
2165	Ridler Creek, Queensland, Australia	-24.4 151.3	250 - 235	250 Gond	-6.0

Porphyry Copper Deposits cont....

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE-LAT.
2174	Coalstoun, Queensland, Australia	-25.6 151.9	258 - 248	250 Gond	-6.1
2175	Mount Stuart, Queensland, Australia	-19.4 146.8	265	250 Gond	-6.9
2176	Kelly's Mountain, Queensland Australia	-19.7 147.3	265	250 Gond	-6.6
2177	Beak's Mountain, Queensland, Australia	-19.9 147.7	265	250 Gond	-6.4
2179	Rocky Creek, Queensland, Australia	-20.4 147.9	265	250 Gond	-6.5
2180	Town Creek, Queensland, Australia	-20.4 146.8	285	300 Gond	1.9
2181	Mount Robin, Queensland, Australia	-20.6 147.0	285	300 Gond	2.0
2182	Mount Wyatt, Queensland, Australia	-20.8 147.3	285	300 Gond	2.2
2183	Wyarra Hills, Queensland, Australia	-21.1 147.6	285	300 Gond	2.4
2184	Mount Lookout, Queensland, Australia	-21.4 147.7	285	300 Gond	2.3
2185	Mountian Maid, Queensland, Australia	-17.0 144.2	285	300 Gond	0.7

Porphyry Copper Deposits cont....

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAE -LAT.
2186	Ruddygore, Queensland, Australia	-17.2 144.6	285	300 Gond	1.0
2187	Eureka Creek, Queensland, Australia	-17.2 145.0	285	300 Gond	1.3
2188	Carbonate Creek, Queensland, Australia	-17.4 145.1	285	300 Gond	1.4
2189	Koombooloomba, Queensland, Australia	-17.8 145.6	285	300 Gond	1.7
2190	Nitchaga, Queensland, Australia	-18.0 145.5	285	300 Gond	1.5
2192	Yuccabine, Queensland, Australia	-18.3 145.7	285	300 Gond	1.6
2194	Mount Darcy, Queensland, Australia	-18.3 143.3	310	300 Gond	-0.6
2195	Mount Turner, Queensland, Australia	-18.5 143.5	310	300 Gond	-0.4
2081	Gaspe, Quebec, Canada	49.0 -65.5	350	350 Laur	-18.2
2080	Evandale, New Brunswick, Canada	45.6 -66.1	364	350 Laur	-20.7
2084	Chandalar, Alaska, USA	67.5 -48.2	370	350 Laur	-5.9
2196	Titov, Queensland, Australia	-18.0 146.4	394	400 Gond	51.3

Porphyry Copper Deposits cont....

A.NO.	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAEO -LAT.
2197	Turkey Gully, Queensland, Australia	-20.0 146.4	394	400 Gond	50.3
2199	Keans, Queensland, Australia	-20.1 146.5	394	400 Gond	50.4
2200	Cameron, Queensland, Australia	-20.2 146.3	394	400 Gond	50.2
2078	Eagle Lake, New Brunswick, Canada	45.2 -66.4	408 - 360	350 Laur 400 Laur	-20.9 -36.4
2082	Recontre East, Newfoundland, Canada	47.5 -55.2	408 - 360	350 Laur 400 Laur	-23.1 -37.6
2083	Woodstock, New Brunswick, Canada	46.0 -67.6	408 - 360	350 Laur 400 Laur	-19.8 -35.3
2085	Catheart, Maine, USA	45.6 -70.2	457	450 Laur	-45.8
2086	Sally Mountain, Maine, USA	45.6 -70.3	505 - 438	450 Laur	-45.7
2201	El Teniente, Chile	-34.2 -70.8	11 - 2	0	-34.2
2202	Los Pelambres, Chile	-32.1 -70.9	11 - 2	0	-32.1
2203	Michiquillay, Peru	-7.0 -78.8	21	0	-7.0
2204	Hualgayoc, Peru	-6.8 -79.0	24 - 5	0	-6.8
2205	Morococha, Peru	-11.5 -76.5	24 - 5	0	-11.5

Porphyry Copper Deposits cont.....

A.NO.	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAEO -LAT.
2206	Farallon Negro-Mivida dist., Argentina	-27.5 -67.1	24 - 5	0	-27.5
2207	Paramillos dist.,Argentina	-32.8 -69.8	24 - 5	0	-32.8
2208	Potrerillos, Chile	-26.2 -69.7	42 - 24	50 S.Am	-29.8
2209	El Salvador, Chile	-26.2 -69.9	42 - 24	50 S.Am	-29.7
2210	Chuquicamata, Chile	-22.2 -69.4	42 - 24	50 S.Am	-25.8
2211	El Abra, Chile	-22.0 -69.0	42 - 24	50 S.Am	-25.7
2212	Toquepala, Peru	-17.2 -70.9	65 - 54	50 S.Am	-20.7
2213	Cuajone, Peru	-17.0 -71.0	65 - 54	50 S.Am	-20.4
2214	Quellavero, Peru	-17.1 -71.0	65 - 54	50 S.Am	-20.5
2215	Cerro Verde, Peru	-16.4 -71.9	65 - 54	50 S.Am	-19.7
2216	Panguna, Bougainville, PNG	-6.0 155.0	c.65	50 Sund	-33.5
2217	Plesyumi, New Britain, PNG	-6.0 155.0	c.65	50 Sund	-31.5
2218	Mamut dep., Sabah, Malaysia	6.0 117.0	c.65	50 Sund	-2.0
2219	Taysan dep., Luzon I., Phillipines	16.5 121.5	c.65	50 Sund	3.6

Porphyry Copper Deposits cont

A.NO.	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAEO -LAT.
2220	Sipalay, Negros, Phillipines	10.0 123.0	c.65	50 Sund	-2.5
2221	Tapadaa, Sulawesi, Indonesia	-2.0 120.0	c.65	50 Sund	-10.1
2222	Almalyk deposit, Kazakhstan, USSR	42.3 68.5	296 - 286	300 Sib	26.6
2223	Kounrad deposit, Kazakhstan, USSR	47.3 75.0	296 - 286	300 Sib	33.4

Epithermal Gold Deposits

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAEO -LAT.
2119	Florida Canyon, Nevada, USA	40.6-118.2	1 - 0	0	40.6
2120	Comstock Lode, Nevada, USA	39.0-119.6	12	0	39.0
2121	Goldfield, Nevada, USA	37.7-117.2	20 - 5	0	37.7
2117	Tonopah-Divide, Nevada, USA	38.1-117.2	20 - 15	0	38.1
2118	Summitville, Colorado, USA	37.4-106.6	21	0	37.4
2098	Cinola, Queen Charlotte I., Canada	53.3-132.5	24 - 5	0 50 N.Am	53.3 62.8
2103	Delamar, Idaho, USA	43.0-116.8	24 - 15	0 50 N.Am	43.0 49.4
2099	Round Mountain, Nevada, USA	38.7-117.1	25	0 50 N.Am	38.7 45.4
2123	Relief Canyon, Nevada, USA	40.0-119.8	30	0 50 N.Am	40.0 47.3
2104	Alligator Ridge, Nevada, USA	39.8-115.8	30 - 5	0 50 N.Am	39.8 46.1
2105	Jerritt Canyon, Nevada, USA	41.0-115.8	65	0 50 N.Am	41.0 47.2
2100	Picacho, California, USA	33.0-114.6	65 - 2	0 50 N.Am	33.0 39.3
2101	Buckhorn, Nevada, USA	40.1-116.6	65 - 2	0 50 N.Am	40.1 46.6

Epithermal Gold Deposits cont.....

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAEO -LAT.
2102	Borealis, Nevada, USA	38.5-118.6	65 - 2	0 50 N.Am	38.5 45.6
2124	Golden Sunlight, Montana, USA	46.0-112.5	65 - 5	0 50 N.Am	46.0 51.0
2106	Taylor, Nevada, USA	39.0-115.5	70 -65	0 50 N.Am	39.0 45.3
2108	Pinson, Nevada, USA	41.0-117.7	85 - 20	50 N.Am 100 N.Am	47.7 49.5
2125	Windfall, Nevada, USA	39.5-116.0	85 - 20	0 50 N.Am 100 N.Am	39.5 45.9 47.6
2107	Nenzel Hill, Nevada, USA	40.3-118.1	86 - 58	50 N.Am 100 N.Am	47.1 49.0
2122	Hasbrouck, Nevada, USA	37.9-117.8	100 - 2	0 50 N.Am 100 N.Am	37.9 44.8 46.8
2109	Quesnel's R., BC, Canada	53.0-122.5	231 - 188	200 N.Am 250 Laur	40.8 30.7
2110	Cracow, Queensland, Aust.	-25.3 150.3	258 - 231	250 Gond	-7.2
2112	Drake, NSW, Australia	-28.9 152.4	260 - 240	250 Gond	-7.5
2111	Mt. Rawdon, Queensland, Australia	-25.2 151.5	286 - 248	250 Gond 300 Gond	-6.2 4.3

Epithermal Gold Deposits cont...

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAEO -LAT.
2113	Yalwal, NSW, Australia	-35.3 150.4	380 - 360	350 Gond 400 Gond	16.9 43.5
2114	Pambula, NSW, Australia	-37.0 150.0	380 - 360	350 Gond 400 Gond	16.1 42.1
2115	Temora, NSW, Australia	-34.4 147.5	420	400 Gond	42. 5
2116	Peak Hill, NSW, Australia	-32.8 148.2	438 - 400	400 Gond 450 Gond	44.0 48.6
2223	Chinkuashih, Taiwan	24 0 121.0	1	0	24.0
2224	Pachuca Ag deposit, Mexico	19.8 -98.8	3	0	19.8
2225	Baguio, Luzon, Phillipines	17.0 121.0	5	0	17.0
2226	Tui mine, Hauraki, N. Z.	-37.0 175.8	7 - 2.5	0	-37.0
2227	Vatukoula, Viti Levu, Fiji	-17.5-177.8	5 - 4	0	-17.5
2228	Huachocolpa Dist., Peru	-11.2 -77.6	8 - 4	0	-11.2
2229	Lepanto Mine, Bontoc, Luzon	16.9 12.8	11 - 3	0	16.9
2230	Finlandia vein, Lima, Peru	-12.0 -77.0	10	0	-12.0
2231	Monte Cristo, El Salvador	13.7 -88.3	24 - 2	0	13.7
2232	San Sebastian, El Salvador	13.8 -87.7	24 - 2	0	13.8

Epithermal Gold Deposits cont......

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAEO -LAT.
2233	Limon, Costa Rica	12.8 -86.8	24 - 2	0	12.8
2234	La India, Costa Rica	12.8 -86.2	24 - 2	0	12.8
2235	Santo Domingo, Nicaragua	12.2 -85.1	24 - 2	0	12.2
2236	Rodalquilar area, Spain	36.9 -2.5	24 - 5	0	36.9
2237	Sanru dep., Hokkaido, Japan	44.2 141.8	24 - 5	0	44.2
2238	Date, Hokkaido, Japan	42.4 140.5	24 - 5	0	42.4
2239	Mutsu, Honshu, Japan	40.5 140.1	24 - 5	0	40.5
2240	Kawazu, Honshu, Japan	34.7 138.8	24 - 5	0	34.7
2241	Bajo, Kyushu, Japan	33.5 131.5	24 - 5	0	33.5
2242	Guanajuato Ag dep., Mexico	20.5-101.3	30 - 27	50 C.Am	23.9
2243	Angangueo Ag dep., Mexico	19.2-100.0	30 - 27	50 C.Am	22.3
2244	Cosala Ag dep., Mexico	23.8-106.9	30	50 C.Am	28.7
2245	Tayoltita Ag dep., Mexico	23.5-106.0	30	50 C.Am	28.0
2246	Acari district, Peru	-15.2 -74.7	144	130 S.Am	-18.1

Epithermal Gold Deposits cont.....

A.NO	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAEO -LAT.
2247	Tocopilla district, Chile	-22.1 -70.5	144	130 S.Am	-24.6
2248	Carrizal Alto, Chile	-28.3 -71.2	144	130 S.Am	-30.8
2249	Huantajaya, Chile	-20.2 -70.3	144	130 S.Am	-22.7
2250	Arqueros, Chile	-30.0 -71.5	144	130 S.Am	-32.5
2251	Punta del Cobre dist., Chile	-27.5 -70.5	144	130 S.Am	-29.9
2252	La Africana, Chile	-33.8 -71.3	144	130 S.Am	-36.3
2253	Tres Puntas dist., Argentina	-47.0 -66.0	144	130 S.Am	-48.9

ADDENDUM

The examiners advised inclusion of the following examples omitted from the thesis when originally submitted. The material in no way detracts from the thesis and only adds to conclusions previously stated. This completes the database.

Additional References for Oolitic Ironstone Examples.

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Oolitic Ironstone Deposits (continued).

A.NO.	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAEO -LAT.
3001	Achtal-Kressenberg area, nr. Traunstein, FRG	47.9 12.7	50 - 45	50 Eur	38.8
3002	Hollfeld, Oberpfalz, FRG	49.5 12.4	97 - 65	50 Eur 100 Eur	40.4 32.1
3003	Peine, Harz Foreland, FRG	52.3 10.3	97 - 65	50 Eur 100 Eur	43.3 35.0
3004	Meckelfeld, nr. Hamburg, FRG	53.5 10.0	144 - 97	100 Eur 130 Eur	36.3 47.8
3005	Salzgitter, FRG	52.1 10.4	144 - 125	100 Eur 130 Eur	34.8 46.4
3006	Hansa, Harz Foreland, FRG	51.9 10.5	163 - 144	130 Eur	46.2
3007	nr. Minden, FRG	52.3 8.9	169 - 163	130 Eur	46.6
3008	Salzgitter, FRG	52.1 10.4	175 - 144	130 Eur 200 Eur	46.4 46.2
3009	Gifhorn, FRG	52.5 10.5	175 - 144	130 Eur 200 Eur	46.8 46.6
3010	Nr. Metz, France	49.1 6.2	188 - 144	130 Eur 200 Eur	43.6 42.5
3011	Nr. Belfort, France	47.6 6.8	188 - 144	130 Eur 200 Eur	42.1 41.2
3012	Nr. Ehingen, FRG	48.3 9.7	188 - 144	130 Eur 200 Eur	42.6 42.5

Oolitic Ironstone Deposits cont.....

A.NO.	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS		AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAEO -LAT.
3013	Staffhorst, nr. Nienburg, FRG	52.6	9.2	188 - 163	200 Eur	46.4
3014	Nr. Freiburg, FRG	50.9	13.3	188 - 181	200 Eur	45.8
3015	Aalen-Wasseralfingen, FRG	48.8	10.1	188 - 181	20 Eur	43.1
3016	Geislingen, FRG	48.6	9.8	188 - 181	200 Eur	42.8
3017	Hersbruck, nr.Nuremberg, FRG	49.5	11.4	188 - 181	200 Eur	44.0
3018	Nordlinger Ries, FRG	48.8	10..5	188 - 181	200 Eur	43.2
3019	Toul, France	48.7	5.9	194 - 188	200 Eur	42.0
3020	Vitry-le-Francois, France	48.7	4.5	194 - 188	200 Eur	41.7
3021	Verdun, Meuse, France	49.2	5.4	194 - 188	200 Eur	42.4
3022	Orne, France	48.7	0.0	194 - 188	200 Eur	40.7
3023	Langres, France	47.9	5.3	194 - 188	200 Eur	41.1
3024	Keilberg, FRG	49.0	12.1	200 - 194	200 Eur	43.7
3025	Echte, Harz Mts, FRG	51.1	5.9	200 - 194	200 Eur	44.3
3026	Lenglern, FRG	51.5	9.9	200 - 194	200 Eur	45.5

Oolitic Ironstone Deposits cont.....

A.NO.	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAEO -LAT.
3027	Friederike, Harz Mts, FRG	51.9 10.5	206 - 200	200 Eur	46.1
3028	Balingen, Wurttemberg, FRG	48.7 9.0	213 - 206	200 Eur	42.7
3029	Fourmies, Ardennes, France	50.0 4.0	387 - 380	400 Eur	-32.6
3030	Dielette, France	49.5 -1.9	394 - 387	400 Eur	-34.3
3031	St. Brieuc, France	48.5 -2.8	401 - 394	400 Eur	-35.4
3032	Herkimer, N.Y., USA	43.0 -75.0	438 - 421	400 Laur 450 Laur	-34.3 -46.0
3033	Lockport, N.Y., USA	43.2 -78.7	438 - 421	400 Laur 450 Laur	-32.3 -44.2
3034	Rochester, N.Y., USA	43.2 -77.7	438 - 421	400 Laur 450 Laur	-32.8 -44.6
3035	Blue Mountain, Penn., USA	45.2-119.0	438 - 421	400 Laur 450 Laur	-35.9 -18.3
3036	Bloomsburg, Penn., USA	41.0 -76.5	438 - 421	400 Laur 450 Laur	-35.0 -46.9
3037	Cumberland, Maryland, USA	39.7 -78.7	438 - 421	400 Laur 450 Laur	-34.7 -46.7
3038	Bluefield, W.Virginia, USA	37.3 -81.1	438 - 421	400 Laur 450 Laur	-34.8 -47.0

Oolitic Ironstone Deposits cont.....

A.NO.	DEPOSIT/MINE NAME AND LOCATION	PRESENT CO-ORDS	AGE MINZ'N (M.Y.)	ROTATION (M.Y.)	PALAEO -LAT.
3039	Messac, Anjou, France	47.8 -1.8	450	450 Eur	-30.1
3040	Le Bois dep., Anjou, France	47.7 -0.9	450	450 Eur	-29.7
3041	La Ferriere, Normandy, France	48.6 -0.7	450	450 Eur	-29.1
3042	Estrees, Normandy, France	48.9 -0.2	450	450 Eur	-28.6
3043	St. Andre, Normandy, France	49.7 1.0	450	450 Eur	-27.5

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