Apparent Polar Wandering with Respect to North America Since the Late Precambrian

Thesis by

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ABSTRACT

To achieve a better understanding of the relative significance of plate tectonics and true polar wandering, it is important that paleomagnetic poles be as accurate as possible and that knowledge of apparent polar wandering be extended to increasingly remote times. In Chapter 1, advantages of using the mode in analysis of paleomagnetic vectors are discussed, and a computer technique is described for contouring and precisely locating the modes of vector distributions that may be highly skewed. In contrast to conventional determinations of the mode, unit vectors from a given data set are treated not as discrete points, but as identical Fisherian probability density functions defined (at an angle θ from the unit vector) by: $p=\exp(sk(\cos\theta-1))$, where k is the estimate of the Fisherian concentration parameter, and s is an arbitrarily assigned "smoothing parameter." A grid, representing the cumulative probability distribution of the total sample of vectors, is contoured to provide a graphical display of the distribution around the most probable value, the mode. By repeatedly contouring the same sample of vectors with successively larger values of s, and by treating the mode as a vector with length given by the total probability value at the mode, "progressive modal diagrams" can be constructed, to aid in determining the stable position of the mode of skewed distributions. In addition, a new statistic, "β_{95}" is suggested as an error estimator for the mode. The statistic β_{95} is derived from the largest subset of the total sample that has a mean identical with the mode of the total sample.

This statistic is defined as the Fisherian half-angle of the cone of 95% confidence for the mean of this subset.

In Chapters 2 and 3, results are presented of a paleomagnetic investigation of upper Precambrian through Middle Cambrian sedimentary rocks from the Cordilleran geosyncline. Over 800 oriented samples were obtained from a homoclinal, conformable sequence of terrigenous and carbonate miogeoclinal strata in the Desert Range, Nevada. Thermal demagnetization isolated similar characteristic magnetization directions in red-purple mudstones of the Wood Canyon Formation (Lower(?) and Lower Cambrian) and in gray limestones of the Carrara and Bonanza King Formations (Lower and Middle Cambrian). Lithologic and magnetic evidence suggest these magnetizations were acquired penecontemporaneously with deposition. The similarity of the characteristic magnetizations in these strata implies that little apparent polar wander occurred with respect to North America from early Early through middle Middle Cambrian time. The divergence of these directions from those from the partly coeval Tapeats Sandstone of the Colorado Plateau probably resulted from a net 36° clockwise rotation of the Desert Range section about a vertical axis. This rotation is probably due to mid-Tertiary oroflexural bending, but may in part have been caused by Mesozoic thrusting. The characteristic magnetization of the uppermost Johnnie Formation (upper Precambrian; ∿650 m.y.) has two polarities, which allows its direction to be established despite incomplete removal of a secondary component of recent origin. The pole from the uppermost Johnnie is about 47° from the pole for the Wood Canyon Formation. It seems probable that at least 45° of apparent polar wandering occurred with respect to North America between about 650 m.y.B.P. and Early Cambrian time.

In Chapter 4, a method for constructing apparent polar wander (APW) paths is presented that includes: (1) grouping data into standard time intervals of about 22 m.y. duration; (2) using new criteria for selection of paleomagnetic poles to be given unit weight; and (3) using the mode to represent the estimated pole position for each time interval. This method is applied to paleomagnetic data from about 250 references to revise the Phanerozoic APW path for North America. The revised path is documented with an interval-by-interval review of all reliable North American paleomagnetic poles for the Phanerozoic. Unlike the smooth and simple APW paths first obtained in the late 1950's, the revised Phanerozoic APW path for North America is characterized by frequent, abrupt changes in the direction and rate of apparent polar wandering. Some anomalous but apparently reliable poles that lie off the revised path may be attributable to (1) complex tectonic histories of parts of the present North American plate, and (2) possibly, complex structure of the geomagnetic field for periods of time that may have been as long as 10^5 to 10^6 years.

In Chapter 5, a speculative synthesis is presented of the revised Phanerozoic APW path for North America and an inferred north polar wandering path for Mars. It has been proposed that quasi-circular features in the Martian polar regions might represent margins of nearly circular caps that had formed symmetrically about the poles. These features would then in effect be "fossil latitudinal circles," and the offsets in their centers of curvature from the present geographic poles might provide evidence for polar wandering on Mars. If this interpretation is correct, then a polar wander path for Mars can be obtained by connecting the centers of curvature of successive margins of the polar caps. A polar wander path was derived by fitting the most prominent quasi-circular features in the north polar region to portions of circles of nearly constant radius ($\sim 3.4^{o}$) and then connecting the centers of curvature in sequence. An interesting characteristic of the Martian north polar wander path so obtained is its tendency to exhibit two major changes in direction during a time interval estimated as between 50 and 500 m.y. These two bends in the Martian north polar wander path are reminiscent of the "hairpin" bends that occur about every 100 to 130 m.y. in the Phanerozoic APW path for North America. In fact, a comparison between apparent polar wandering with respect to North America and the inferred true polar wandering on Mars reveals a remarkable similarity: both processes can be modeled as the superposition of a quasi-periodic component and a secular component. On both Earth and Mars, the quasi-periodic component is elliptical and has a period in the 10^{8}-year range; and on both planets, the secular component propagates at a nearly constant speed in a direction nearly perpendicular to the major axis of the oscillatory component. Moreover, the length of the secular shift per cycle is roughly the same as the length of the major axis of the oscillatory component. These unexpected similarities suggest that a highly ordered form of true polar wandering occurs on Earth and Mars as a consequence of a physical process common to both planets. An oscillatory form of mantle convection is tentatively proposed to be the excitation function of quasi-periodic polar wandering on Earth and Mars.

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INTRODUCTION

Paleontologic and paleoclimatic studies have long been thought to suggest that, over geologic timescales, the earth's rotation axis appears to have shifted with respect to the land surface. The advent of the plate tectonics hypothesis provided one mechanism by which this apparent polar wandering (APW) can result from movement of segments of the earth's crust relative to the mantle. Paleomagnetism, which figured prominently in the growing acceptance of the plate tectonics hypothesis, allowed the first quantitative determinations of paleolatitude, thereby putting constraints on paleogeographic reconstructions. By the early 1960's, it had come to be realized that the position of paleomagnetic poles from most continents had shifted over 90° since Cambrian time (ca. 600 to 500 m.y.B.P.). The full significance of this apparent polar motion is still the subject of considerable controversy. Early in the debate, it had been suggested that true polar wandering, or coherent motion of the solid body of the earth with respect to the astronomically-defined spin axis, accounts for much of the apparent secular shift of the pole, whereas plate tectonics, or differential motion of segments of the earth's crust, accounts for the offset of this shift along different lines of longitude as seen from different plates. Although considerable effort has been expended on reconstructing APW paths, the relative contribution of plate tectonics versus true polar wandering remains largely unknown.

To achieve a better understanding of these two geophysical processes, it is important that paleomagnetic poles be as accurate as possible and that knowledge of apparent polar wandering be extended to increasingly remote times. In Chapter 1, a computer technique is described for precisely locating the modes of samples of paleomagnetic vectors. The use of this technique in many cases permits a more accurate analysis of the characteristic direction of a sample of vectors than use of the traditional vector mean. In Chapters 2 and 3, a paleomagnetic investigation of upper Precambrian through Middle Cambrian sedimentary rocks from the Desert Range, Nevada, is described that extends knowledge of apparent polar wandering with respect to North America to a time for which very little paleomagnetic data have previously been obtained. In Chapter 4, a method for constructing APW paths is presented and applied to revise the North American APW path for the Phanerozoic. Speculations on the planetological significance of the revised path are discussed in Chapter 5.

Chapter 1

ANALYSIS OF THE MODES OF DIRECTIONAL DATA WITH PARTICULAR REFERENCE TO PALEOMAGNETISM

Accurate determination of paleomagnetic directions and poles is becoming increasingly important in regional and global tectonic syntheses and in stratigraphic correlation and age determination. Although the usual procedure for calculating these directions and poles has been to use statistical methods developed by Fisher (1953) and refined by Watson (1956a, 1956b, 1960, 1961, 1962, 1965) and Watson and Irving (1957), these methods assume a particular model of dispersion on a sphere roughly analogous to a two-dimensional, Gaussian (normal) distribution. Commonly, however, paleomagneticists are confronted by distributions of vectors that are not symmetric about the mean. In these instances, computed vector means and the other Fisherian statistics are of questionable geophysical significance. The mode, on the other hand, is a more characteristic measure of central tendency, especially for skewed distributions. In this chapter, some advantages of using the mode in paleomagnetism are discussed and a computer program is described that maps the probability distributions of sets of directional data and precisely locates their modes.

Advantages of Using the Mode

Fisherian parametric statistics are based upon populations of vectors distributed in direction with probability density:

$$P = (\kappa/4\pi\sinh\kappa)\exp(\kappa\cos\theta) \tag{1.1}$$

where θ is the angle between a given direction and the mean, and κ is a constant called the "precision" or "concentration" parameter, describing the dispersion about the mean. Large values of κ indicate small dispersion, and $\kappa = 0$ indicates a uniform distribution over the sphere. The probability distribution given by (1.1), generally called a Fisherian distribution, is symmetric about the mean.

Watson and Irving (1957) have described a χ^2 test for the goodness of fit of a sample of vectors to the Fisherian distribution. They applied this test to two samples containing 70 and 77 paleomagnetic directions from the Torridonian Sandstone and to a sample containing 30 mean site directions from the Tasmanian dolerite sills. The stably magnetized units yielded vector distributions that were found to be consistent with the Fisherian distribution at the 95% confidence level. However, Baag and Helsley (1974) discussed the problems in paleomagnetic applications of the χ^2 test, especially regarding expected frequencies, number of classes, minimum sample size, and number of degrees of freedom. As others have found, they showed that the outcome of the χ^2 test is dependent on the choice of class boundaries and can give inconsistent results when repeatedly applied to the same data set.

A serious disadvantage of the χ^2 test is that it requires a sample size larger than $N = 30$ (Baag and Helsley, 1974) to obtain any useful test of the distribution of the data. In making paleogeographic reconstructions, however, paleomagneticists compare mean poles representing approximately the same interval of geologic time from several lithospheric plates; yet, these mean poles, which are taken to represent points on apparent polar wander paths of individual plates, are commonly based on $N < 10$ (e.g., McElhinny, 1973; Van der Voo and French, 1974). For these small data sets, the assumption of the Fisherian distribution is not readily subject to test. Moreover, estimates of the Fisherian statistical parameters, especially of κ, are probably unreliable for N less than about 7 (Tarling, 1971). When treating these small samples, even a few wildly deviant vectors (outliers) will significantly influence the mean and its confidence estimates, since the mean has the property that it is affected by each vector in a data set. Outliers in distributions of paleomagnetic vectors can be caused by: (1) magnetization during polarity transitions or short-period excursions of the geomagnetic field; (2) the presence of unresolved multiple components of magnetization; (3) isothermal remanent magnetization resulting from lightning strikes; (4) misorientation of specimens either in the field or laboratory; and (5) post-magnetization tectonic rotations, often of unknown amplitude. Treatment of outliers in paleomagnetic data is a difficult problem that has been approached in the past by somewhat subjective methods (e.g., Watkins, 1969; Helsley and Steiner, 1974; Stone and Packer, 1976).

In addition to difficulties in testing the form of the distribution and in applying Fisherian statistics to small samples of data, actual distributions of paleomagnetic vectors are frequently skewed or "streaked." This tendency toward skewness may result from apparent polar wandering during the time interval being investigated (Beck, 1972) or, more commonly, from incomplete removal of secondary components of magnetization (Helsley, 1973). In this latter case, the mean for a given sample of vectors will always be biased by the unremoved secondary component. Indeed, Watson (1966) has cautioned against the application of Fisherian statistical procedures to non-Fisherian distributions:

> If the data indicates several modes or shows marked skewness, then the case for using the [Fisherian] distribution to represent it is slight and the geologist should be content to show his data graphically. It may often be the case that the deviations from a symmetrical unimodal distribution indicate some complicating phenomena--nonrandom sampling, a mixture of populations, etc.--of either conceptual or technical interest.

Unlike the mean, the mode is not affected by outliers and has the property that, for a truly Fisherian distribution, it is identical to the mean. In many cases of "streaked" distributions (e.g., when primary and secondary magnetizations are incompletely resolved), the mode rather than the mean may be more representative of the primary component of magnetization. The main disadvantage in using the mode is that confidence limits are not as readily determinable as they are for the mean. Thus, it has been difficult, for example, to judge whether the estimated mean for a sample of vectors from a Fisherian distribution and the estimated mode for a sample from a non-Fisherian one were significantly different. Probably because of this statistical disadvantage, the mode, although commonly used in structural geology and petrofabric studies, has rarely been employed in paleomagnetic investigations, except by means of graphical displays in the more recent literature (e.g., Klootwijk, 1975; French et al., 1977; Lowrie and Alvarez, 1977; Morris and Tanner, 1977; Vandenberg et al., 1978).

Computer Technique for Determining Modes of Directional Data

A frequently used procedure for determining modes of sets of directional data consists of contouring equal point densities on stereographic projections (e.g., Knopf and Ingerson, 1938, pp. 245-251). A circular test element, usually with an area of one percent of a hemisphere, is moved along a grid of desired fineness; the number of points falling in this element when centered at each grid-point is determined, and contours (isopleths) of equal point density are drawn. Modes, then, are identified as maxima in the resulting plot. Brotzen (1975) has demonstrated that the location and number of modes on contour plots derived in this manner are a strong function of the size and, to a lesser extent, the shape of the test element:

> As long as the element is smaller than the individual clusters, acceptable maps will be obtained, but only if it is large enough to contain an adequate number of points. If the test element is large and includes considerable portions of two or more clusters, high N values may become associated with individual points, which actually are surrounded by low point densities.

One problem with the point-count contouring technique, in paleomagnetic applications, is that it is of little value in determining modes of samples containing fewer than about 15 unit vectors; yet, samples of paleomagnetic vectors are commonly smaller than N = 10. A second problem with this technique is that the optimum size of the test element, upon which the character of the resulting plot is highly dependent, is imprecisely defined. To be useful in paleomagnetism, any contouring technique must be capable of treating highly variable point densities, since estimated Fisherian concentration parameters of paleomagnetic vector samples range over more than two orders of magnitude.

To circumvent these two difficulties associated with the point-count contouring method, a new technique for determining modes of directional data was devised and incorporated in a FORTRAN computer program (Appendix A). The chief modification of the conventional contouring procedure is that vectors from a given data set are represented not as discrete points, but as identical probability density functions. Specifically, the probability is assigned a value of 1 at the peak of the distribution centered on each unit vector, and decays with angular distance away from this peak according to the probability function:

$$p = \exp(sk(\cos\theta-1)) \tag{1.2}$$

where k is the estimated Fisherian concentration parameter for the total sample of vectors, and s is an arbitrarily-assigned "smoothing parameter."

Use of the probability function (1.2) in determining modes solves both of the problems inherent in the traditional point-count contouring technique. First, because each vector is treated as a probability distribution, it is possible to draw smooth contours for very small samples of vectors. Second, the degree of smoothing of the data is scaled to a calculable statistic (k) related to the actual dispersion of each data set. (Increasing the dispersion of the probability function given by (1.2) has the same effect as increasing the test element size in the point-count contouring technique.) Thus, samples of vectors with radically different estimated concentration parameters can effectively be smoothed to a similar degree by being contoured with similar values of the smoothing parameter, s.

For each vector in a given data set, solutions to (1.2) are determined at the grid-points of a small Cartesian grid of specified fineness, with the observed vector at the center and with a width defined by the angular distance at which $p = 0.01$. The probability distribution given by (1.2) that is used to represent each vector is computed only once for a given data set, since all vectors are to be treated equally. The computed probabilities at the points of each small grid are then transferred to a second, larger Cartesian grid representing an entire hemisphere on an azimuthal equidistant projection and with the same grid-point spacing. This transfer is accomplished by centering each small grid on the coordinates of the large grid that correspond to the direction of the associated vector; the probabilities at the points of each small grid are then added to the probabilities at the points of the large grid, which have initially been set equal to zero. After the probabilities of the last small grid have been added, the probability values at the grid-points of the large grid are scanned for a maximum; the mode is identified as that grid-point with the highest probability value. A computer-driven plotter then generates a map of the cumulative probability distribution by contouring the probability values of the large grid. The contours may be spaced at any desired interval in units of percent of the probability at the mode.

Using the computer technique described above, it is possible to treat even multimodal distributions. However, since only one mode (i.e., that with the highest probability) is recognized on one iteration, precise location of multiple modes requires several repetitions of the program. After the highest mode has been identified, the subset of the total sample that has a mean identical with this mode can be removed from the data set. The probability distribution of the modified data set can then be mapped to locate the next highest mode.

The form of the contoured distribution is partly dependent on the location of the observed vectors with respect to the pole of the projection on which they are displayed. For a large vector sample drawn from a Fisherian distribution and with a mean at the center of a stereographic projection, the contours are essentially concentric circles around the mean. If the pole of this projection were then rotated through 90°, and the same vector sample were contoured again, the contours would become roughly elliptical because of distortions inherent in the projection. For this reason, the computer program is written so that the pole of the projection is positioned near the mode of interest before the probability distribution is mapped. Skewness in the distribution of vectors around this mode is then easily recognized by inspection of the probability map.

Application to Specific Examples of Paleomagnetic Data Sets

In most paleomagnetic studies, the investigator wishes to find a single value most representative of a direction of magnetization or a paleomagnetic pole. All samples of paleomagnetic vectors, however, contain local clusters that usually have no statistical or geophysical significance. Whether a particular probability map is unimodal or multimodal is a strong function of the degree of smoothing of the data. In the probability mapping technique described above, the degree of smoothing of the data is inversely proportional to the smoothing parameter, s. The following generalizations are based on application of this technique to a wide variety of paleomagnetic vector samples. When dealing with distributions that are not markedly skewed, using small values of s ($\sim$ 0.1) leads to oversmoothing and nearly circular contour lines; since the contribution of outliers to the mode increases for small values of s, the mode approaches the estimated mean, and the probability at the mode approaches N, the number of unit vectors. Values of s near 1.0 generally lead to unimodal probability maps, but the mode diverges from the estimated mean if the distribution is not axially symmetric or if there are asymmetrically-scattered outliers in a small sample. In addition, the contours exhibit more structure, by conforming more closely to the actual distribution of vectors in the sample; the maximum probability is commonly about 0.5N. At larger values of s (> 5), the probability map is generally undersmoothed, resulting in multiple modes that have no statistical significance.

The effects of varying the smoothing parameter are well illustrated by mapping probability distributions of actual paleomagnetic data sets. The first example is of 78 virtual geomagnetic poles (VGP's) representing site poles from Newark Group diabases (Beck, 1972). These 78 VGP's constitute a relatively large, nearly axially-symmetric vector sample, although Baag and Helsley (1974) challenged the validity of Beck's χ^2 test for the Fisherian distribution. A 0.5° difference between the mean of the 78 VGP's determined in this paper and Beck's published mean results from assigning a common location coordinate to sites from the same intrusion in attempting to reproduce the VGP's from Beck's published site magnetization directions.

Figure 1.1 graphically shows the effects of contouring these 78 VGP's with s varying from 0.1 to 20.0; Table 1.1 documents the slight changes in location of the mode. It is informative to view these changes by treating the mode as a vector with a length given by the probability at the mode. The changes in the direction and probability at the mode can then be plotted on diagrams similar in construction to vector demagnetization diagrams (Zijderveld, 1967). In these "progressive modal diagrams," the probability at the mode is the analogue of the intensity of magnetization in a vector demagnetization diagram. Figure 1.2 shows that for the 78 Newark Group VGP's, paths for both latitude and longitude are nearly linear to the origin over a range in smoothing parameter of two orders of magnitude. This example demonstrates that the modes of large samples drawn from approximately Fisherian distributions have high directional stability; the solution for the mode is nearly independent of the smoothing parameter.

Close inspection of Table 1.1 and Figure 1.2 reveals that the mode and mean of the 78 VGP's are identical for $s \leq 0.25$, but that the mode decreases 1° in latitude between $s = 0.25$ and $s = 0.50$. This new direction persists until $s = 2.50$, when another 1° directional change occurs. The additional slight change in the direction of the mode that occurs at larger s values (about 1° by $s = 20.0$) is probably not significant, because a major portion of the total sample does not contribute to the probability at the mode.

Figure 1.1 - Probability contour maps for a relatively large vector sample with an approximately Fisherian distribution. The azimuthal equidistant projection at the center shows 78 virtual geomagnetic poles (VGP's) from the Newark Group (Upper Triassic to Lower Jurassic) (computed from data of Beck, 1972). Around this projection are probability maps for the same 78 vectors derived by using different values of the "smoothing parameter," s, but holding k constant at 117 in equation (1.2) of the text. Contours are shown for 15%, 35%, 55%, 75%, and 95% of the maximum probability at the mode. The projection for the probability maps has been rotated so that its pole is centered on the mean of the total sample.

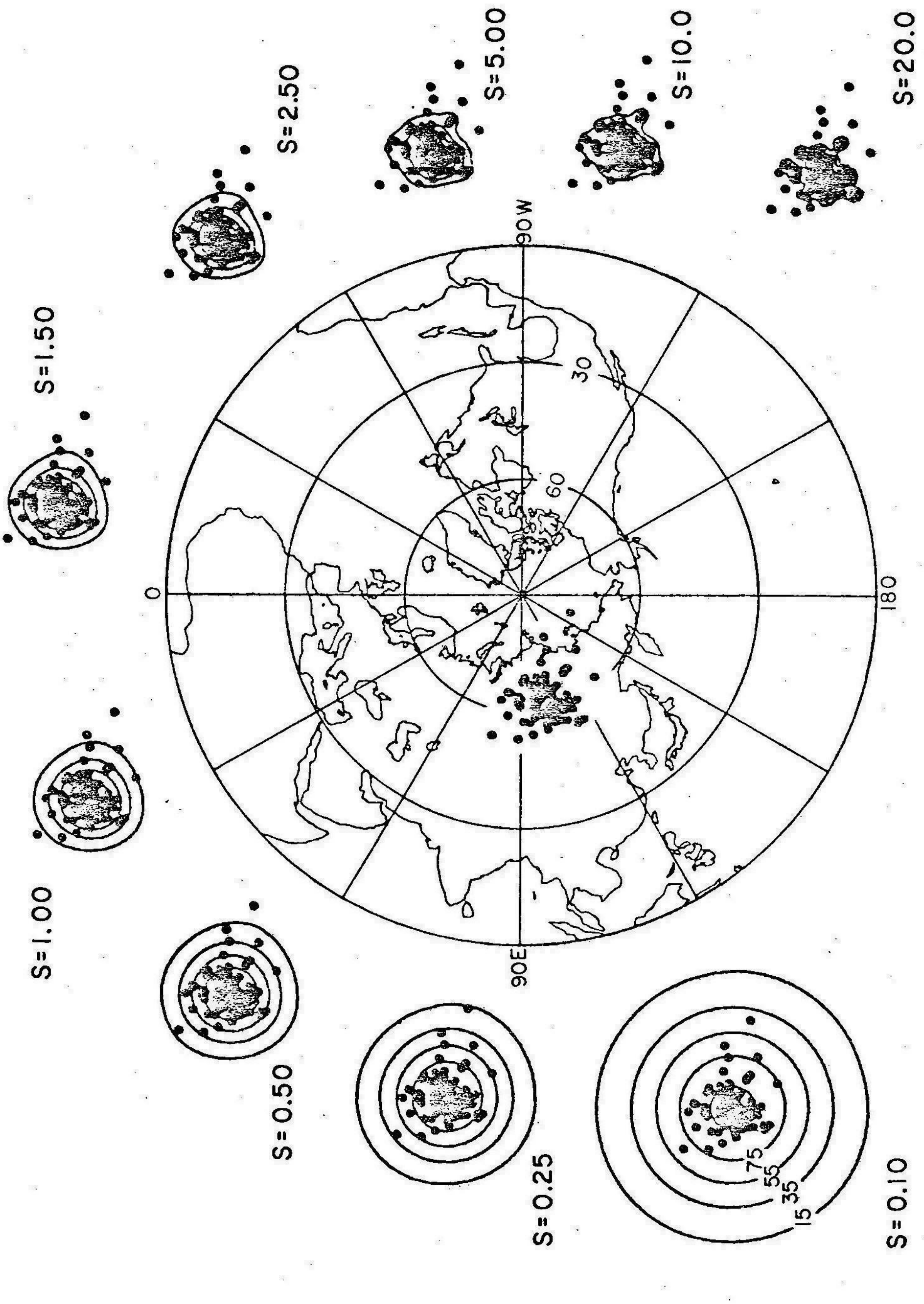

S=0.10
S=0.25
S=0.50
S=1.00
S=1.50
S=2.50
S=5.00
S=10.0
S=20.0
90W
0
180
90E
30
60
75
55
35
15

TABLE 1.1

Effects of Varying the Smoothing Parameter (s) on the Mode for 78 Virtual Geomagnetic Poles from the Newark Group*

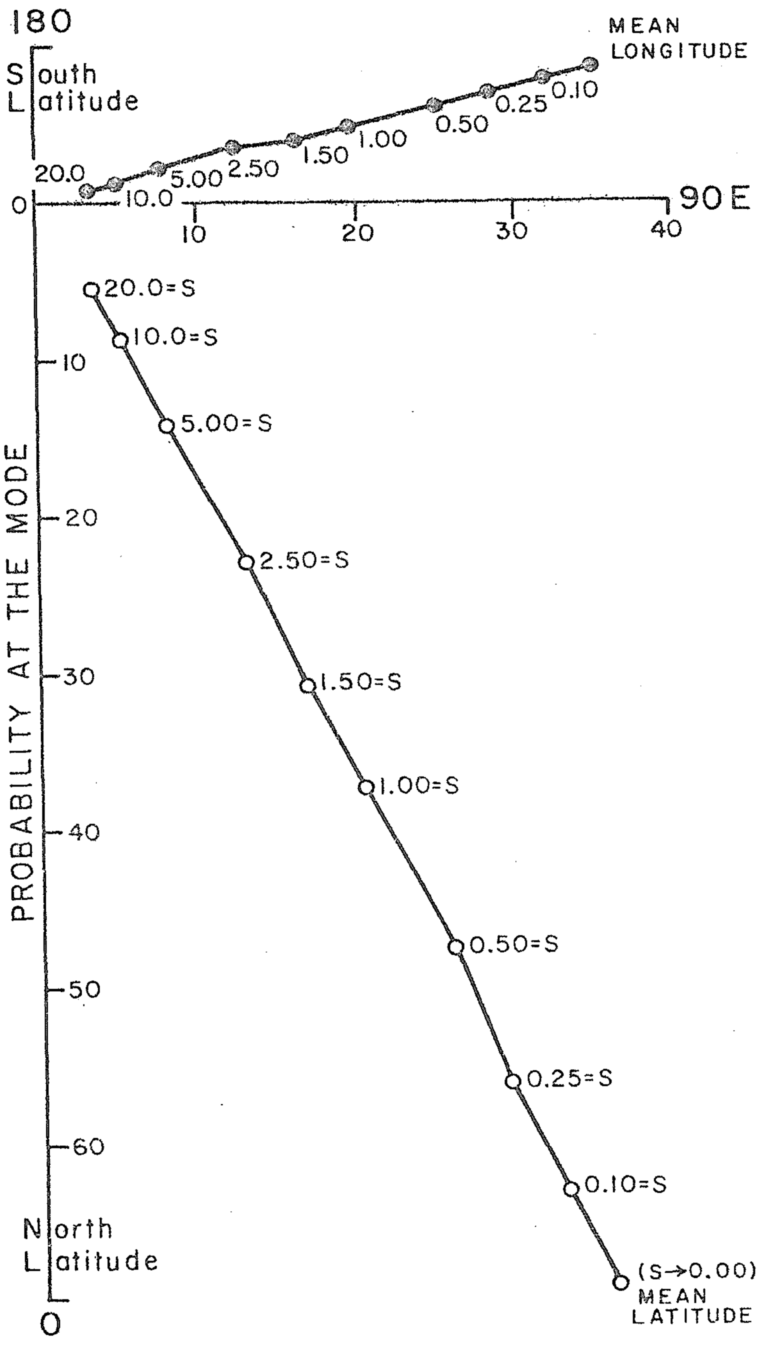
s	P†	Mode	
		Lat°N	Long°E
0.00**	78.0	62.5,	103.7
0.10	71.2	62.5,	103.7
0.25	63.4	62.5,	103.7
0.50	54.1	61.5,	103.7
1.00	42.4	61.5,	103.7
1.50	35.0	61.5,	103.7
2.50	26.3	60.5,	105.8
5.00	16.5	60.5,	105.8
10.00	10.1	59.5,	103.7
20.00	6.4	59.5,	103.7

*Data slightly modified from Beck (1972) (see text).

**As s approaches 0.00, the mode approaches the mean (62.5N, 103.7E).

† P is the maximum probability value at the mode.

Figure 1.2 - Progressive modal diagram of 78 Newark Group VGP's. This diagram illustrates the effect on the mode of varying the smoothing parameter, s. The mode is treated as a vector, with a length given by the probability at the mode derived from the probability maps shown in Figure 1.1. The conventions used in constructing this diagram are adapted from the vector demagnetization diagrams of Roy and Park (1974). Shaded circles represent the endpoint of the modal vector projected onto the equatorial plane; the longitude of the mode is the angle (measured counterclockwise from 0°E) between the 0-180°E axis and the equatorial component. The distance from open circles to the origin is the total length of the modal vector (i.e., the probability at the mode); the latitude of the mode is the angle between the abscissa (which represents the equatorial plane) and the ray from the origin through the open circle. Note that open circles do not represent a projection of the modal vector onto a fixed plane.


180
South Latitude
MEAN LONGITUDE
0.10
0.25
0.50
1.00
1.50
2.50
5.00
10.0
20.0
0
90E
10
20
30
40
20.0=S
10.0=S
5.00=S
2.50=S
1.50=S
1.00=S
0.50=S
0.25=S
0.10=S
(S→0.00) MEAN LATITUDE
PROBABILITY AT THE MODE
10
20
30
40
50
60
North Latitude
0

The differences between the mode and the mean are more readily apparent in treating the smaller vector samples more commonly encountered in paleomagnetism. The mean and its statistical confidence for small N are especially susceptible to bias caused by outliers. This is illustrated by Figure 1.3 and Table 1.2, which show the results of probability mapping of 10 paleomagnetic poles from North America for the time interval 0-3 m.y.B.P. (Van Alstine and de Boer, 1978). Clearly, the mean of these 10 poles (Table 1.2) is biased by an outlier at 60°N, 241°E and is "near-sided" in the sense of Wilson and McElhinny (1974). On the other hand, the modes at all values of s (Table 1.2) are "far-sided," except for those computed at small s (0.20), which represent oversmoothing of the vector distribution. The mode of this small sample at s = 1.00 (the smoothing recommended for nonskewed distributions) deviates by 3^{o} from the mean, although this difference is well within the α_{95} (half-angle of the cone of 95% confidence; Fisher, 1953) for the mean.

Results of other applications of the probability mapping technique to small, nearly axially-symmetric vector samples have been reported by Van Alstine and de Boer (1978). In revising the Phanerozoic apparent polar wander path for North America, they determined both modes and means for 10 time intervals for which the number of poles given unit weight was at least 3. For an average sample size of slightly over 5 paleomagnetic poles per interval, the average difference between the mode and mean (at s = 1.0) was about 3°, with a range of 1° to 5°. Thus, differences of up to several degrees between the mode and mean are

Figure 1.3 - Probability contour maps for a relatively small vector sample with an approximately Fisherian distribution. The azimuthal equidistant projection at the center shows 10 paleomagnetic poles from North America representing the past 3 m.y. (interval 1A of Van Alstine and de Boer, 1978). Around this projection are probability maps for the same 10 vectors derived by using different values of the smoothing parameter, s, but holding k constant at 30 in equation (1.2) of the text. Contours are shown for 15%, 35%, 55%, 75%, and 95% of the maximum probability at the mode. The projection for the contour maps has been rotated slightly so that its pole is centered on the mean of the total sample (marked by the cross).

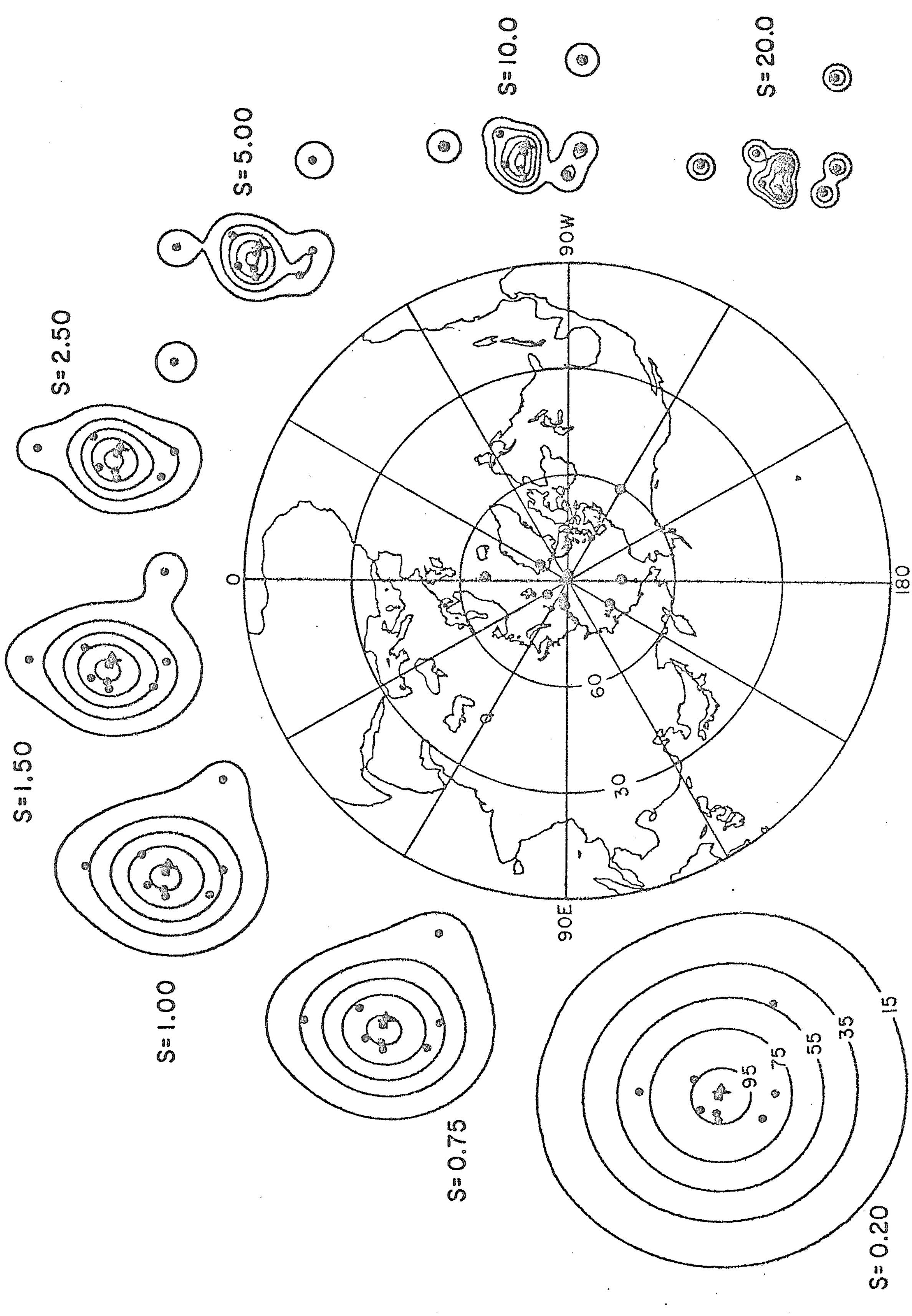

S=1.50
S=2.50
S=5.00
S=10.0
S=20.0
0
90W
180
90E
60
30
S=1.00
S=0.75
S=0.20
95
75
55
35
15

TABLE 1.2

Effects of Varying the Smoothing Parameter (s) on the Mode for 10 Paleomagnetic Poles from North America*

s	P†	Mode	
		Lat°N	Long°E
0.00**	10.0	88.8,	265.9
0.10	9.2	89.8,	265.9
0.20	8.6	89.8,	265.9
0.25	8.3	89.2,	85.9
0.50	7.3	89.2,	85.9
0.75	6.6	88.0,	56.4
1.00	6.2	88.0,	56.4
1.50	5.5	88.0,	56.4
2.50	4.7	87.3,	37.4
5.00	3.7	87.3,	37.4
10.00	2.6	86.1,	71.0
20.00	1.9	84.1,	76.0
40.00	1.5	84.1,	76.0

*Poles represent data for time interval 1A (0-3 m.y.B.P.) of Van Alstine and de Boer (1978).

**As s approaches 0.00, the mode approaches the mean (88.8N, 265.9E).

† P is the maximum probability value at the mode.

typical for small, nearly axially symmetric vector samples. The mode seems preferable for constructing apparent polar wander paths because it is unaffected by outliers; hence, its position is less likely to change as more data are accumulated.

Application to Markedly Skewed Distributions

Identification of the stable position of the mode in a markedly skewed or "streaked" distribution requires some modifications to the procedures described above. For example, Figures 1.4 and 1.5 and Table 1.3 show the effect of probability mapping on the skewed distribution of 122 directions of magnetization from the Wood Canyon Formation after thermal demagnetization at 640°C (data of Gillett and Van Alstine, Chapter 2). In deriving these probability maps, s was varied between 0.4 and 112.0, whereas k was held constant at 9.0 (i.e., the k of the total $N = 122$ sample). The mean of the total sample is biased by specimens containing a high-blocking-temperature component of recent origin aligned roughly with the present axial dipole field at the sampling site. Because this secondary component has not been completely removed from all specimens, both the mean and the "mode" for s near 1.0 misrepresent the direction of the "characteristic magnetization" (Zijderveld, 1967). As shown by the progressive modal diagram (Figure 1.5), however, the "mode" at small values of s is unstable; paths for neither the declination nor the inclination are linear to the origin until about $s = 14$ and above. The direction represented by the linear portion of the progressive modal diagram is a more reliable estimate of the characteristic magnetization direction than is the mean of the distribution. As illustrated by this

Figure 1.4 - Probability contour maps for a large vector sample with a markedly skewed distribution. The azimuthal equidistant projection at the center shows 122 directions of magnetization from the Wood Canyon Formation (Lower Cambrian) after thermal demagnetization at 640°C and after correction for the dip of the beds (data of Gillett and Van Alstine, Chapter 2). Solid (open) symbols are on the lower (upper) hemisphere, respectively. Around this projection are probability maps for the same 122 vectors derived by using different values of the smoothing parameter, s, but holding k constant at 9 in equation (1.2) of the text. Contours are shown for 15%, 35%, 55%, 75%, and 95% of the maximum probability at the mode. The projection for the probability maps has been rotated so that its pole is centered on the mean of the total sample (marked by the open cross). The star marks the direction of the present axial dipole field at the sampling site after correction for the dip of the beds.

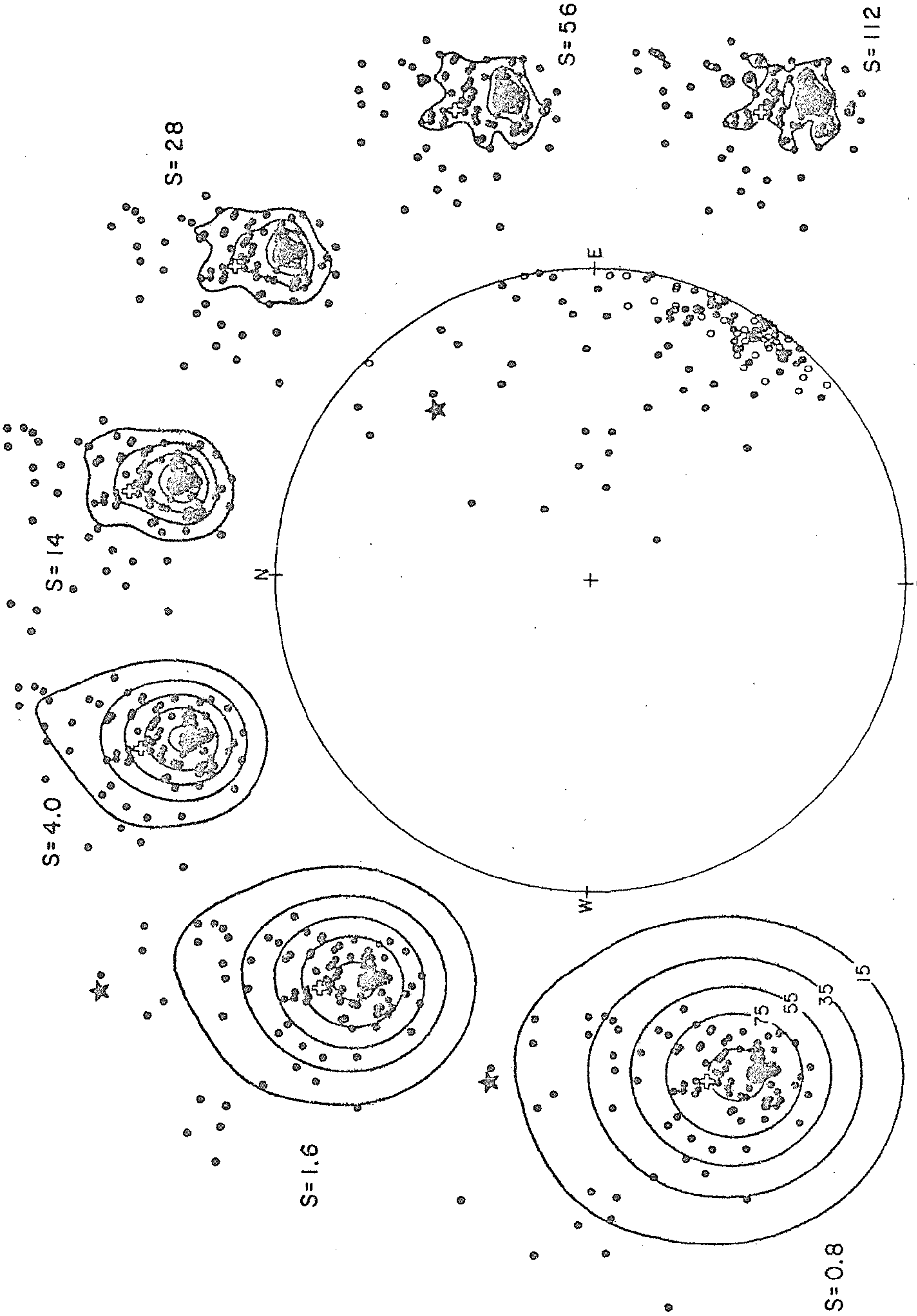
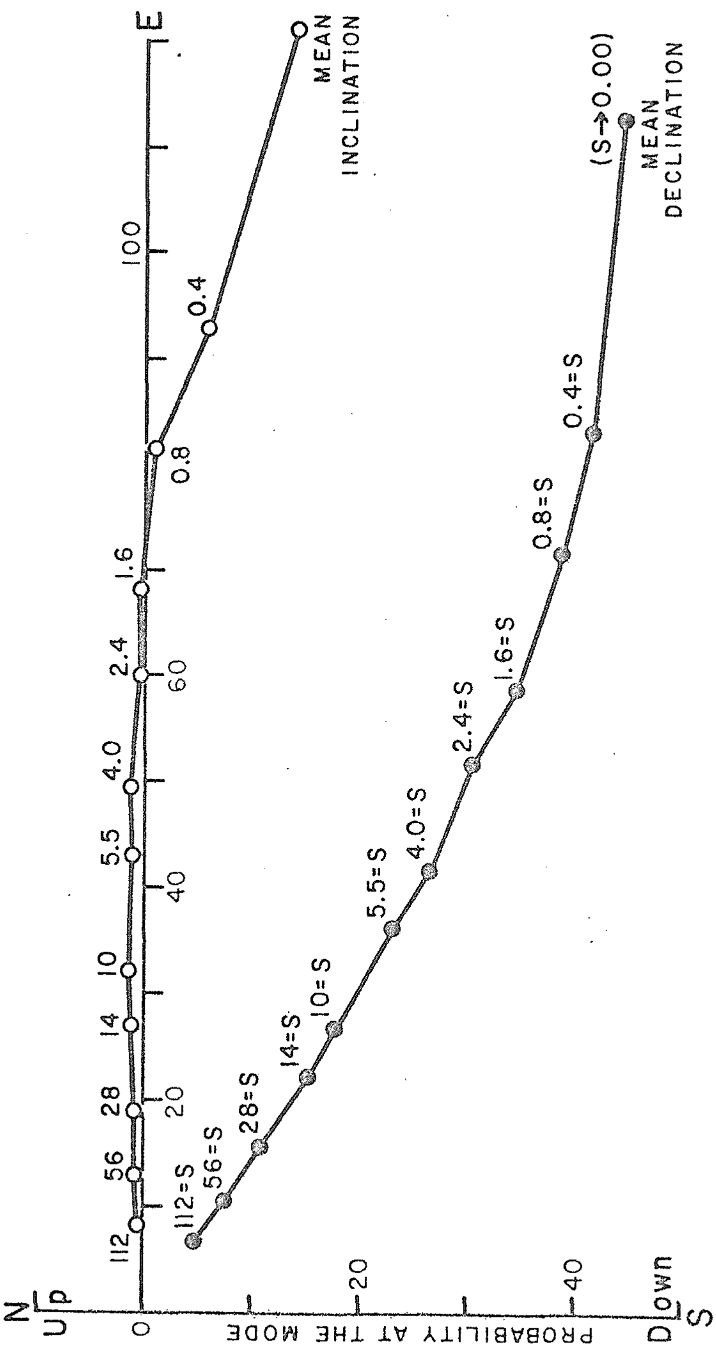

S=0.8
S=1.6
S=4.0
S=14
S=28
S=56
S=112
N
E
S
W
15
35
55
75

Figure 1.5 - Progressive modal diagram of 122 magnetization directions from the Wood Canyon Formation. This diagram illustrates the effect on the mode of varying the smoothing parameter, s. The mode is treated as a vector, with a length given by the probability derived from probability maps, some of which are shown in Figure 1.4. The conventions used in constructing this diagram are similar to those used in the vector demagnetization diagrams of Roy and Park (1974). Shaded circles represent the endpoint of the modal vector projected onto the horizontal plane; the declination of the mode is the angle (measured clockwise from due north) between the north-south axis and the horizontal component. The distance from open circles to the origin is the total length of the modal vector (i.e., the probability at the mode); the inclination of the mode is the angle between the abscissa (which represents the horizontal plane) and the ray from the origin through the open circle. Note that the open circles do not represent a projection of the modal vector onto a fixed plane.



N
Up
E
0
20
40
60
100
PROBABILITY AT THE MODE
Down
S
112
56
28
14
10
5.5
4.0
2.4
1.6
0.8
0.4
MEAN INCLINATION
112=S
56=S
28=S
14=S
10=S
5.5=S
4.0=S
2.4=S
1.6=S
0.8=S
0.4=S
(S→0.00)
MEAN DECLINATION

TABLE 1.3

Effects of Varying the Smoothing Parameter (s) on the Mode for 122 Directions of Magnetization from the Wood Canyon Formation*

s	P†	Mode Dec.	Mode Inc.
0.0**	122.0	111.5,	+6.6
0.4	93.1	116.5,	+3.6
0.8	81.6	118.4,	+0.6
1.6	68.4	120.4,	-0.4
2.4	60.1	120.4,	-0.4
4.0	49.5	122.4,	-1.5
5.5	43.1	122.4,	-1.5
10.0	32.3	123.4,	-2.5
14.0	27.2	124.4,	-2.5
28.0	19.1	124.4,	-2.5
56.0	13.5	125.3,	-2.6
112.0	8.2	125.4,	-3.5

*Directions are from the 640°C demagnetization step and have been corrected for the dip of the beds; data of Gillett and Van Alstine (Chapter 2).

**As s approaches 0.00, the mode approaches the mean (D = 111.5, I = +6.6).

† P is the maximum probability value at the mode.

example, in order to obtain an accurate estimate of the mode of a markedly skewed distribution, the product $s \times k$ in equation (1.2) must be considerably larger than 1.0 times the k of the total vector sample. For these distributions, the best procedure is to derive probability maps with progressively larger values of s until the estimate of the mode stops migrating away from the tail of the distribution.

Although the procedure of progressive probability mapping is recommended for determining the mode of a skewed distribution, a good estimate can be obtained in a single step. This can be accomplished by substituting in equation (1.2) not the k of the entire vector sample, but instead the k for a set of vectors near the mode; s can then be set at 1.0, and the mode will be very close to that derived by progressive probability mapping. In the Wood Canyon Formation example, contouring with $s = 1$, $k = 642$ (the k of the 25 directions closest to the mode; cf. Table 1.4) would yield the same mode as contouring with $s = 71$, $k = 9$ (the k of the total $N = 122$ sample), since it is the product $s \times k$ in equation (1.2) that determines the probability distribution. Even if the 48 vectors closest to the mode were chosen and the probability distribution were mapped with their k of 152 and $s = 1$, the resulting mode would still fall on the linear portion of the progressive modal diagram (Figure 1.5) at about $s = 17$.

In the Wood Canyon example, the mode identified by progressive probability mapping is about 16° different from the mean of the total sample. This illustrates the danger in attaching geophysical significance to the mean of skewed distributions of paleomagnetic directions. It must be emphasized that probability mapping can only be successful in identifying the direction of the characteristic magnetization if that direction is represented by a maximum in the distribution of vectors. However, the mode rather than the mean provides a better estimate of the characteristic magnetization direction as long as the remanent magnetization of the majority of samples is dominated by the characteristic component.

Precision of the Mode and Estimates of Confidence Limits

The fineness of the Cartesian grid used for probability mapping limits the precision with which the mode can be determined. Decreasing the grid-point spacing increases the amount of computer time and storage required, while decreasing the error in locating the mode. Since the estimate of the mode is that grid-point with the highest probability value, the maximum error in locating the mode introduced by finite grid size is $G \times 0.707$, where G is the grid-point spacing in degrees. A 2° grid-point spacing has been found to be adequate for working plots or for ascertaining the effects of varying the smoothing parameter for a particular data set. A 1° grid-point spacing, which is recommended for more precise determination of the mode, could introduce a maximum error of 0.7° and has a probable error of 0.4°. Because these errors are considerably smaller than most published α_{95} values, the increased precision obtained in using finer grids generally does not justify the added computing costs.

An estimate of confidence limits for the mode can be made, based on the property that the mode and mean of a Fisherian distribution are identical. This property suggests that a new statistic, designated β_{95}, can provide an estimate of the 95% confidence limits for the mode. The statistic β_{95} is derived from the largest subset of the total sample that has a mean identical with the mode of the total sample. This statistic is defined as the Fisherian half-angle of the cone of 95% confidence for the mean of this subset.

This definition of β_{95} must be qualified in two respects. First, because of the effects of finite grid and sample sizes, generally none of the subsets of a given sample will have a mean identical with the mode of the total sample. Hence, β_{95} should be derived from the largest subset having a mean that differs from the mode of the total sample by no more than $0.707 \times G$, where G is the grid-point spacing in degrees. Second, if β_{95} is to be an accurate estimate of the 95% confidence limits for the mode, then the subset of the total sample from which it is derived must have a Fisherian distribution. In a rigorous sense, therefore, tests for consistency with the Fisherian distribution should accompany all calculations of β_{95} for the mode, just as they should accompany all calculations of α_{95} for the mean.

In Table 1.4, a procedure for estimating β_{95} is illustrated for the example of the skewed distribution of directions from the Wood Canyon Formation. The stable position of the mode is first determined by progressive probability mapping (Figures 1.4 and 1.5 and Table 1.3). A contracting cone is then centered on this mode, and Fisherian statistics are computed for the enclosed subsets of progressively smaller size. This yields an estimate of β_{95} of about 1.7°. Appreciably smaller or larger estimates for β_{95} are invalid, because they are derived from truncated or skewed distributions, respectively.

TABLE 1.4

Fisherian Statistics* of Points Increasingly Close to the Mode**

for 122 Directions of Magnetization from the Wood Canyon Formation†

Cone Half-Angle (°)	# Contained	"K"	"α_{95}"	Mean Dec., Inc.	Difference (Mean-Mode)
(90.0)	122	9.0	4.5	111.5, +6.6	15.8
54.0	106	21.2	3.1	116.1, +2.3	9.6
51.3	105	22.2	3.0	116.5, +2.3	9.2
48.6	104	23.3	2.9	116.9, +2.2	8.9
45.9	103	24.4	2.9	117.3, +2.3	8.6
43.2	101	26.4	2.8	117.9, +2.1	8.0
40.5	97	31.7	2.6	119.0, +1.5	6.7
37.8	97	31.7	2.6	119.0, +1.5	6.7
35.1	95	34.6	2.5	119.1, +1.1	6.4
32.4	92	39.0	2.4	119.4, +0.4	5.8
29.7	90	41.9	2.3	119.8, +0.2	5.3
27.0	86	47.8	2.2	120.5, -0.2	4.5
24.3	83	52.7	2.2	121.0, -0.6	3.9
21.6	78	61.9	2.1	121.7, -1.1	3.0
18.9	76	64.7	2.0	122.1, -1.2	2.6
16.2	69	79.8	1.9	123.1, -1.5	1.7
13.5	62	97.5	1.8	123.2, -1.9	1.3
10.8	48	151.9	1.7††	124.4, -2.0	0.5
8.1	31	392.1	1.3	124.6, -3.2	0.8
5.4	25	642.2	1.1	124.8, -2.9	0.5
2.7	11	1711.7	1.1	125.1, -2.8	0.8

*Fisher (1953). Note that the estimate of the Fisherian concentration parameter (K) and the half-angle of the cone of 95% confidence for the mean (α_{95}) are being applied to non-Fisherian distributions.

**The mode has been determined by progressive probability mapping and is at D = 124.4, I = -2.5.

†Directions are from the 640°C demagnetization step and have been corrected for the dip of the beds; data of Gillett and Van Alstine (Chapter 2).

††The value 1.7° is suggested as an estimate for the half-angle of the cone of 95% confidence for the mode (β_{95}) (see text).

Conclusions

The techniques described in this study can complement the more conventional, Fisherian method for determining the central tendency of directional data. Probability mapping is especially helpful when dealing with small samples containing outliers or with distributions that are skewed or multimodal. Use of this technique in deriving paleomagnetic poles can increase the accuracy of apparent polar wander paths and paleogeographic reconstructions.

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Chapter 2
PALEOMAGNETISM OF LOWER AND MIDDLE CAMBRIAN SEDIMENTARY ROCKS FROM THE DESERT RANGE, NEVADA

A fairly extensive body of paleomagnetic data from rocks of latest Precambrian and earliest Paleozoic age from the Gondwana continents (McElhinny and Embleton, 1976) suggests that rates of apparent polar wander (APW) were much higher during the late Precambrian and Early Paleozoic than in the Mesozoic and Cenozoic. On this basis, McElhinny _et al._ (1974) predicted that "extensive studies of late Precambrian and Cambrian rocks of North America . . . will reveal a complicated apparent polar wander path with a polar shift of at least 180°." Whereas a relatively continuous APW path can be drawn for North America since the Early Ordovician (Van Alstine and de Boer, 1978), the APW path of the latest Precambrian and Cambrian remains poorly known.

To help extend the continuous North American APW path into these times, we have undertaken a paleomagnetic investigation of upper Precambrian and Lower Paleozoic strata in the Cordilleran Geosyncline. A miogeoclinal sequence exposed in the Desert Range, Nevada, was chosen for study because it is conformable from the upper Precambrian through Lower Ordovician, occurs in a generally homoclinal fault block, shows little if any evidence of regional metamorphism, and is well exposed. Besides the advantage of stratigraphic continuity, this miogeoclinal sequence is known to include the Precambrian-Cambrian boundary, unlike any sequence on the craton. A disadvantage of this section is that it lies within a structurally complex area, so that the probability of tectonic rotations about a vertical axis must be reckoned with. Such rotations may be corrected for, however, by calculating the declination correction required to make paleomagnetic poles from these rocks match poles of the same age from the craton.

The work reported in this and the companion paper (Chapter 3) represents results from over 800 independently oriented samples collected in the Desert Range during four sampling trips over two years. Of this total, about one third come from upper Precambrian strata and two thirds from Lower and Middle Cambrian strata. This chapter focuses on the paleomagnetism of the Cambrian part of the section, with particular emphasis on determining whether there has been post-magnetization rotation of the fault block about a vertical axis. In Chapter 3, we present results from the upper Precambrian part of the section.

Regional Geologic Setting of the Desert Range

The Desert Range is located in Clark County, southeastern Nevada (Figure 2.1). It is an eastward dipping block which is much cut by faulting and which has been interpreted by Longwell (1945, 1960) to be the eastern limb of a broad, north-plunging anticline. Immediately east of the Desert Range is the Sheep Range, part of a huge allochthonous plate which has overridden rocks of the Las Vegas Range, farther to the east, on the Gass Peak Thrust. The large-scale thrusting occurred during the Sevier Orogeny in Late Cretaceous time (Fleck, 1970). Burchfiel *et al.* (1970) have shown that a belt of older

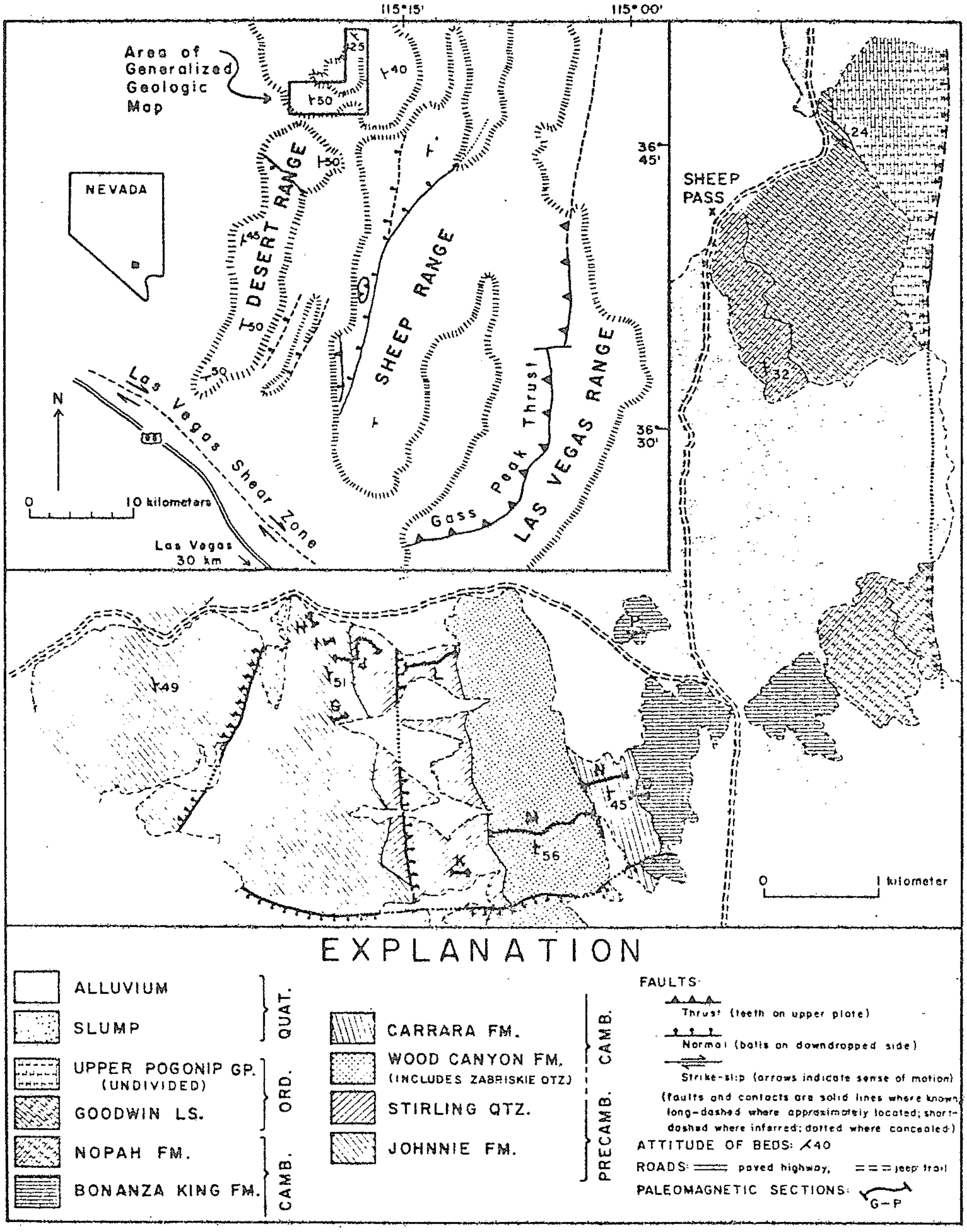


Figure 2.1. Generalized tectonic and geologic maps showing the location of upper Precambrian and Cambrian paleomagnetic sections. Geologic mapping by D.R. Van Alstine, J.L. Kirschvink, and S.L. Gillett (1975-1977).

(probably late Triassic to early Jurassic) thrust faults is exposed about 25 km west of the Desert Range. Normal faults, related to late Tertiary Basin and Range extension, separate the downfaulted east side of the Desert Range from the Sheep Range, and similar faults probably define the western side of the range. Because of its position with respect to the Mesozoic thrust belts, the Desert Range is probably allochthonous, but the inferred basal thrust is hidden.

To the south, the Desert Range is truncated by the Las Vegas Shear Zone, a fault zone of right lateral displacement which is a major tectonic element in southern Nevada (e.g., Longwell, 1960, 1974; Stewart, 1967). The time of the major movement on this shear zone probably lies in the range 17-11 m.y.B.P. (Ekren et al., 1968; Longwell, 1974). The pronounced westward bend of the strike of the beds at the southern tip of the range has been interpreted as drag on this shear zone (cf. Burchfiel, 1965; Longwell et al., 1965; Albers, 1967; Stewart, 1967). These authors have also commented on similar westward bends in the structural grain and in isopach trends of strata in ranges nearby. Albers and Stewart, in particular, have presented evidence for major right-lateral displacement and large-scale oroflexural bending in this part of the Basin and Range Province; Albers attributes arcuate trends of many of the ranges to such oroflexural bending. It seems likely that the moderate northwesterly change in trend of the Desert Range north of the area sampled, reflecting a change in strike of the underlying strata, also represents an oroflexural bend.

Upper Precambrian and Paleozoic rocks exposed in the Desert Range are miogeoclinal deposits typical of the Cordilleran Geosyncline (Stewart, 1970). Deposition in the geosyncline was essentially continuous from late Precambrian into Early Permian time; and the total thickness of strata in this area probably exceeds 9 km (cf. Longwell et al., 1965; Burchfiel et al., 1974). The section studied is a spectacularly exposed homocline which includes about 6500 m of upper Precambrian through Ordovician strata (Figure 2.1). In this chapter, we will present paleomagnetic data on rocks from the Wood Canyon Formation (Lower(?) and Lower Cambrian), the Zabriskie Quartzite (Lower Cambrian), the Carrara Formation (Lower and Middle Cambrian), and the lowermost Bonanza King Formation (Middle Cambrian).

Sampling, Laboratory, and Statistical Procedures

Only oriented block samples could be taken from the Desert Range, because the area is part of the Desert National Wildlife Range, where motorized coring is forbidden. We therefore devised a block sampling technique that is reasonably fast and which works equally well with either smooth rocks such as quartzites or ones with jagged surfaces, such as solution-pitted carbonates.

Glass microscope slides, trimmed to 2.5 cm squares, were glued flat with automobile weather-stripping adhesive to the blocks to be collected. The upper edge of each slide was leveled by means of a hand-held spirit level. After the adhesive had fully set, the strike and dip of the slide were measured with a Brunton compass, and the block was collected. At this point, the slide was also outlined on the surface of the rock sample, with the top indicated, so that an approximate orientation could be recovered even if the slide were accidentally broken off in transit to the laboratory. This technique for sample orientation is probably more accurate than conventional block sampling, because a flat surface with a machined straightedge is affixed to each rock and oriented independently of the character of the rock surface itself.

Blocks were collected in stratigraphic sequence. The stratigraphic intervals in the initial reconnaissance were estimated by pacing, but the sections later sampled in detail were measured with a tape and/or Jacob's staff.

All laboratory work was performed at the U.S. Geological Survey Paleomagnetics Laboratory at Flagstaff, Arizona. Generally two cylindrical cores 2.5 cm in diameter were drilled from each block sample perpendicular to the glass slide. For measurement, only a single specimen 2.5 cm long was trimmed from a core, although duplicate specimens for additional demagnetization experiments were cut as needed. Magnetic measurements were made on an ScT superconducting rock magnetometer interfaced with a DEC PDP 8/M computer. This system has a background noise level less than 1×10^{-7} emu (1×10^{-10} Am^2) and allows immediate, on-line data reduction.

Progressive thermal and/or alternating-field (AF) demagnetization was performed on all samples. Thermal demagnetization was done with a non-inductively wound electrical oven in which the magnetic field within the specimen area was ~20 nT. Specimens were placed in non-systematic orientations during thermal demagnetization, so that systematic spurious components could not be imparted by stray fields. AF demagnetization below 50 mT (500 Oe) was performed with a Schonstedt specimen demagnetizer. To minimize acquisition of anhysteritic remanent magnetization, AF demagnetization above 50 mT was carried out with a 3-axis tumbler.

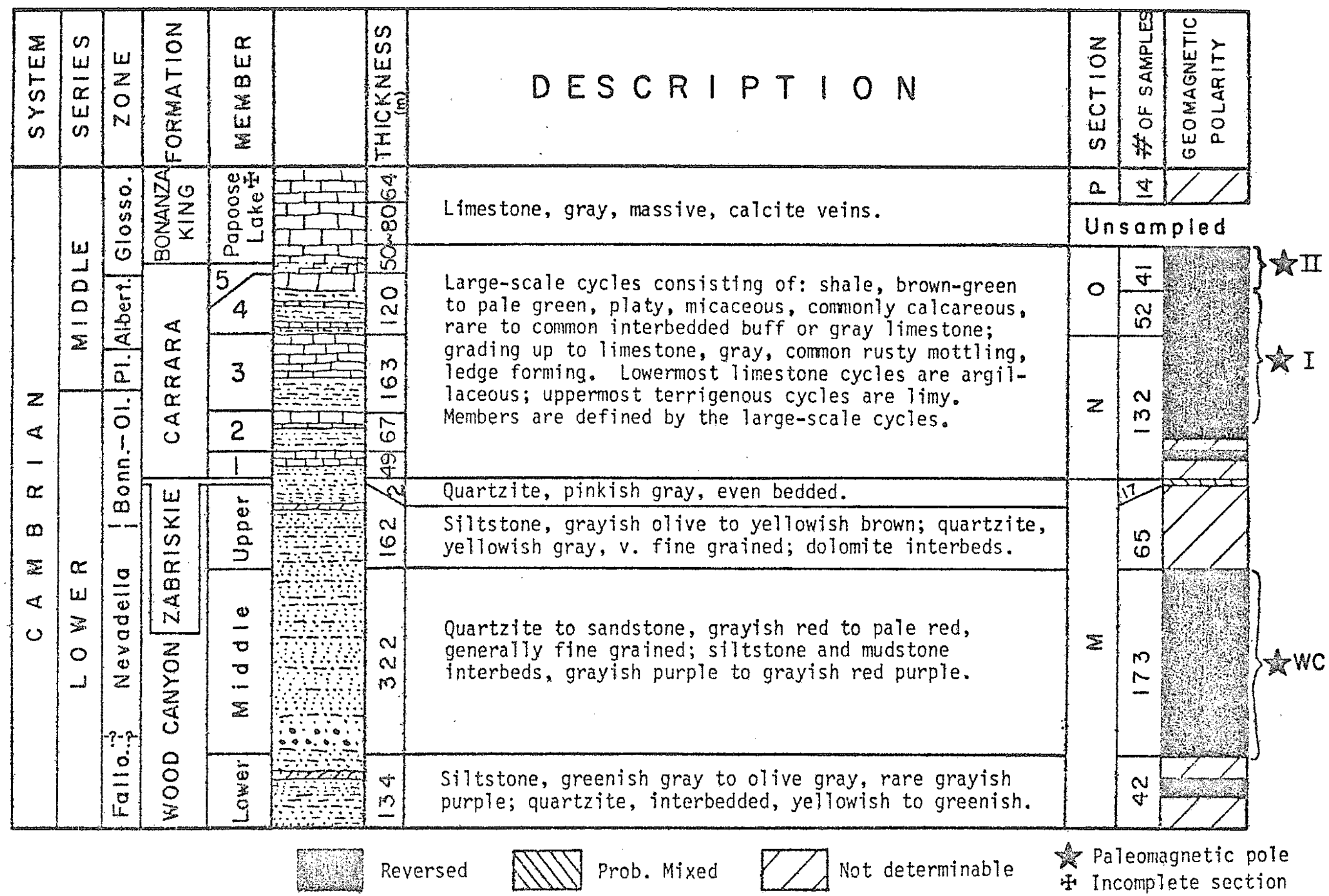
In this paper and its companion (Chapter 3), "geographic coordinates" and "stratigraphic coordinates" refer to the declination and inclination of the magnetic vector in coordinate systems uncorrected and corrected for the strike and dip of bedding, respectively. "Characteristic magnetization" is used here in the sense of Zijderveld (1967, p. 263) as "that magnetization which one is apt to call primary without having definite proof." The progressive demagnetization data from nearly all specimens were plotted by computer, using a modification of Zijderveld's (1967) demagnetization diagrams similar to that employed by Roy and Park (1974).

Because the distributions of magnetic directions encountered in this study are skewed, application of the statistics of Fisher (1953) is inappropriate. This skewness results from the incomplete removal of secondary components of magnetization from all samples. The mode rather than the mean of such distributions provides a better estimate of the characteristic magnetization direction as long as the remaining remanence is dominated by the characteristic component. In Chapters 2 and 3, we have determined modes and estimated their confidence limits by the techniques described in Chapter 1.

Wood Canyon Formation

The Wood Canyon is the lowest formation in the Desert Range sequence that is demonstrably at least partly of Cambrian age on the basis of fossils it contains. The Wood Canyon is composed of a heterogeneous sequence of siltstone, quartzite, and minor carbonate rock, in which Stewart (1970) has described three informal members (lower, middle, and upper) of regional extent (Figure 2.2). At several localities where the Wood Canyon has been studied, Early Cambrian trilobites referable to *Judomia*(?) and *Nevadella* are reported from the upper half of the upper member (Stewart, 1970; Palmer, 1971). *Skolithos* tubes (vertical worm burrows) are also locally abundant in the upper part of the upper member, and pelmatozoan, archeocyathid, and brachiopod fragments are also found in these beds. Because the lowest stratigraphic occurrence of trilobite fossils falls within the upper member, the Wood Canyon has been assigned a latest Precambrian and earliest Cambrian age (Stewart, 1970). However, if lithologic correlations of the Desert Range strata with strata in the White-Inyo Mountains of eastern California are accepted (Stewart, 1970; Nelson, 1978), the lowest trilobites (*Fallotaspis*) in the White-Inyo Mountains section occur in strata correlative with the lower and middle members of the Wood Canyon. Moreover, in the White-Inyo section, *Skolithos*, *Cruziana*, and other possibly Cambrian trace fossils occur even lower than the lowest trilobites. On this basis, Nelson (1978) suggested that the Precambrian-Cambrian boundary may lie

Figure 2.2 Chart showing description of strata, trilobite zones, stratigraphic distribution of samples, and inferred geomagnetic polarity for Lower and Middle Cambrian formations sampled in this study. "Fallo.", "Bonn.-Ol.", "Pl.", "Albert.", and "Glosso." stand for *Fallotaspis*, *Bonnia-Olenellus*, *Plagiura-Kochaspis*, *Albertella*, and *Glossopleura* Zones, respectively. Units labeled 1 through 5 in the Carrara Formation are the large-scale sedimentary cycles (see text). Members comprising the cycles are: Cycle 1, Resting Springs Member of the Zabriskie Quartzite, Eagle Mountain Shale, and Thimble Limestone; Cycle 2, Echo Shale and Gold Ace Limestone; Cycle 3, Pyramid Shale and Red Pass Limestone; Cycle 4, Pahrump Hills Shale and Jangle Limestone; Cycle 5 (incomplete), Desert Limestone. The top of cycle 5 is defined by the basal strata of the Bonanza King Formation (see text). Descriptions of Wood Canyon Formation and Zabriskie Quartzite are adapted from Stewart and Barnes (1966) and Stewart (1970). Descriptions of the Carrara and Bonanza King Formations are from unpublished data of Gillett and Van Alstine.


SYSTEM
SERIES
ZONE
FORMATION
MEMBER
THICKNESS (m)
DESCRIPTION
SECTION
OF SAMPLES
GEOMAGNETIC POLARITY
CAMBRIAN
MIDDLE
LOWER
Glosso.
Pl. Albert.
Bonn.-Ol.
Nevadella
Fallo.
BONANZA KING
CARRARA
ZABRISKIE
WOOD CANYON
Papoose Lake
5
4
3
2
1
Upper
Middle
Lower
64
50~80
120
163
67
49
2
162
322
134
Limestone, gray, massive, calcite veins.
Large-scale cycles consisting of: shale, brown-green to pale green, platy, micaceous, commonly calcareous, rare to common interbedded buff or gray limestone; grading up to limestone, gray, common rusty mottling, ledge forming. Lowermost limestone cycles are argillaceous; uppermost terrigenous cycles are limy. Members are defined by the large-scale cycles.
Quartzite, pinkish gray, even bedded.
Siltstone, grayish olive to yellowish brown; quartzite, yellowish gray, v. fine grained; dolomite interbeds.
Quartzite to sandstone, grayish red to pale red, generally fine grained; siltstone and mudstone interbeds, grayish purple to grayish red purple.
Siltstone, greenish gray to olive gray, rare grayish purple; quartzite, interbedded, yellowish to greenish.
P
14
Unsampled
O
41
52
N
132
M
17
65
173
42
II
I
WC
Reversed
Prob. Mixed
Not determinable
Paleomagnetic pole
Incomplete section

in strata correlative with the upper part of the Stirling Quartzite, which comformably underlies the Wood Canyon.

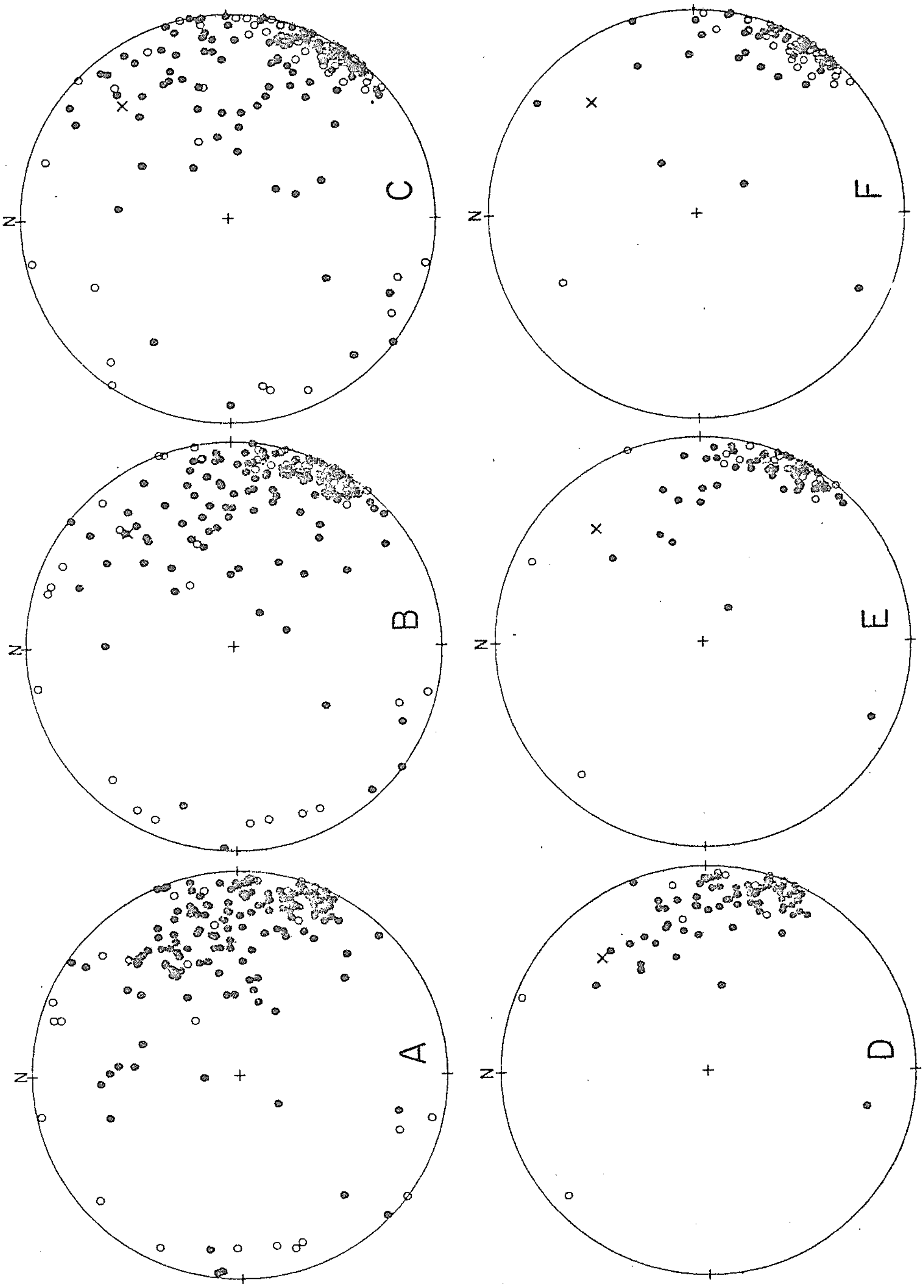
Initially, about 120 samples, representing a variety of rock types, were collected from the Wood Canyon Formation at approximately 5 m intervals (section M in Figure 2.1). Resampling was concentrated mostly in red-purple mudstone interbeds in the middle member and in a few horizons in both the lower and upper members that had given promising preliminary results.

All the initial samples from the Wood Canyon were progressively thermally demagnetized at 6 steps between 300 and 610°C; most were then further demagnetized at 640°C. Some of the lower thermal steps were omitted for the latest samples collected, but in no case were fewer than 3 thermal demagnetization steps performed on any sample. In addition, 24 samples showing directions farthest from that of the present axial dipole field at the sampling site were further demagnetized: 12 at 650°C, and the remaining 12 at 670°C. Results from the middle member, in which a characteristic magnetization could be identified, are presented first; discussion of the generally scattered directions from the rest of the formation is deferred until later.

Middle Member, Wood Canyon Formation

NRM directions for most samples from the middle member of the Wood Canyon were distributed between the present axial field direction and a direction in stratigraphic coordinates with a declination of about 110° and an inclination near 0° (Figure 2.3). After

Figure 2.3 Directions at selected steps from progressive thermal demagnetization of samples from the middle member of the Wood Canyon Formation. Equal-area projection in stratigraphic (dip-corrected) coordinates. Open (shaded) circles are on upper (lower) hemisphere, respectively. "X" marks the direction of the present axial dipole field at the sampling site after rotation for the mean attitude of beds. Directions from all samples are shown in "A," "B," and "C" (top row): (A) NRM; (B) 550°C; (C) 640°C. "D," "E," and "F" show the directions from red-purple mudstones only (bottom row): (D) NRM; (E) 550°C; (F) 640°C.


A
B
C
D
E
F
N

demagnetization at 640°C, the directions had become better grouped with a mode at a low negative inclination and a declination near 125°; fewer samples showed a pronounced present-axial-field component. Directions from the samples further demagnetized at 650°C (not illustrated) showed increased scatter; all samples demagnetized at 670°C exhibited scattered directions and viscous components of magnetization.

In detail, the thermal demagnetization behavior of a given sample from the middle member is a function of its grain size and color. As summarized in Figure 2.2, the middle member of the Wood Canyon consists predominantly of pale reddish quartzites, with darker purplish and red-purple siltstone and mudstone interbeds. Most of the directions far from that of the present axial field before and after demagnetization are found in the finer-grained rocks (Figure 2.3). The finer-grained rocks are also the darkest (5 RP 4/2 versus 5 RP 6/2 or 5 P 6/2). In addition, samples which are red-purple (5 RP 4/2) rather than purple (5 P 4/2) tend to have directions farther from that of the present axial field.

The outliers in Figure 2.3 are probably caused by lightning strikes. This is suggested by the anomalously high intensities of magnetization in a number of these samples. Several of these samples were progressively AF demagnetized up to several hundred milliteslas in an attempt to remove a possible isothermal remanent magnetization (IRM), but the directions from none changed significantly with either AF or thermal demagnetization. While some samples with anomalous directions do not have anomalous intensities, in general they were obtained near samples which do. A few anomalous directions are nearly antipolar to the mode defined by the mudstones, which suggests they might represent zone(s) of opposite polarity. However, these directions were not reproduced in samples collected at stratigraphically equivalent horizons but off the ridge crest where most samples of the middle member were taken.

The high blocking temperature of the characteristic magnetization of the middle member, as well as the deep red-purple color of these rocks, demonstrates that this component resides in hematite. Before this magnetization can be accepted as representative of the Early Cambrian geomagnetic field, however, evidence for the time of its acquisition must be sought. Accordingly, we now present textural, petrographic, and stratigraphic evidence that the characteristic remanence of the middle member probably was acquired penecontemporaneously with deposition.

All samples were examined with a low-power binocular microscope. Thin sections of 8 red-purple mudstones, 3 purplish siltstones, and one coarser-grained, pale quartzite were examined in both transmitted and reflected light.

The petrographic observations indicate that at least 3 types of hematite are present in the rocks of the middle member of the Wood Canyon Formation. (1) Discrete grains of specular hematite, generally 5-25 μm in diameter. A detrital origin for these grains is inferred from rounding, which suggests transport; apparent size sorting; lack of alteration halos; occurrence of imbricate trains suggestive of heavy mineral lenses; and elongation subparallel to the bedding fabric, suggesting depositional control. (2) Aggregates containing minute grains ($\sim$1 μm) of specular hematite and/or red, earthy, microcrystalline hematite, presumably formed by alteration of ferroan precursors. An authigenic origin is suggested because such aggregates apparently could not survive transport, and because commonly they have diffuse contacts suggestive of alteration halos. Both of the preceding forms of hematite occur in all the rock types. (3) Microcrystalline red cement, occurring primarily in the red-purple mudstones. This cement, which is responsible for the deep red-purple (5 RP 4/2) color of these rocks, is largely confined to very fine-grained parts of the rock; coarser, quartzo-feldspathic laminae and lenses are light-colored.

All three forms of hematite may contribute to the characteristic magnetization of these rocks. The presence of detrital specular hematite in all rock types of the middle member suggests that part of the characteristic remanence may be a detrital remanent magnetization (DRM). A chemical remanent magnetization (CRM) in the authigenic specular hematite may also contribute to the characteristic magnetization; if so, constraints similar to those discussed below apply to the timing of the acquisition of this CRM. A component of the characteristic magnetization can be discerned in most specimens from the middle member, even in the pale quartzites which contain very little of the red cement. Thus, the detrital specular hematite and/or the hematitic aggregates apparently contribute in part to the characteristic magnetization.

The red cement, however, is probably the most important carrier of the characteristic magnetization. This cement is the most distinguishing feature of the red-purple mudstones and must postdate their deposition to some degree; as elaborated above, the red-purple mudstones also show the best grouping of directions, farthest from the direction of the present axial field. Together, these observations indicate that a large part of the characteristic magnetization may be a CRM residing in the red cement.

Textural relations in these mudstones support an Early Cambrian time of hematization and cementation, suggesting that any CRM associated with these processes is also Early Cambrian. Within the red-purple mudstones, bioturbation is suggested by light-colored, commonly rounded pods that are filled with coarser, quartzo-feldspathic material; the bedding fabric of the mudstone bends around the pods. The pods probably represent burrows infilled with some of the overlying, coarser-grained sediment. Burrowing is also suggested by distorted, light-colored laminae that contain disrupted and bent pieces of the surrounding red-purple mudstone. Even where bioturbation is present, the contact of coarser, light-colored material and fine, red-purple mudstone is sharp at the scale of the thin sections. Hence, at least the precursor of the hematitic cement was present prior to the bioturbation. One would expect a pervasive, later alteration to show more diffuse contacts. Further, such an alteration should follow preferentially the coarser-grained areas, because of their higher permeability. In the middle member, however, it is the least permeable beds that are the most hematitic.

An Early Cambrian time of acquisition of the characteristic magnetization of the middle member is also likely on regional stratigraphic grounds. A superimposed, post-Early Cambrian chemical alteration should occur somewhat sporadically and have other factors affecting its regional distribution (e.g., degree of fracturing, metamorphic grade) besides the physical stratigraphy, as is indeed observed with albite metasomatism at some localities in the lower member (Stewart, 1970). However, the hematitic middle member, as Stewart shows, is regionally persistent over a distance of about 400 km along the sedimentary strike and has no correlation with post-Cambrian structure. Further, the middle member is in most places overlain by several hundred meters of predominantly greenish shale, siltstone, and pale quartzite in the upper Wood Canyon, Zabriskie, and Carrara Formations. These rocks, as indicated by their color, are in general more reduced than the hematite-rich middle member. They are also not very permeable; indeed, they act as barriers for ground water movement in the present Basin and Range topographic regime (Winograd and Thordarson, 1968). A post-Early Cambrian oxidizing mechanism that would regionally penetrate these beds and oxidize the middle member, not just preferentially but exclusively, seems most implausible. Similar arguments apply to any oxidizing mechanism which might be postulated to have penetrated the strata from below. The Wood Canyon Formation lies in the middle part of a miogeoclinal sequence and was deeply buried throughout Paleozoic time; hence, near-surface oxidation at any time in the Paleozoic also is implausible. Together, these observations indicate that the bulk of the hematite of the middle member is an intrinsic feature of the unit, not a product of late secondary alteration.

The characteristic magnetization direction for the middle member (declination (D) = 124°, inclination (I) = -3°; Table 2.1) was calculated from the mode of directions from the fine-grained rocks (the red-purple mudstones and purplish siltstones) at the 640°C demagnetization step. This mode is probably representative of the characteristic magnetization direction, since sample directions at higher demagnetization temperatures no longer show consistent changes but abruptly scatter. However, the existence of only one polarity of the characteristic magnetization removes an independent check on whether all secondary components have been removed.

The polarity of the characteristic magnetization direction of the middle member is interpreted as reversed, based on its proximity to the reversed direction (D = 93°, I = -4°) from the Lower and Middle Cambrian Tapeats Sandstone of the Grand Canyon (Elston and Bressler, 1977). Virtually all discrete beds of fine-grained rock thicker than 2 cm exposed at section M have been sampled, with the average sampling interval approaching 1 m at the top and base of the member, where these rock types are more abundant. Fine-grained beds are rare in the middle of the member, but the magnetization of even the coarser-grained quartzites, which were sampled at about 3 m intervals, is of reversed polarity. Thus, any intervals of normal polarity that might be present in the middle member of the Wood Canyon Formation would have to be thin, and probably none exist.

Table 2.1 EARLY AND MIDDLE CAMBRIAN PALEOMAGNETIC POLES FROM THE DESERT RANGE, NEVADA

Formation	Section (Fig.2.1)	Mean Attitude (str.,dip)	$\overline{J}_{mode}$[§] (Am^2/kg $\times 10^{-7}$)	$\overline{J}_{mode}/\overline{J}_{NRM}$	Geog. Coords. (D, I)[†]	Strat. Coords. (D, I)[††]	β_{95}[¶] $\left(\frac{N_{mode}}{N_{tot.}}\right)$	Apparent North Pole (Lat,Long)[Ψ]	Corrected North Pole (Lat,Long)[δ]
Group II[*]	O	350, 47E	1.2	0.09	129,+15	130,-18	2°(47/76)	37N, 136E	9N, 160E
Group I[**]	N,O	351, 43E	0.4	0.05	131,+15	130,-14	4°(26/48)	36N, 134E	7N, 158E
Wood Canyon[***]	M	356, 58E	40.8	0.49	139,+40	124,-3	2°(56/122)	28N, 134E	1S, 157E

[*] Directions from gray, rusty-mottled limestone of the Jangle and Desert Limestone Members of the Carrara Formation, and lowermost Papoose Lake Member of the Bonanza King Formation.

[**] Directions from gray, rusty-mottled limestone of the Gold Ace, Red Pass, and Pahrump Hills Members of the Carrara Formation.

[***] Directions from red-purple mudstones and purplish siltstones of the middle member.

[§] $\overline{J}_{mode}$ is the geometric mean of the specific intensities of magnetization at the demagnetization step(s) upon which the mode is based (see text).

[†] Geographic coordinates are the mode of specimen directions uncorrected for the bedding attitude.

[††] Stratigraphic coordinates are the mode of specimen directions after correction for local bedding attitudes, not for the mean attitude of the entire formation.

[¶] β_{95} is an estimate of the 95% confidence limits for the mode. The statistic β_{95} is derived from the largest subset (N_{mode}) of the total sample (N_{tot}) that has a mean identical with the mode of the total sample. This statistic is defined as the Fisherian half-angle of the cone of 95% confidence for the mean of this subset.

[Ψ] Apparent north paleomagnetic pole calculated from stratigraphic coordinates before correction for the 36° clockwise rotation with respect to the Colorado Plateau (see text).

[δ] North paleomagnetic pole calculated from stratigraphic coordinates after correction for the 36° clockwise rotation with respect to the Colorado Plateau (see text).

Lower and Upper Members, Wood Canyon Formation

The directions of magnetization of about two-thirds of 42 samples taken from the lower member are much more scattered than the directions found in the middle member; in addition, intensities of magnetization are about an order of magnitude weaker. Moreover, the directions of most samples change dramatically and apparently randomly above about 600°C. Commonly, a large component of magnetization roughly aligned with the present axial field is present, particularly in the more weakly magnetized samples.

Upon thermal demagnetization, directions from the remaining third of the samples from the lower member are distributed between the characteristic magnetization direction of the middle member and the direction of the present axial field. These samples were taken from an approximately 70 m stratigraphic interval in the middle of the lower member (Figure 2.2). The rock types present in this interval are mostly green siltstone with purplish laminae, interbedded with light-colored to purplish quartzites. Directions of magnetization similar to the characteristic magnetization of the middle member are observed only in the purplish rocks. The characteristic magnetization direction of this interval is poorly defined, apparently owing to the presence of a significant present-axial-field component even at the highest temperature step (640°C). In this respect, these samples resemble the quartzites from the middle member. Although the direction of characteristic magnetization of the middle interval of the lower member could not be precisely determined, the polarity of all 15 samples from this interval is unambiguously reversed.

In samples from the upper member of the Wood Canyon, as in those from much of the lower member, intensities of magnetization are weaker than for the middle member, directions are scattered and change erratically upon demagnetization, and a present-axial-field component predominates.

Zabriskie Quartzite

At most localities in the southern Great Basin, the contact of the Zabriskie Quartzite with the underlying Wood Canyon Formation is gradational. In the Desert Range, however, the contact may represent a depositional hiatus (Stewart, 1970), since it is sharp and the Zabriskie is exceptionally thin. The Zabriskie, as Stewart recognized it in the Desert Range, is a fine-grained, pinkish-gray, even-bedded quartzite about 2 m thick. Palmer and Halley (in press) have reassigned beds gradational in lithology between the Zabriskie and the overlying Eagle Mountain Shale Member of the Carrara Formation to the Zabriskie, naming them the Emigrant Pass Quartzite Member of the Zabriskie Quartzite. The beds assigned to the Emigrant Pass are about 10 m thick in the Desert Range (Palmer and Halley, in press). In this study, Stewart's original definition of the Zabriskie will be used, and the Emigrant Pass Member will be discussed with the Carrara Formation.

Seventeen samples were obtained from the 2 m thick Zabriskie; this represents a higher sampling density than for any other formation in the Desert Range. These samples were progressively thermally demagnetized at 6 steps between 400 and 640°C; a few were intitially demagnetized at 300°C as well.

The NRM directions of about half the Zabriskie samples were near the direction of the present axial dipole field. These directions changed only slightly upon thermal demagnetization, even at the highest temperature steps. Nevertheless, vector demagnetization diagrams show that demagnetization paths for nearly all the samples do not trend toward the origin, but have declinations consistently displaced toward the northwest. Hence, these rocks contain some component of magnetization in addition to the component aligned roughly with the present axial field. This component is not well defined, but it could have a direction antipolar to the characteristic magnetization direction from the middle member of the Wood Canyon Formation.

In the remaining half of the Zabriskie samples, thermal demagnetization more convincingly revealed a component not aligned with the direction of the present axial field (Figure 2.4). The direction from one sample is consistently east to southeast with shallow inclination, near the characteristic magnetization direction from the middle member of the Wood Canyon. Directions from the other samples, stratigraphically both above and below this sample, generally trend toward a direction antipolar to the characteristic direction from the middle member. These data suggest that zones of both normal and reversed polarity may be present in the Zabriskie. Because the number of samples obtained was small, the direction of characteristic magnetization of the Zabriskie cannot be defined precisely.

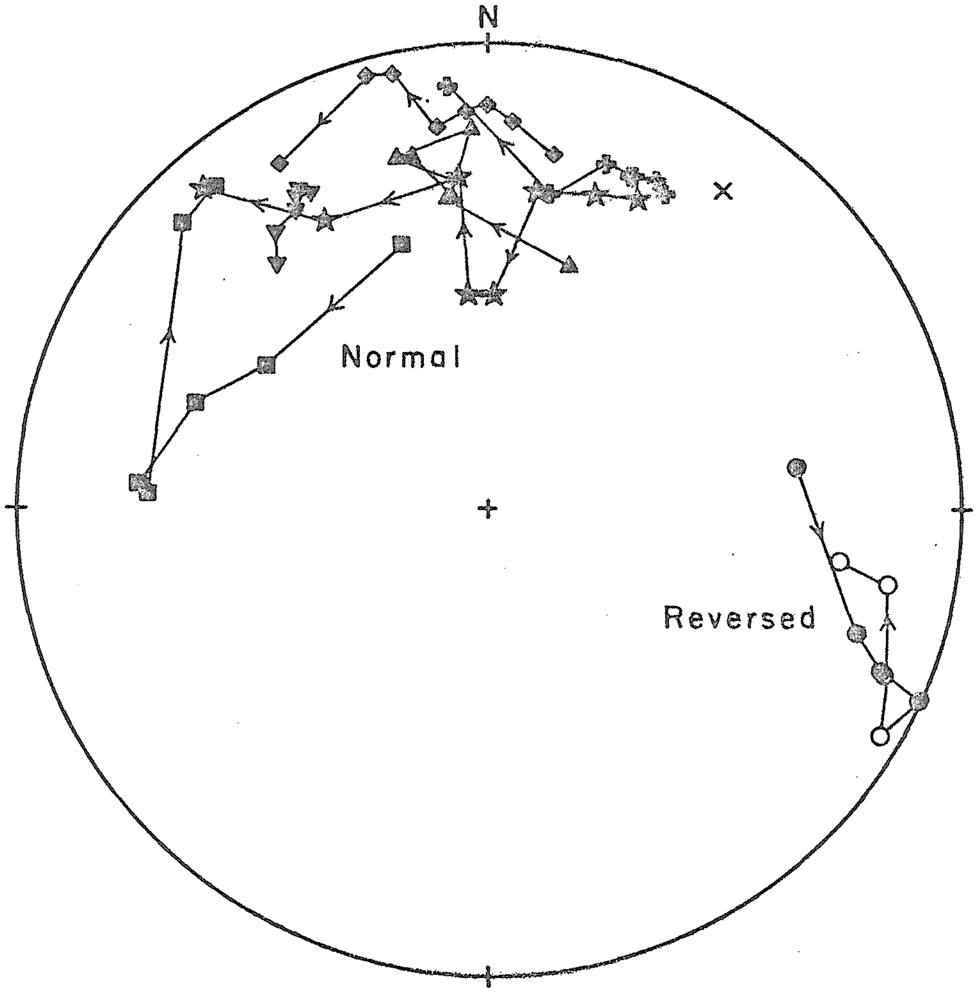


Figure 2.4. Progressive thermal demagnetization behavior of selected samples from the Zabriskie Quartzite, showing one reversed and 6 possibly normal-polarity directions. "X" is as in Figure 2.3. Shown are the NRM and demagnetization steps between 300 and 640°C. Arrows point to the higher-temperature steps. Each sample is represented by a different symbol. Open (shaded) symbols are on upper (lower) hemisphere, respectively.

Thin sections of 4 Zabriskie specimens were examined in both transmitted and reflected light. Specular hematite, comprising about 1% of the rock, is the dominant opaque mineral which is observed as discrete grains. These grains range from one micrometer to several micrometers across. As in the middle member of the Wood Canyon Formation, many of the grains are apparently detrital and can be readily distinguished from aggregates of hematite, also occurring in these rocks, which appear to have replaced ferroan precursors. Overall, the samples from the Zabriskie resemble the quartzites from the middle member of the Wood Canyon Formation. Perhaps, as suggested for those rocks, the NRM includes a DRM residing in specular hematite and an early-formed CRM, as well as a predominant CRM related to recent weathering.

Carrara and Lowermost Bonanza King Formations

The Carrara Formation of Early and Middle Cambrian age conformably overlies the Zabriskie Quartzite. In turn, the Carrara is overlain with a gradational contact by the Bonanza King Formation, an immense section of carbonate rock of Middle and early Late Cambrian age. Thus, deposition of the Carrara represents a transition from dominantly terrigenous sedimentation in the late Precambrian and Early Cambrian to carbonate sedimentation in later Paleozoic time.

Several sedimentary cycles, tens of meters thick, characterize the Carrara Formation (Palmer, 1971). Each cycle includes fine-grained, terrigenous rocks (siltstones and shales, with minor quartzite) in the lower part which grade to limestone in the upper part. The basal terrigenous rocks of the overlying cycle rest with a sharp contact on the limestone of the underlying cycle. On the basis of these cycles, Palmer and Halley (in press) have divided the Carrara into 9 members. The Eagle Mountain Shale and Thimble Limestone comprise the lower terrigenous and upper limestone parts of the first cycle, respectively. (In this study, data from Palmer and Halley's Emigrant Pass Member of the Zabriskie Quartzite will also be considered with the Carrara Formation.) The Echo Shale and Gold Ace Limestone define the lower and upper parts of a second cycle; the Pyramid Shale and Red Pass Limestone comprise a third cycle; and the Pahrump Hills Shale and Jangle Limestone make up a fourth cycle. The stratigraphically highest member is the argillaceous Desert Limestone, which forms the base of a fifth sedimentary cycle.

The Bonanza King Formation has been divided into a lower, Papoose Lake Member and an upper, Banded Mountain Member (Palmer, 1971). In the Desert Range, the Papoose Lake Member is about 550 m thick, and the Banded Mountain Member totals about 650 m (estimated from data in Palmer (1971)). Two short sections were sampled in the lower part of the Papoose Lake Member (Figure 2.2). The lower section immediately overlies the Desert Limestone Member of the Carrara Formation and roughly defines the top of the fifth large-scale sedimentary cycle. Seen from afar, this unit has a brownish cast which contrasts with dark gray limestone above; it is composed predominantly of gray, rusty-mottled limestone very similar to the limestones within the Carrara. The other section in the Bonanza King Formation was sampled in the dark gray limestone, about 80 m stratigraphically above the lower section.

The lower third of the Carrara Formation, up through the lowest part of the Pyramid Shale, contains latest Early Cambrian faunas (Palmer, 1971; Palmer and Halley, in press) and has been assigned to the Bonnia-Olenellus Zone (Nelson, 1978). Trilobites characteristic of the earliest Middle Cambrian are found in the lowest Red Pass Member, which has been assigned to the Plagiura-Kochaspis Zone (Palmer and Halley, in press). An Albertella-Zone fauna is found from the top of the Red Pass Limestone to the basal part of the Jangle Limestone. Trilobites representative of the Glossopleura Zone occur in the Desert Limestone Member of the Carrara Formation and in the immediately overlying section in the Papoose Lake Member of the Bonanza King Formation (Palmer, 1971); the higher section in the Papoose Lake probably also lies within the Glossopleura Zone (cf. Palmer, 1971).

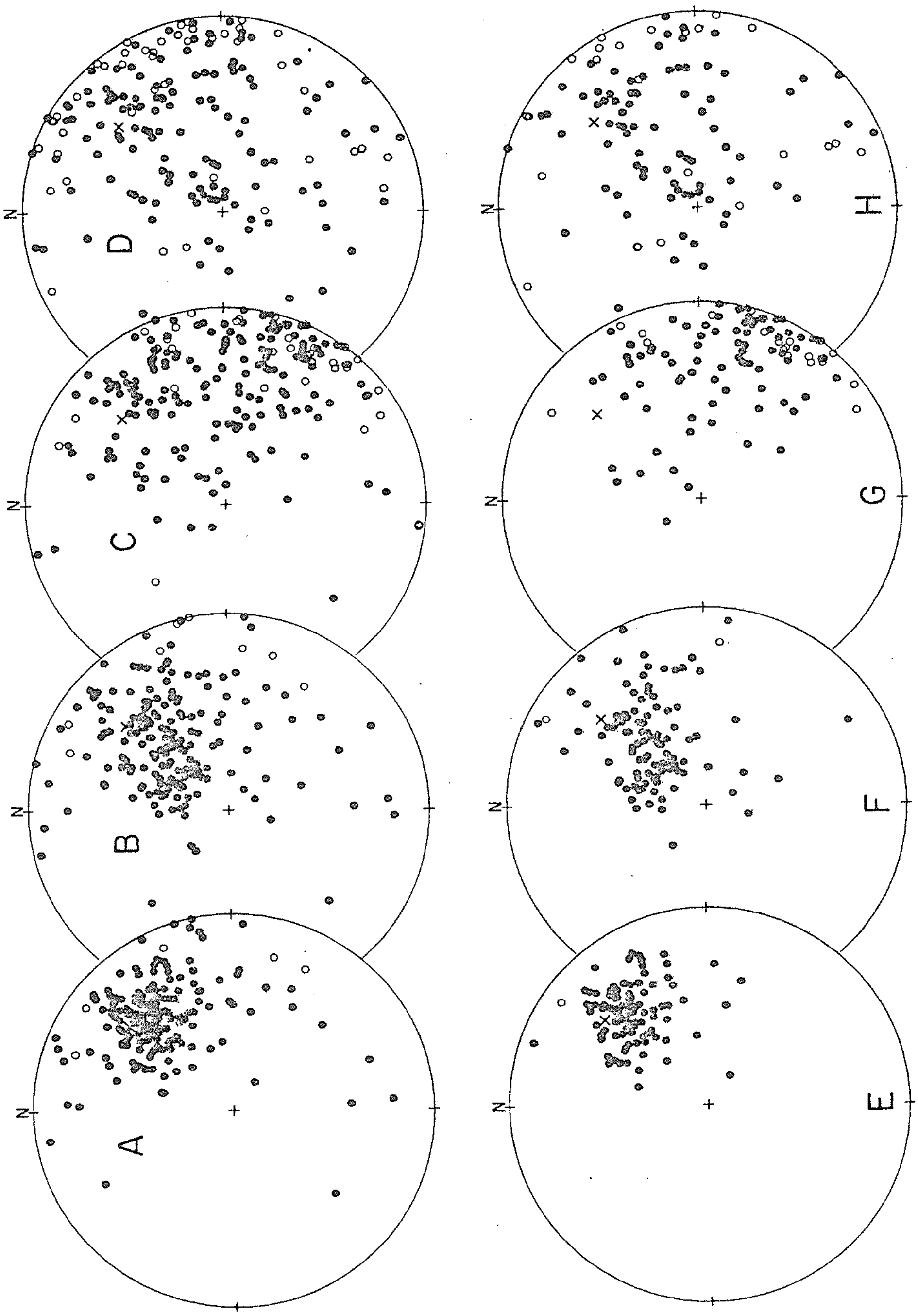
Initially, the entire Carrara Formation and the lower section in the Papoose Lake were sampled at intervals of 5 to 6 meters at sections N and O (Figure 2.1). About 75 samples were taken, representing a wide variety of rock types. The most promising paleomagnetic results came from gray, rusty-mottled limestones occurring in the limestone members of the Carrara and in the lower Papoose Lake section; hence, effort was concentrated on this rock type during resampling, when another 150 samples were obtained.

Demagnetization data from the strata sampled at sections N and O will be presented in two groups. This is convenient because (1) the demagnetization steps of the two groups are somewhat different; (2) the magnetic behavior of the two groups, while similar, differs slightly; and (3) paleomagnetic poles are calculated differently from the two data sets. Group I comprises all the Carrara Formation up through the Pahrump Hills Shale. Group II comprises the Jangle and Desert Limestone Members of the Carrara Formation, and the lower section in the Papoose Lake Member of the Bonanza King Formation. All of the initial samples in Group I were progressively thermally demagnetized at 200, 300, 400, 500, and 550°C; a few with sufficient remaining intensity were further demagnetized at 590 and 630°C. The Group I resamples were also demagnetized at these steps, but with two additional steps at 350 and 450°C. The Group II samples were all demagnetized at 6 or 7 steps between 200 and 500°C; the initial samples had steps performed at 120 and 550°C as well.

To complement the thermal demagnetization experiments, 13 duplicate specimens, all from Group I, were progressively AF demagnetized in 8 steps at peak fields between 5 and 150 mT. For all rock types, AF demagnetization generally resulted in a loss of intensity ranging from as little as 20% to over 90%, with little change from an NRM direction near that of the present axial field. Gray, rusty-mottled limestones and buff carbonates showed the least response to AF demagnetization; they lost 25-40% of their remanence in the first few tens of millitesłas and little thereafter. AF demagnetization alone provided no information regarding any characteristic magnetization that might be present in these rocks.

In contrast, progressive thermal demagnetization revealed the presence of several components of magnetization in the Group I samples (Figure 2.5). The NRM directions were clustered near the present axial field direction. At 200°C, most directions steepened slightly in

Figure 2.5 Directions at selected steps from progressive thermal demagnetization of samples from Group I (Carrara Formation up through Pahrump Hills Shale Member). Symbols are as in Figure 2.3. Directions from all samples are shown in "A," "B," "C," and "D" (top row): (A) NRM; (B) 200°C; (C) 400°C; (D) 550°C. "E," "F," "G," and "H" show the directions from gray, rusty-mottled limestones only (bottom row): (E) NRM; (F) 200°C; (G) 400°C; (H) 550°C.


A
B
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G
H
N

inclination and moved slightly counterclockwise; in addition, their distribution became somewhat more scattered. At 400°C, the directions of many samples moved toward a low, negative inclination and a declination near 130°. Finally, by 550°C, the component of magnetization with low inclination and declination near 130° was almost entirely removed.

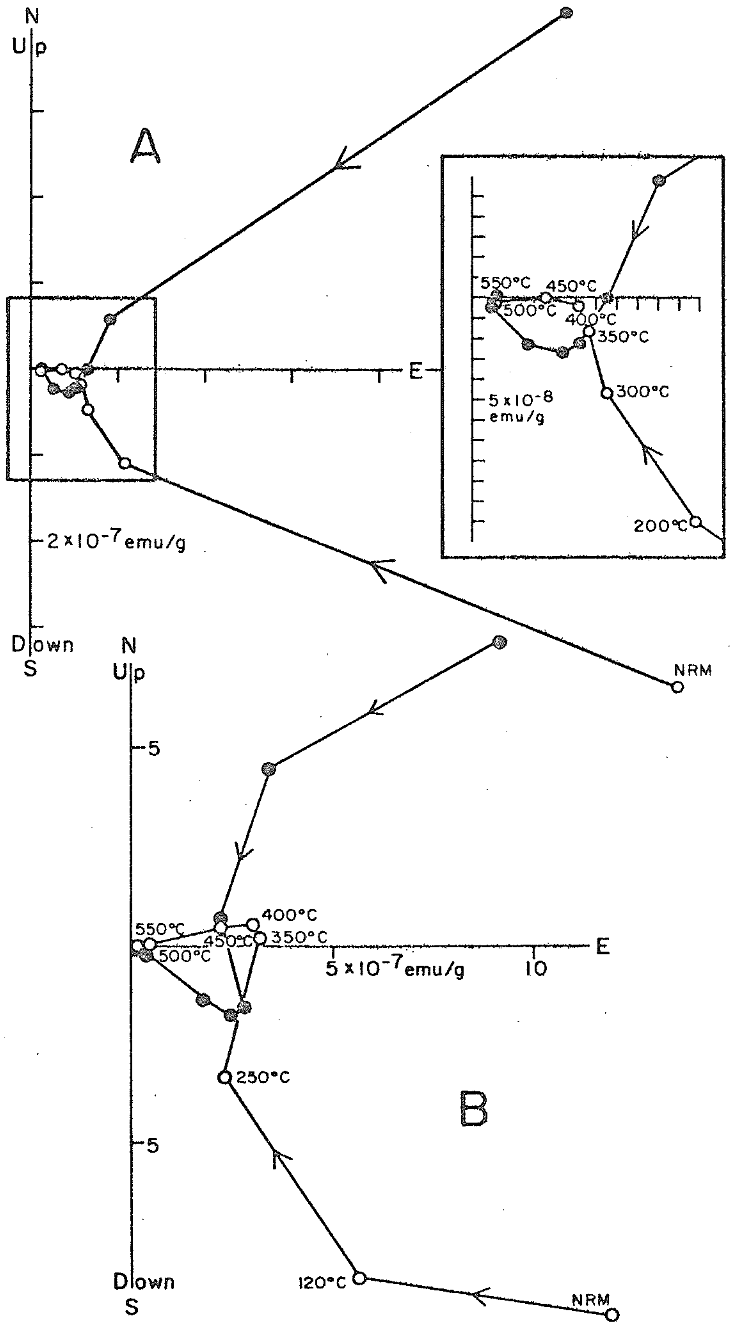
In detail, the behavior of a given sample upon thermal demagnetization was a function of its lithology. Directions of magnetization from gray, rusty-mottled limestones of the Thimble, Gold Ace, Red Pass, and interbeds in the Pahrump Hills Members most consistently showed the demagnetization behavior outlined above: i.e., a shift from NRM directions near that of the present axial field toward a direction with low inclination and south-easterly declination by 400°C (Figure 2.5). Directions from some of the quartzite interbeds in the Eagle Mountain and Echo Shales are distributed like those from the gray, rusty-mottled limestones; however, a present-axial-field component was more dominant in the quartzites at all demagnetization steps. Directions from green and brown siltstones of the Carrara generally changed erratically upon demagnetization. Carbonate rocks that are buff or orange rather than predominantly gray showed present-field directions which did not change appreciably upon demagnetization.

A vector demagnetization diagram of a specimen of gray, rusty-mottled limestone from Group I emphasizes several distinctive features of the demagnetization behavior of this rock type (Figure 2.6A). First, about 80% of the NRM intensity is lost by 200°C, followed by a smaller loss to 400°C. Second, the direction of magnetization at 200°C is not on the great-circle path from NRM to 400°C, but is displaced toward a

Figure 2.6 Vector demagnetization diagrams of samples of gray, rusty-mottled limestone from the Carrara and lowermost Bonanza King Formations. The conventions used in constructing these diagrams are similar to those used by Roy and Park (1974). Shaded circles represent the endpoint of the magnetic vector projected onto the horizontal plane; the declination is the angle (measured clockwise from due north) between the north-south axis and the horizontal component. The distance from open circles to the origin is the total length of the magnetic vector; the angle between the abscissa and the ray from the origin through the open circle is the inclination. Note that open circles do not represent a projection of the magnetic vector onto a fixed plane. The demagnetization steps are labeled beside the points on the inclination path. (1 emu/g = 1 Am^2/kg)

(A) Behavior upon thermal demagnetization of a sample from the Gold Ace Limestone Member of the Carrara Formation (Group I).

(B) Behavior upon thermal demagnetization of a sample from near the base of the Papoose Lake Member of the Bonanza King Formation (Group II).


N
Up
A
E
2 x 10^-7 emu/g
Down
S
550°C
450°C
500°C
400°C
350°C
5 x 10^-8
emu/g
300°C
200°C
NRM
N
Up
5
550°C
400°C
450°C
350°C
500°C
5 x 10^-7 emu/g
10
E
250°C
B
5
Down
S
120°C
NRM

steeper inclination. Third, the demagnetization path between 400 and 500°C is essentially linear to the origin, indicating removal of a single component with moderately high blocking temperature. This demonstrates that at least 3 components of magnetization are present in these limestones: (1) a component which is roughly aligned with the direction of the present axial field and which has a low blocking temperature or chemical stability limit; (2) a component of intermediate thermal stability (present at $\sim$200°C but removed thereafter), contributing to a direction with a steep inclination; and (3) a characteristic component, isolated between 400 and 500°C, with 5-10% of the NRM intensity, a stratigraphic declination near 130°, and a shallow inclination. This last component is removed by 550°C.

The Group II samples showed magnetic behavior very similar to that of the gray, rusty-mottled limestones from Group I (Figure 2.7); the three components of magnetization are again present. (A few samples, taken from a 10 m thick whitish limestone unit in the Jangle Member, had essentially no magnetization remaining above 250°C; they are excluded from discussion and from Figure 2.7.) The rock types in Group II are also mostly gray, rusty-mottled limestones, so the similarity in behavior is not surprising; however, the characteristic magnetization was much more effectively isolated by thermal demagnetization in the Group II samples, as is shown by their better grouping at 400 and 450°C. This may be due to less deep weathering in the Group II samples, leading to less remagnetization. In general, the low-temperature, present-axial-field component in the Group II samples is much smaller, as shown by a vector demagnetization diagram of a typical Group II sample

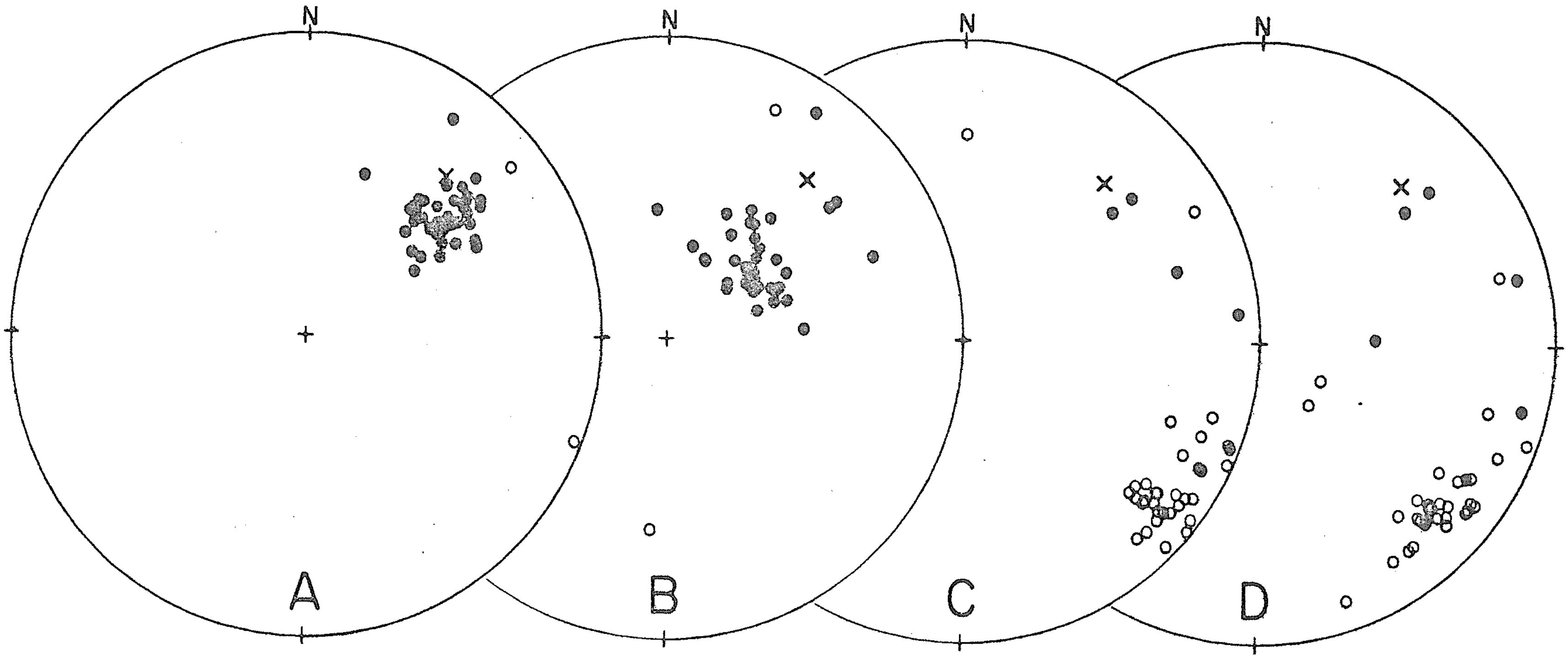


Figure 2.7 Progressive thermal demagnetization behavior of all samples from Group II (Jangle and Desert Members of the Carrara Formation, and lowermost Papoose Lake Member of the Bonanza King Formation). Symbols are as in Figure 2.3. (A) NRM; (B) 200 or 250°C; (C) 400°C; and (D) 450°C.

(Figure 2.6B); this lends some support to this idea. However, no difference in degree of weathering was apparent in the field between the Group I and Group II samples.

The component in both Groups I and II stable only at low temperature probably in part represents a CRM developed during recent weathering. This component may reside either in goethite (which has a Neel point near 120°C; e.g., Strangway (1970), p. 35) or perhaps in very fine-grained hematite; it is probably related to the rusty seams and mottling noted in hand specimen. The inital drop in intensity of magnetization upon AF demagnetization (in the Group I duplicate specimens), however, suggests that part of this component may be a viscous magnetization residing in magnetite.

The cluster of directions observed at 200-250°C in Groups I and II probably represents a vector sum of a component of intermediate thermal stability, the present-axial-field component, and the characteristic component. The exact direction of the component of intermediate thermal stability is difficult to determine because the relative intensities of all 3 components are not known. This component resembles a component with similar thermal stability found in upper Precambrian limestones at the base of the Desert Range section (Van Alstine and Gillett, unpublished data). The component with intermediate thermal stability is ascribed to the Sevier Orogeny of Late Cretaceous age.

Most important for interpreting the characteristic magnetization of both Groups I and II is the identification of the magnetic species in which this component resides. As the characteristic magnetization is removed above 500°C, it is possible that it is a DRM carried by magnetite. From the thermal demagnetization experiments alone, however, the possibility cannot be excluded that this component resides in hematite with a moderate blocking temperature. Accordingly, 25 duplicate specimens of gray, rusty-mottled limestones, all from Group I, were subjected to progressive AF demagnetization at peak fields up to 50 mT following thermal demagnetization at 410^{o}C (above the thermal stability range of the two apparently secondary components, but below that of the characteristic component). Vector demagnetization diagrams show that by 50 mT, AF demagnetization has removed about 75% of the characteristic magnetization present after thermal demagnetization at 410^{o}C (Figure 2.8). This suggests that the characteristic magnetization of these limestones resides in a species with a coercivity typical of magnetite and may well be a DRM. In addition, the specimen directions that were farthest from the present-axial-field direction at 410^{o}C became better grouped upon AF demagnetization (Figure 2.9).

Forty thin sections of gray, rusty-mottled limestone, mostly from the Red Pass and Pahrump Hills Members of the Carrara Formation and including a few specimens which had been thermally demagnetized at 410^{o}C, were examined in transmitted and reflected light. About 10 thin sections were also examined under cathodoluminescence. In addition, polished chips of about 90% of the samples from the Red Pass Member of the Carrara through the lowermost Papoose Lake Member of the Bonanza King Formation were examined under a binocular microscope.

Most of the rocks proved to be typical of the lime mudstone lithofacies of Palmer and Halley (in press), which makes up the greater part of the limestone members of the Carrara Formation. Fenestral fabric

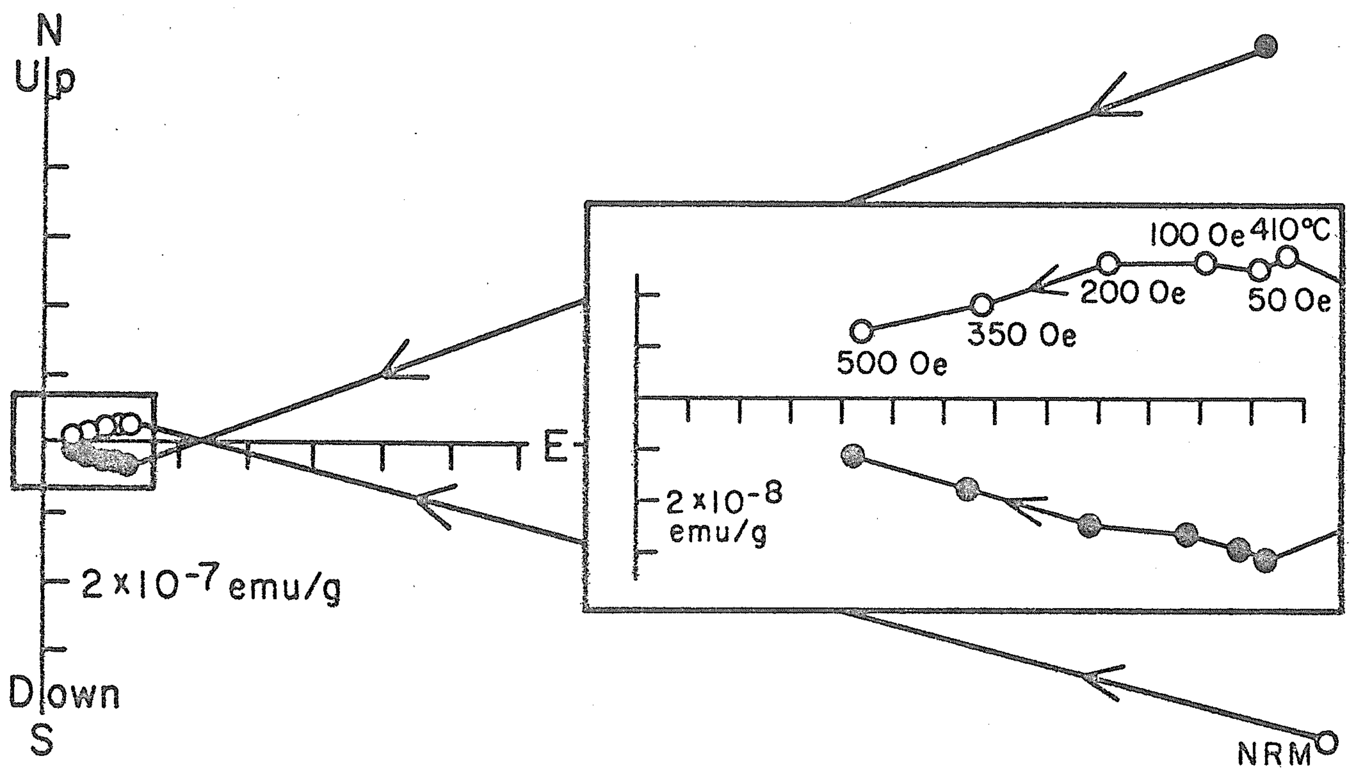


Figure 2.8 Vector demagnetization diagram showing the behavior upon AF-following-thermal demagnetization of a sample of gray, rusty-mottled limestone from the Pahrump Hills Member of the Carrara Formation (Group I). Conventions are as in Figure 2.6. (1 Oe = 0.1 mT ; 1 emu/g = 1 Am^2/kg)

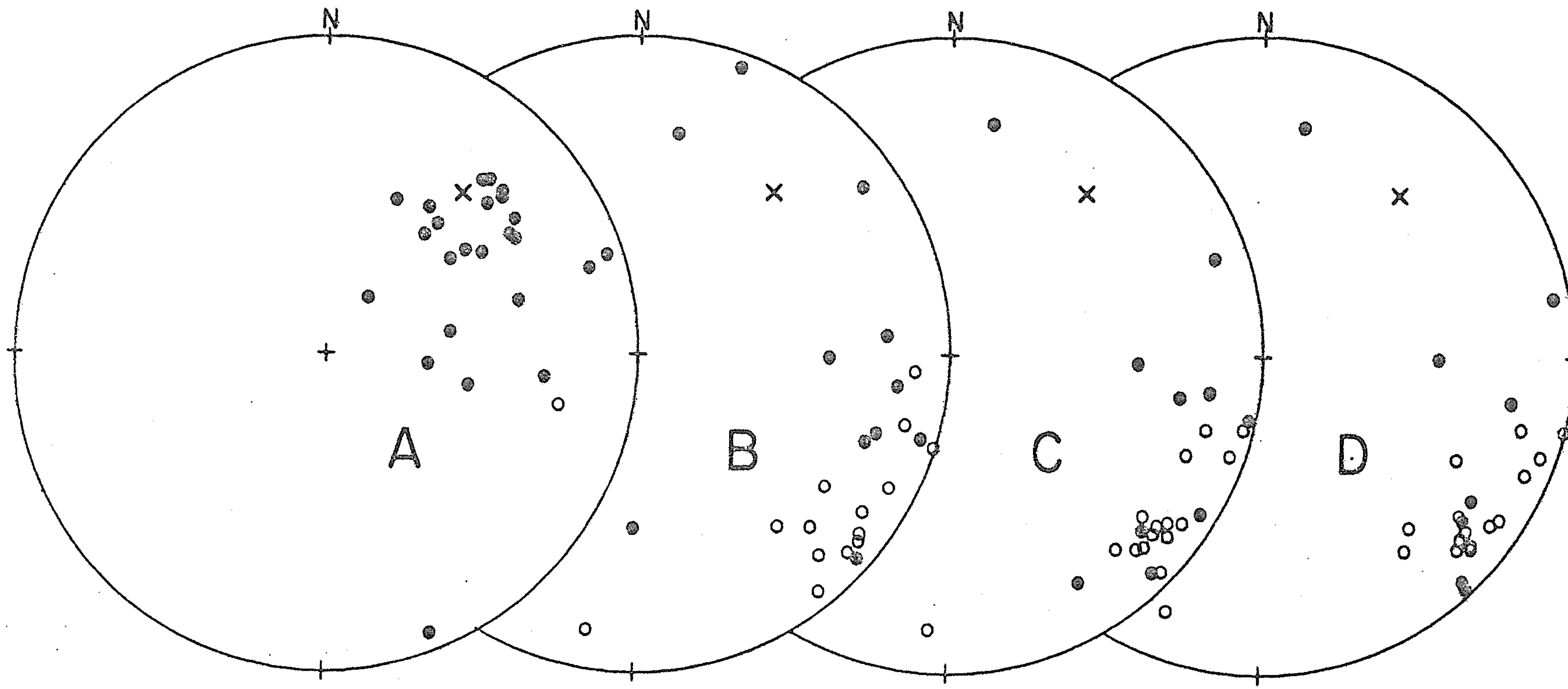


Figure 2.9 Directions from combined AF-following-thermal demagnetization experiments on duplicate specimens of gray, rusty-mottled limestone from Group I (Carrara Formation up through Pahrump Hills Shale Member). Symbols are as in Figure 2.3. (A) NRM; (B) 410^{o}C; (C) 410^{o}C + 20 mT; (D) 410^{o}C + 35 mT.

boundstones (characteristic of Palmer and Halley's algal boundstone lithofacies) and oolites are less common rock types; one intraclast conglomerate was also noted. No systematic correspondence was found between magnetic behavior and the specific limestone type. Staining of all rock types with slightly acidified solutions of alizarin red S (for calcite) and potassium ferricyanide (for Fe) (e.g., Wolf et al., 1967) shows the rocks to be composed predominantly of non-ferroan calcite. No fabric destruction was apparent in the samples that had undergone thermal demagnetization to 410°C; the only observed effect was to redden the orange seams discussed below, presumably from the dehydration of goethite to form hematite.

The rusty seams and mottling which lend these rocks their distinctive appearance in hand specimen are composed of cryptocrystalline, orange to red-orange iron oxides and oxyhydroxides ("limonite") which are probably dominantly goethite. The limonite seams contain occasional red streaks of hematite and abundant rhombs of dolomite. In some specimens, peloids or ooids are also commonly replaced by limonite. Palmer and Halley (in press) presented evidence suggesting that the oxidation of these rocks occurred late in their history. Hence, the limonitic veins may well contain the magnetic species carrying the components of both low and intermediate thermal stability.

Examination of lightly etched surfaces on the thin sections under reflected light showed common grains of insoluble minerals standing up in relief. Quartz and feldspar are most abundant, but some hematitic patches also occur. The euhedral form of most of the feldspars suggests they are authigenic. Under cathodoluminescence, however, many proved to have rounded, blue-luminescing cores. Such blue luminescence is due to a small amount of Ti^{4+} in the lattice (Mariano et al., 1973) and is diagnostic of a high-temperature feldspar (Kastner, 1971). Thus, some of the feldspar grains contain detrital cores. The occurrence of such conclusively detrital material in these limestones makes it likely that a small component of detrital magnetite could also be present.

The characteristic magnetization direction for the Group I samples (Table 2.1) was determined from specimens that had been subjected to AF-following-thermal demagnetization (Figure 2.9). These demagnetization experiments yielded distributions of directions that are less biased toward the present axial field direction than the distributions obtained from thermal demagnetization alone. The characteristic magnetization direction represents the mode of combined specimen directions from the 20 and 35 mT steps. This procedure, which gives greater weight to those samples with more stable directions, was followed because the directions are about equally well grouped at both demagnetization steps. Axial symmetry of the highest probability contours around the mode, and the consistency of this direction at two demagnetization steps, suggest that the direction of characteristic magnetization has been isolated.

The characteristic magnetization direction of the Group II samples (Table 2.1) was determined from the mode of directions combined from the 400 and 450°C demagnetization steps. Like the procedure employed above, this gives greater weight to those samples with more stable directions; it was adopted because the distributions of directions from both steps are about equally well grouped. The characteristic magnetization direction of the Group II samples is within 5° of that for Group I.

Secondary components of magnetization were completely removed from relatively few of the Group I samples. However, analysis of vector demagnetization diagrams of all 185 samples permitted polarity assignments for about 60% of them. All samples that exhibited reasonably smooth thermal demagnetization paths are of reversed polarity; in addition, all Group II samples not showing only a present-field direction are unambiguously reversed. Hence, the characteristic magnetization of the entire Carrara Formation and the lowermost Bonanza King may be entirely reversed (Figure 2.2). As the terrigenous units in the lower part of the Group I strata (Eagle Mountain and Echo Shale Members of the Carrara Formation, and Resting Springs Member of the Zabriskie Quartzite) exhibit mostly scattered or present-axial-field directions, however, their polarity has been shown as "not determinable" in Figure 2.2; intervals of normal polarity up to ∿10 m thick might have been missed in these units.

The Bonanza King limestones from the stratigraphically higher section P not only are lithologically different from those at section O, but also have very different magnetic characteristics. Section P contains dark gray to blue-gray limestone with common sparry calcite veins and seams. The 14 samples collected from this section had scattered NRM directions. Progressive AF demagnetization to peak fields of 50 mT in general caused the directions to cluster more closely around the present axial field direction, with losses of magnetic intensity ranging from 50-90%. Moreover, progressive thermal demagnetization of 5 duplicate specimens yielded scattered directions. Hence, no paleomagnetic information was obtained concerning any characteristic magnetization that might be present in this part of the Papoose Lake Member.

Discussion

The evidence presented above suggests that the characteristic magnetizations of the Desert Range Cambrian rocks were acquired penecontemporaneously with deposition. Additional support for penecontemporaneous acquisition is provided by the similarity of characteristic magnetization directions recorded by two apparently different magnetic species (in the Wood Canyon red-purple mudstones versus the gray, rusty-mottled limestones of the Carrara and Bonanza King). Because the Desert Range lies in a structurally complex region, however, caution must be exercised in interpreting paleomagnetic poles computed from these characteristic magnetizations. Since the rocks of the Desert Range may have been rotated about a vertical axis, poles derived from these rocks will not necessarily coincide with poles from rocks of equivalent age on the craton. In principle, the amount of vertical-axis rotation of the Desert Range can be determined by calculating the declination correction required to make paleomagnetic poles from these rocks match poles of the same age from the craton. The success of this technique depends upon the availability of accurate, well-dated Early to Middle Cambrian poles from the craton.

Most of the published paleomagnetic poles from North America that may represent Early and Middle Cambrian time are either insufficiently accurate or too imprecisely dated to be useful for comparison with the Desert Range Cambrian results (Table 2.2, Figure 2.10). Poles 1 and 2, from extrusive and hypabyssal rocks of the Franklin Magnetic Interval

Table 2.2

POSSIBLY EARLY AND MIDDLE CAMBRIAN PALEOMAGNETIC POLES FROM NORTH AMERICA

Pole No.	Pole Position	Rock Unit, Location	Age	Reference
14	37N, 122E	Intrusives, Colorado	Late Precambrian-Early Ordovician (704-485 m.y.B.P.)	Larson and Mutschler, 1971
13	44N, 100E			French *et al.*, 1977
12	15N, 142E			
11	5N, 174E			
10	48N, 107E			
9	59N, 89E	Abrigo Fm., Arizona	Middle to Late Cambrian	Elston and Bressler, 1977
8	55N, 110E	Muav Fm., Arizona	Middle to Late Cambrian	Elston and Bressler, 1977
7	38N, 144E	Rome Fm., Tennessee	Middle Cambrian	French, 1976
6	13N, 146E	Ophiolite complex, Que.	Middle(?) Cambrian (550(?) m.y.B.P.)	Seguin, 1976
5	5N, 158E	Tapeats Sandstone, Arizona	Early to Middle Cambrian	Elston and Bressler, 1977
4	29N, 167E	Bradore Fm., W. Newfoundland	Early Cambrian	Rao and Deutsch, 1976
3	5N, 172E	Cloud Mtn. Basalt, W. Newfoundland	Late Precambrian-Early Cambrian (605±10 m.y.B.P.)	Deutsch and Rao, 1977
2	4S, 161E	Franklin rocks, Can. Arctic (Group A)	Late Precambrian (675-625 m.y.B.P.)	Palmer and Hayatsu, 1975
1	8N, 166E	(Group B)		

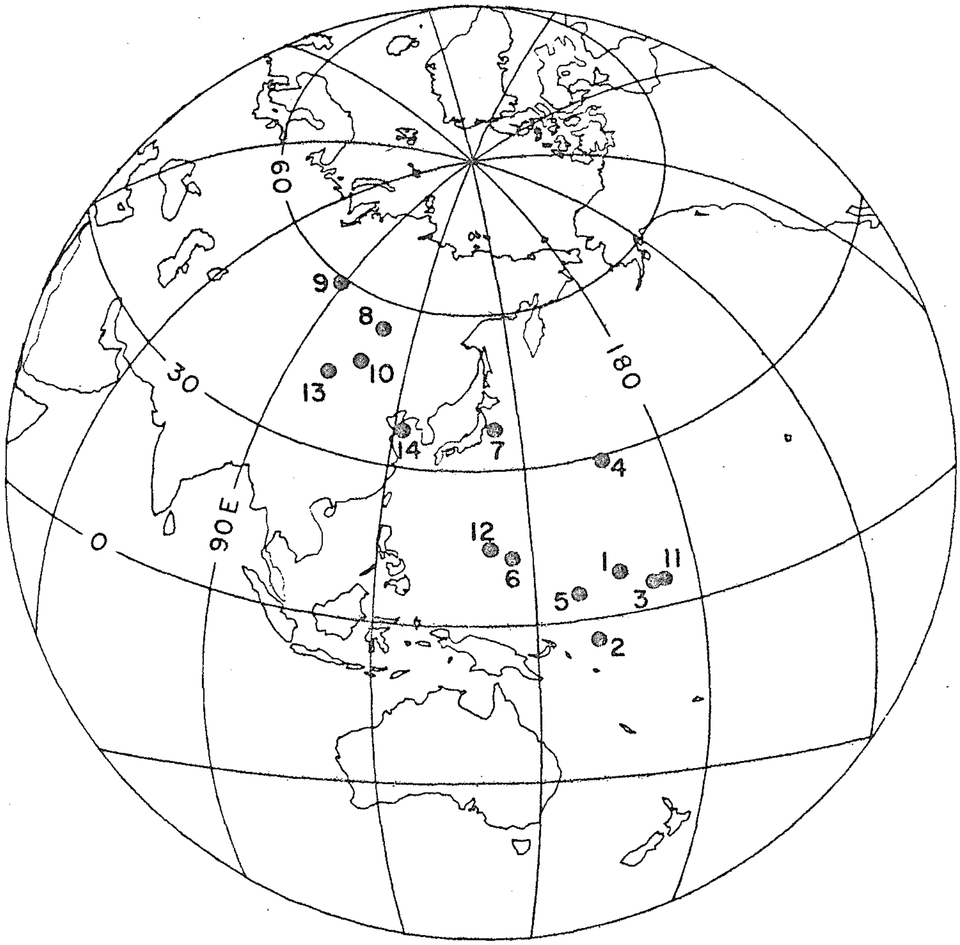


Figure 2.10 Published, possibly Early and Middle Cambrian paleomagnetic poles from North America. Pole numbers are as in Table 2.2.

in the Canadian Arctic, have a reported age of 625 m.y.B.P. from the least altered lavas (Palmer and Hayatsu, 1975). Earlier workers (Fahrig et al., 1971), however, preferred an age of 675 m.y.B.P. for these rocks. Even the younger age seems somewhat too old for middle Wood Canyon time, as the base of the Cambrian is probably in the range 590 ± 20 m.y.B.P. (Lambert, 1971). Pole 3 from 3 flows of the Cloud Mountain Basalt, western Newfoundland, may also be older (605 ± 10 m.y.B.P.) than middle Wood Canyon time and may have incompletely averaged out secular variation (Deutsch and Rao, 1977). In addition, the possibility of a significant rotation of western Newfoundland with respect to cratonic North America has been the subject of a longstanding controversy (cf. Deutsch and Rao, 1977). This uncertainty also applies to pole 4, from the Early Cambrian Bradore Formation of western Newfoundland. Moreover, the attitude correction upon which the Bradore pole is based is ambiguous (Rao and Deutsch, 1976). Pole 6, from a Middle(?) Cambrian ophiolite complex in Quebec (Seguin, 1976), is derived from rocks in a structurally complex area and may reflect a significant vertical-axis rotation. Pole 7, from the Rome Formation, Tennessee, is from a thrust sheet, which may also have rotated about a vertical axis (French, 1976). Although poles 8 and 9 from the Muav and Abrigo Formations, Arizona, are from rocks in structurally simpler settings, on faunal evidence these strata are younger (latest Middle and earliest Late Cambrian) than early Bonanza King time and, in addition, may have been remagnetized (Elston and Bressler, 1977). Intrusives from Colorado studied by Larson and Mutschler (1971) and French et al. (1977) yield ages ranging from 704 to 485 m.y.B.P. and give multimodal distributions of paleomagnetic directions (poles 10 through 14).

Of the poles listed in Table 2.2, that from the Tapeats Sandstone of the Colorado Plateau (pole 5) is the most relevant for comparison with the Desert Range Cambrian paleomagnetic results. The Tapeats pole, which has an α_{95} of 3.3°, is based on 129 samples representing antipolar normal and reversed directions. The Tapeats is unfossiliferous except locally at its top in fine-grained beds gradational in lithology with the overlying Bright Angel Shale. In the westernmost exposures of a typically developed Tapeats section at Frenchman Mountain, Nevada, such transitional beds yield trilobites of latest Early Cambrian age, equivalent to those in the lower Pyramid Shale Member of the Carrara (A. R. Palmer, written communication). The Tapeats, however, becomes younger west to east so that its upper part is probably early Middle Cambrian in the eastern Grand Canyon (McKee, 1945), where the pole of Elston and Bressler (1977) was derived. This implies that the section represented by the Tapeats pole should be partly older than and partly contemporaneous with the section represented by the Group I pole from the Carrara Formation, as the Group I pole was derived from rocks that are mainly early Middle Cambrian in age. Moreover, on the basis of lithologic comparison and regional stratigraphic relationships, Stewart (1970) has suggested that the Tapeats is the rock-stratigraphic equivalent of the Zabriskie Quartzite and upper part of the Wood Canyon Formation; therefore, the entire Tapeats almost certainly is younger than the middle member of the Wood Canyon. In addition, preliminary paleomagnetic results from the Bright Angel Shale, which conformably overlies the Tapeats, suggest that a pole from the Bright Angel would be similar to that from the Tapeats (Elston and Bressler, 1977). The Bright Angel extends into the Middle Cambrian Glossopleura Zone at the top (McKee, 1945) and therefore is partly contemporaneous with the upper Carrara and lowermost Bonanza King Formations. In summary, lithostratigraphic and biostratigraphic evidence strongly indicate that the age of the Tapeats is bracketed by the Desert Range strata.

However, poles computed from the characteristic magnetization directions of the Early and Middle Cambrian strata from the Desert Range differ by about 35° from the Tapeats pole (Figure 2.11). The direction of characteristic magnetization for the Tapeats ($D = 94^{\circ}$, $I = -6^{\circ}$) differs almost entirely in declination from the directions for the middle Wood Canyon ($D = 124^{\circ}$, $I = -3^{\circ}$) and the Carrara Group I samples ($D = 130^{\circ}$, $I = -14^{\circ}$). Hence, the discordance between Cambrian poles from the Desert Range and the pole from the Tapeats can be explained by clockwise rotation of the entire Desert Range Cambrian section. A 36° counterclockwise rotation of the Desert Range declinations will place the pole from the Carrara Group I samples (7°N, 158°E) and that from the Wood Canyon (1°S, 157°E) on either side of the pole from the Tapeats (5°N, 158°E). A vertical-axis rotation of this magnitude is entirely consistent with the structural evidence (summarized in the discussion of the regional geology) for clockwise oroflexural bending during the Cenozoic in the southern Great Basin. Because the Desert Range is part of a thrust plate, however, some component of the net 36° clockwise rotation may have resulted from thrusting in the Mesozoic. Significant vertical-axis rotations have been documented by paleomagnetic investigations in other thrust systems of the Cordillera and have amplitudes up to 60° in

Figure 2.11 Cambrian paleomagnetic poles from the Desert Range before (unprimed symbols) and after (primed symbols) correction for the proposed 36° clockwise vertical-axis rotation of the Desert Range section with respect to the Colorado Plateau. II: Group II (uppermost Carrara and lowermost Bonanza King Formations; I: Group I (remainder of Carrara Formation); WC: Wood Canyon Formation. Star labeled "T" is the pole from the Tapeats Sandstone of the Colorado Plateau (Elston and Bressler, 1977). Cross labeled "DR" marks the sampling site. Also shown is the post-Early Ordovician part of the apparent polar wander path for North America determined in Chapter 4. Numbers beside the path refer to nearly uniform, 22-m.y. time intervals: 1-3, Cenozoic; 4-7, Cretaceous; 7-9, Jurassic; 10-11, Triassic; 12-13, Permian; 14-15, Pennsylvanian; 16-17, Mississippian; 18-19, Devonian; 20-21, Silurian; 21-22, Ordovician.

Wyoming (Grubbs and Van der Voo, 1976).

The polarity of the characteristic magnetizations of the Desert Range Cambrian rocks is consistent with the polarity zonation observed in Cambrian rocks of the Colorado Plateau. Elston and Bressler (1977) have determined that several zones of reversed and normal polarity are represented in the lower and middle parts of the Tapeats Sandstone, whereas the uppermost Tapeats and overlying Bright Angel Shale seem to be entirely reversely magnetized. As discussed previously, the Tapeats is possibly the time equivalent of the upper Wood Canyon through lowermost Carrara Formations. Perhaps the zones of normal polarity inferred in the Zabriskie correlate with some of those found in the Tapeats, and the long reversed interval in the Carrara and lowermost Bonanza King correlates with the reversed interval in the uppermost Tapeats and Bright Angel. The reversed interval in the middle member of the Wood Canyon occurs in rocks that are probably older than the basal Tapeats (Stewart, 1970).

The Cambrian polarity history beginning to emerge from paleomagnetic studies in the southwestern United States is supported by results from other regions. Van Alstine (unpublished data) has found only reversed polarity in rocks of the Canadian Cordillera spanning most of Middle Cambrian time. In addition, Irving and Pullaiah's (1976) synthesis of worldwide polarity data suggests that the geomagnetic field during most of the Cambrian was characterized by a strong reversed-polarity bias.

Conclusions

Paleomagnetic evidence suggests that part of the Desert Range has undergone 36° of net clockwise rotation with respect to the Colorado Plateau. The existence of such large vertical-axis rotations implies that paleomagnetic results from tectonically complicated areas such as the southern Great Basin must be treated with discretion. However, such results also demonstrate the value of paleomagnetism in establishing constraints on structural geologic problems.

The similarity of Early and Middle Cambrian poles from the Desert Range strata strongly suggests that the paleogeographic pole was relatively stationary with respect to North America between middle Wood Canyon (early Early Cambrian) and early Bonanza King (middle Middle Cambrian) time.

A distinctive Cambrian polarity zonation seems to be emerging from paleomagnetic investigations of Early Paleozoic rocks. A change from mixed to dominantly reversed polarity appears to have occurred near the Early-Middle Cambrian boundary and may form a very useful worldwide chronologic horizon.

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Chapter 3

PALEOMAGNETISM OF UPPER PRECAMBRIAN SEDIMENTARY ROCKS FROM THE DESERT RANGE, NEVADA

In a previous paper (Chapter 2), we described the motivation for and the methodology of a systematic paleomagnetic investigation of upper Precambrian through Lower Paleozoic miogeoclinal strata in the Desert Range, Nevada. In that paper, we presented paleomagnetic data from the Lower and Middle Cambrian parts of this section. Beneath these Cambrian beds lie nearly 3000 m of apparently conformable strata, which comprise the Johnnie Formation (upper Precambrian) and Stirling Quartzite (uppermost Precambrian and possibly Lower Cambrian). In this companion paper, we present paleomagnetic results from the uppermost Johnnie Formation and Stirling Quartzite, in order to extend the North American apparent polar wander path into latest Precambrian time and to investigate the polarity history of the geomagnetic field near the Precambrian-Cambrian boundary.

Uppermost Johnnie Formation (Rainstorm Member)

The Johnnie Formation (upper Precambrian) of the southern Great Basin is a heterogeneous sequence of shale, siltstone, quartzite, limestone, and dolomite. In the Desert Range, where the Johnnie attains its maximum known thickness of more than 1500 m, Stewart (1970) has recognized 5 members, the uppermost of which is the Rainstorm Member (Figure 3.1). At a locality about 5 km northwest of that sampled in this study, Stewart and Barnes (1966) and Stewart (1970) have measured and described 3 subdivisions of the Rainstorm Member. In ascending stratigraphic order they are: (1) a siltstone unit, 23 m thick, consisting predominantly of

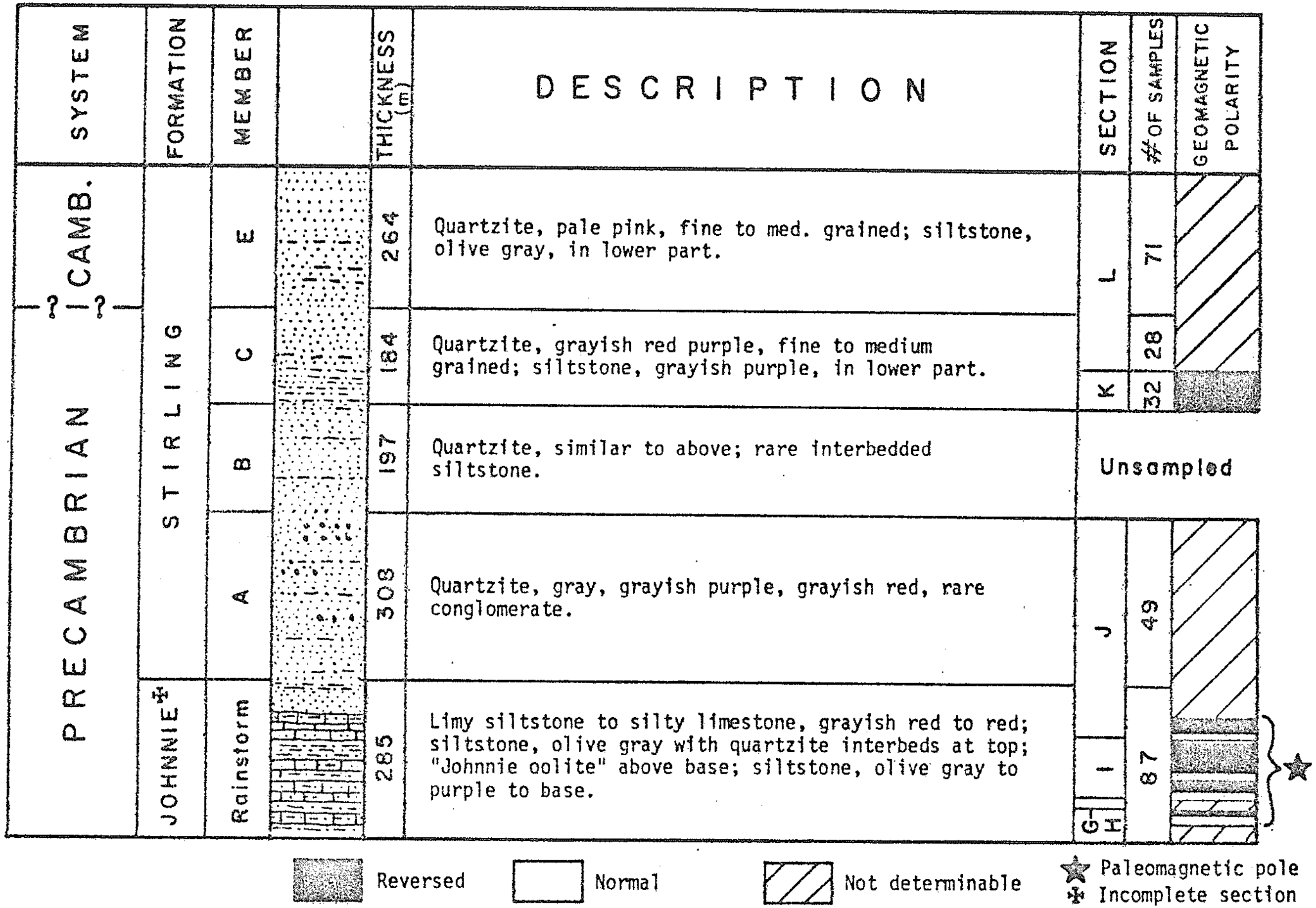

Figure 3.1 Chart showing description of strata, stratigraphic distribution of samples, and inferred geomagnetic polarity for upper Precambrian and Lower(?) Cambrian formations sampled in this study. Descriptions are adapted from Stewart and Barnes (1966). The beds at the top of this figure are conformable with those at the bottom of Figure 2.2.

grayish-olive to purple siltstone and including a distinctive, 1.8 m thick oolitic dolomite (the "Johnnie oolite"); (2) a carbonate unit, 198 m thick, consisting of generally pale red to grayish-red intergradational limestone and siltstone, containing ripple marks, drag marks, and flute casts; and (3) a siltstone and quartzite unit, 64 m thick, consisting of grayish-olive siltstone and pale yellowish-brown quartzite. The stratigraphic thicknesses of the 3 units of the Rainstorm at the locality sampled in this study are about 20% less than those reported at Stewart's (1970) Desert Range locality 70.

The presence of some minor faults and the incomplete exposure of some slope-forming beds necessitated our sampling the Rainstorm Member in 4 stratigraphically overlapping sections (G, H, I, and J, in Figure 2.1). Eighty-seven oriented samples were obtained from the Rainstorm Member, with the following stratigraphic distribution: 8 samples from the siltstone unit (6 samples of purple siltstone and 2 samples of the Johnnie oolite); 74 samples from the carbonate unit; and 5 samples from the siltstone and quartzite unit.

The 6 samples from the siltstone unit of the Rainstorm Member were progressively demagnetized at 6 steps between 400 and 640°C. Directions of natural remanent magnetization (NRM) of most of these samples were displaced from the direction of the present axial dipole field. At progressively higher temperatures, some of the directions became more westerly and some more easterly, although they did not become antiparallel. This demagnetization behavior suggests that, in addition to a component aligned roughly with the direction of the present axial field, the siltstone unit of the Rainstorm contains a two-polarity magnetization with a much different direction. This remanence, which represents the characteristic magnetization of these rocks, appears to have a shallow inclination and either easterly or westerly declination. The characteristic magnetization directions cannot be precisely defined, however, because a significant part of the present-axial-field component has a blocking temperature at least as high as that of the characteristic magnetization.

In contrast to the siltstone samples, the 2 samples from the Johnnie oolite had NRM directions roughly aligned with that of the present axial field. Upon progressive demagnetization beginning at 200°C, the samples lost most of their intensity of magnetization by 300°C, and the directions of magnetization changed erratically at higher temperatures. Hence, not even the polarity of the Johnnie oolite could be ascertained. Examination of a thin section from the Johnnie oolite shows it to be an ooid grainstone, with voids commonly remaining between the grains. Rusty-colored stylolitic veins are common, and a rusty color pervades the entire rock. This rusty material, which is probably iron-bearing and of recent origin, in all likelihood dominates the NRM and the magnetic behavior at low-temperature demagnetization steps.

About half of the 74 samples from the carbonate unit of the Rainstorm Member were progressively demagnetized at 10 steps between 200 and 670°C; the rest were demagnetized at 5 steps between 550 and 670°C. The NRM directions of samples from this unit were scattered about the direction of the present axial field (Figure 3.2). At successively higher temperatures, directions from most samples migrated away from the

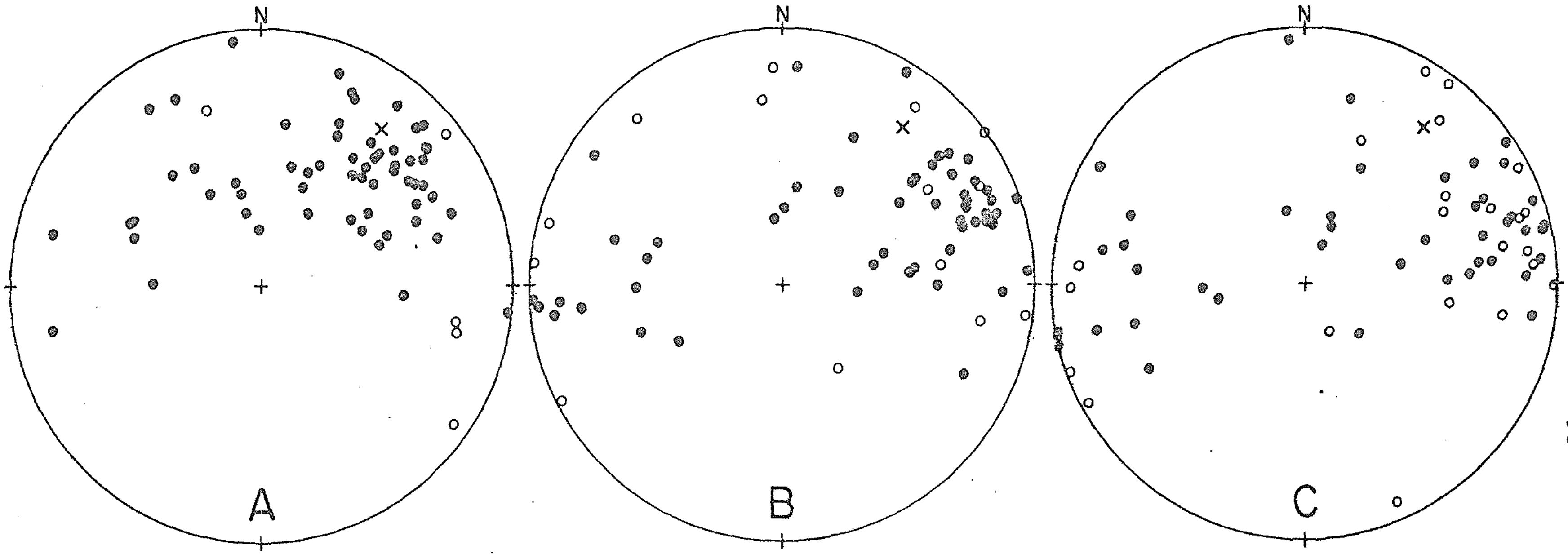

Figure 3.2 Directions of magnetization at selected steps from progressive thermal demagnetization of samples from the carbonate unit of the Rainstorm Member of the Johnnie Formation. Open (shaded) circles are on upper (lower) hemisphere, respectively. "X" marks the direction of the present axial dipole field at the sampling site after rotation for the mean attitude of beds. The directions from the stratigraphically highest 10 samples are not plotted, because they are from poorly exposed, slightly rotated blocks. (A) NRM; (B) 550°C; (C) 650°C.

present axial field direction to form bimodal distributions. The modes have low, positive inclinations and declinations in the northeast and southwest quadrants. By 650°C, the intensity of magnetization of most samples had decreased to 45-75% of their NRM values and the directions had become nearly antiparallel. At 670°C, the directions from most samples abruptly scattered.

Twelve duplicate specimens from the Rainstorm Member were subjected to progressive chemical demagnetization by soaking in concentrated HCl up to a maximum of 500 hr. The specimens were ring-shaped to increase the surface area for leaching; they were made by drilling holes about 1.5 cm across in wafers about 7 mm thick. Chemical demagnetization resulted in incomplete separation of magnetic components. A distribution somewhat skewed toward the positions revealed by thermal demagnetization could be discerned after prolonged immersion, and the intensity of magnetization decreased by about 50%. However, scatter of directions remained high, the directions of some specimens changed rather erratically, and many directions remained near that of the present axial field. Moreover, the abundance of carbonate in some samples caused them to disintegrate.

Six thin sections from the carbonate unit of the Rainstorm Member were examined in transmitted and reflected light. The rocks consist predominantly of silt-sized (20-30 μm) quartz grains, with subordinate feldspar and rare detrital white mica. Vague lamination defined by grain-size variations is present. The carbonate cement that occurs in these rocks is micritic and probably is calcite, as it reacts readily with cold, dilute HCl. The rocks are pervasively tinted a light pinkish-orange. Local reddish or red-orange hematitic patches and aggregates occur; these probably were formed by alteration of iron-bearing precursors. Minute (∿1 μm) flecks of specular hematite are occasionally associated with the aggregates; these flecks presumably also are authigenic, in contrast to the probably detrital grains described below.

Discrete grains of specular hematite, ∿10 μm across, constitute about 1% of these rocks. Many of them have squarish outlines, suggesting they may be martite. Most of these grains are probably detrital, as suggested by the following observations: (1) The grains are generally concentrated in lenses, stringers, and laminae that apparently are heavy mineral trains. (2) The grains commonly show rounding, suggestive of transport. (3) The grains have crisp contacts with the surrounding rock.

The 5 samples from the poorly exposed siltstone and quartzite unit at the top of the Rainstorm Member behaved unstably upon progressive thermal demagnetization. Not even the polarity of these samples could be ascertained.

Modes representing each of the two polarities of the characteristic magnetization of the Rainstorm Member were determined from the 650°C demagnetization step of the carbonate unit (Figure 3.2 and Table 3.1). These modes are not antiparallel and the distributions of directions are skewed. Evidently the present-axial-field component has not been completely removed, even by 650°C. An estimate of the direction representing one polarity of the characteristic magnetization can be made by rotating one of the modes through 180° and then averaging it with the

Table 3.1 LATE PRECAMBRIAN PALEOMAGNETIC POLE FROM THE DESERT RANGE, NEVADA

Formation	Section**	Mean Attitude (str.,dip)	$\overline{J}_{mode}$§ (Am^2/kg $\times 10^{-7}$)	$\frac{\overline{J}_{mode}}{\overline{J}_{NRM}}$	Geog.† Coords. (D, I)	Strat.†† Coords. (D, I)	β_{95}¶ $\left(\frac{N_{mode}}{N_{tot.}}\right)$	Apparent$^{\Psi}$ North Pole (Lat,Long)	Corrected$^{\delta}$ North Pole (Lat,Long)
Johnnie* (normal)	H,I,J	344, 51E	6.6	0.50	265,-41	261,+7	11°(10/20)		
Johnnie* (reversed)	H,I,J	344, 51E	5.1	0.31	73,+62	74,+8	7°(31/45)		
Johnnie* (average)	H,I,J	344, 51E				78,+1		10S, 162E	37S, 188E

*Directions from the carbonate unit of the Rainstorm Member of the (uppermost) Johnnie Formation. Normal=normal polarity; reversed=reversed polarity; average=average of the mode of normal-polarity directions with the mode of reversed-polarity directions (after first rotating the normal-polarity mode through 180°).

**Refers to paleomagnetic section locations of Figure 2.1

§ $\overline{J}_{mode}$ is the geometric mean of the specific intensities of magnetization at the demagnetization step upon which the mode is based (i.e., 650-660°C).

†Geographic coordinates are the mode of specimen directions uncorrected for the bedding attitude.

††Stratigraphic coordinates are the mode of specimen directions after correction for local bedding attitudes, not for the mean attitude of the entire formation.

¶ β_{95} is an estimate of the 95% confidence limits for the mode. The statistic β_{95} is derived from the largest subset (N_{mode}) of the total sample (N_{tot}) that has a mean identical with the mode of the total sample. This statistic is defined as the Fisherian half-angle of the cone of 95% confidence for the mean of this subset.

$^{\Psi}$Apparent north paleomagnetic pole calculated from the average stratigraphic coordinates before correction for the 36° clockwise rotation with respect to the Colorado Plateau (see text).

$^{\delta}$North paleomagnetic pole calculated from the average stratigraphic coordinates after correction for the 36° clockwise rotation with respect to the Colorado Plateau (see text).

mode of opposite polarity (cf. McElhinny, 1973, p. 85). This procedure yields a direction with declination(D) = 78° and inclination(I) = +1° in stratigraphic (i.e., dip-corrected) coordinates for the characteristic magnetization with easterly declination. This direction is assumed to represent reversed polarity, because of its proximity to the reversed-polarity direction from the middle member of the Lower Cambrian Wood Canyon Formation (Chapter 2).

An apparent polarity zonation is observed in the carbonate unit of the Rainstorm Member (Figure 3.3). The carbonate unit appears to contain at least 10 polarity zones, each of which is represented by at least two samples and is therefore locally reproducible.

The high blocking temperature (>650°C) of the characteristic magnetization of the carbonate unit of the Rainstorm Member demonstrates that this magnetization resides in hematite. Whether this magnetization is a detrital remanent magnetization (DRM) or a chemical remanent magnetization (CRM) is ambiguous. The thin-section observations, which show common, apparently detrital specular hematite in these rocks, suggest that the specular hematite might contain a significant fraction of the characteristic magnetization. In this case, the magnetization could be a DRM, acquired contemporaneously or penecontemporaneously with deposition. However, authigenic hematite also occurs in these rocks and could contribute a component of CRM. Thermal demagnetization experiments cannot distinguish conclusively between remanences residing in authigenic versus detrital hematite. Chemical demagnetization (which, in favorable cases, can distinguish CRM from DRM by preferentially leaching away

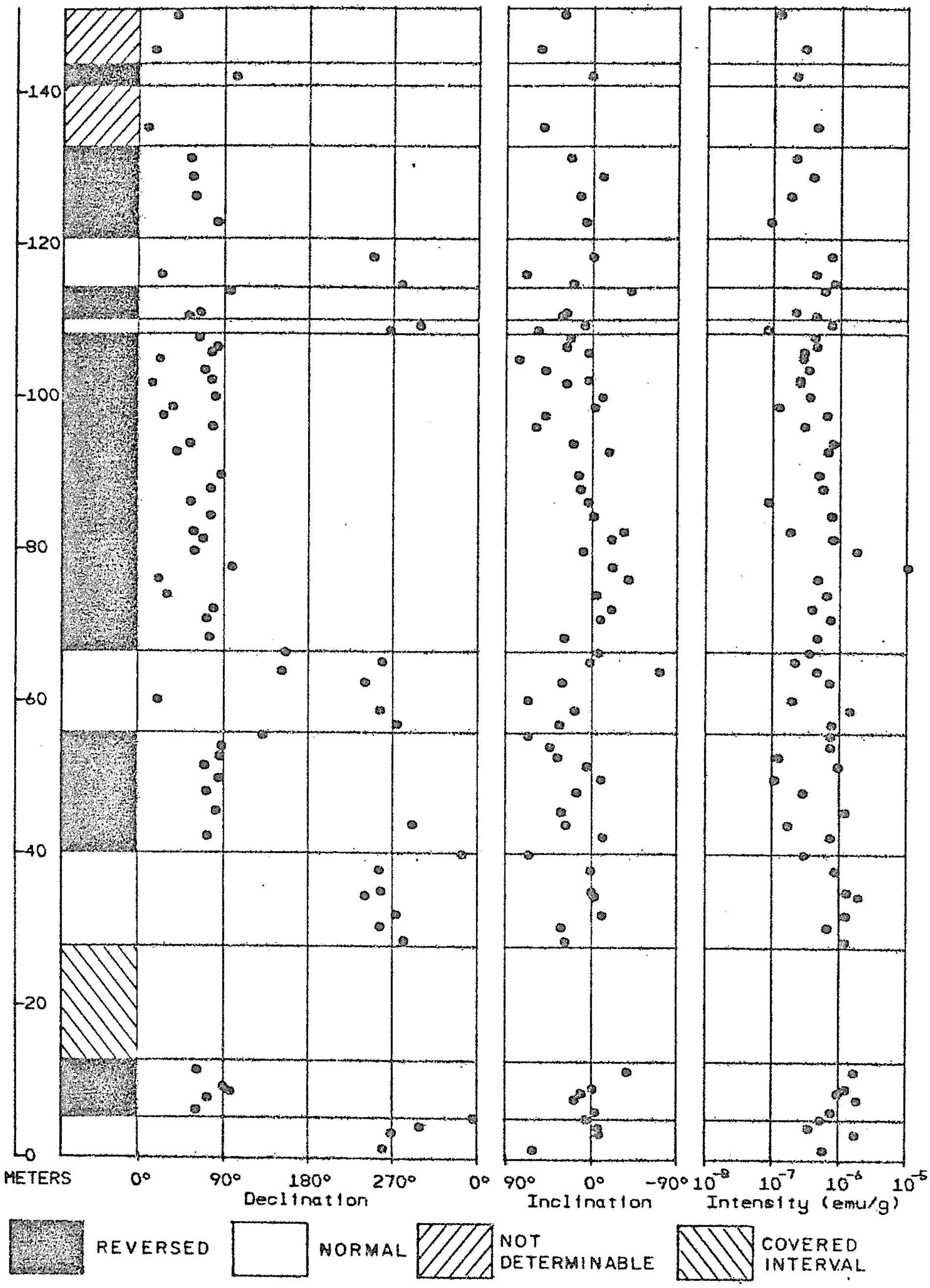

Figure 3.3 Apparent polarity zonation of the carbonate unit of the Rainstorm Member of the Johnnie Formation in the Desert Range, Nevada. Polarity assignments have been made by inspection of vector demagnetization diagrams of each sample. The base of the section is the top of the "Johnnie oolite"; the top of the section is the top of the carbonate unit.

fine-grained authigenic hematite) gave equivocal results for rocks from the Rainstorm Member.

Even if the characteristic magnetization were largely a CRM, stratigraphic evidence suggests that it was acquired in the late Precambrian. The Rainstorm Member, unlike the underlying strata of the Johnnie Formation, contains distinctive units that can be correlated in detail over a wide area. In particular, the Johnnie oolite and the reddish middle carbonate unit are present throughout the southern Great Basin (Stewart, 1970). Moreover, hematite-bearing beds that can be confidently correlated on lithologic grounds with the Rainstorm Member occur in late Precambrian strata exposed several hundred kilometers to the north, in eastern Nevada and western Utah (Stewart, 1974). If the hematite of the carbonate unit of the Rainstorm had been produced chiefly by late alteration in the Phanerozoic, its distribution would probably be more spotty than is suggested by the regional stratigraphy.

As discussed in Chapter 2, deposition in this part of the miogeocline was essentially continuous throughout the late Precambrian and Paleozoic. By the late Paleozoic, the Rainstorm Member was buried to depths of about 7 km, far below surface weathering and oxidation. Furthermore, the Rainstorm is sandwiched between hundreds of meters of quartzite, siltstone, and shale. Because these rocks are not very permeable (Winograd and Thordarson, 1968), it seems implausible that oxidizing groundwaters could penetrate them to oxidize the Rainstorm Member on a regional scale in the Phanerozoic.

Another indication of a late Precambrian time of acquisition of the characteristic magnetization of the Rainstorm Member comes from consideration of its corresponding paleomagnetic pole position. A paleomagnetic pole for the Rainstorm computed directly from its characteristic magnetization direction (in stratigraphic coordinates) lies more than 60° from the post-Early Ordovician APW path for North America (Figure 3.4). This suggests that the characteristic magnetization of the Rainstorm predates the Ordovician. As discussed in Chapter 2, the pole during the earliest Paleozoic seems to have been relatively stationary with respect to North America, near the position of the paleomagnetic pole from the Lower and Middle Cambrian Tapeats Sandstone of the Colorado Plateau (Elston and Bressler, 1977). Even the pole from the Tapeats, however, is about 15° from that determined directly from the characteristic magnetization direction of the Rainstorm Member (Figure 3.4). This suggests that the magnetization of the Rainstorm was acquired prior to the Early Cambrian.

In Chapter 2, it was shown that characteristic magnetization directions derived from Lower and Middle Cambrian strata of the Desert Range were consistent with a net 36° clockwise rotation of this part of the range. This vertical-axis rotation was inferred from the mismatch in paleomagnetic declination between the Cambrian strata from the Desert Range and the partly coeval Tapeats Sandstone. The net rotation was attributed chiefly to oroflexural bending in the late Cenozoic, but may in part have been caused by thrusting in the Mesozoic. Because the Rainstorm Member was sampled in the same homocline as the Cambrian strata, a 36° declination correction must also be applied to the characteristic magnetization of the Rainstorm. The corrected north paleomagnetic pole from the Rainstorm (37°S, 188°E; Table 3.1) falls even farther from both

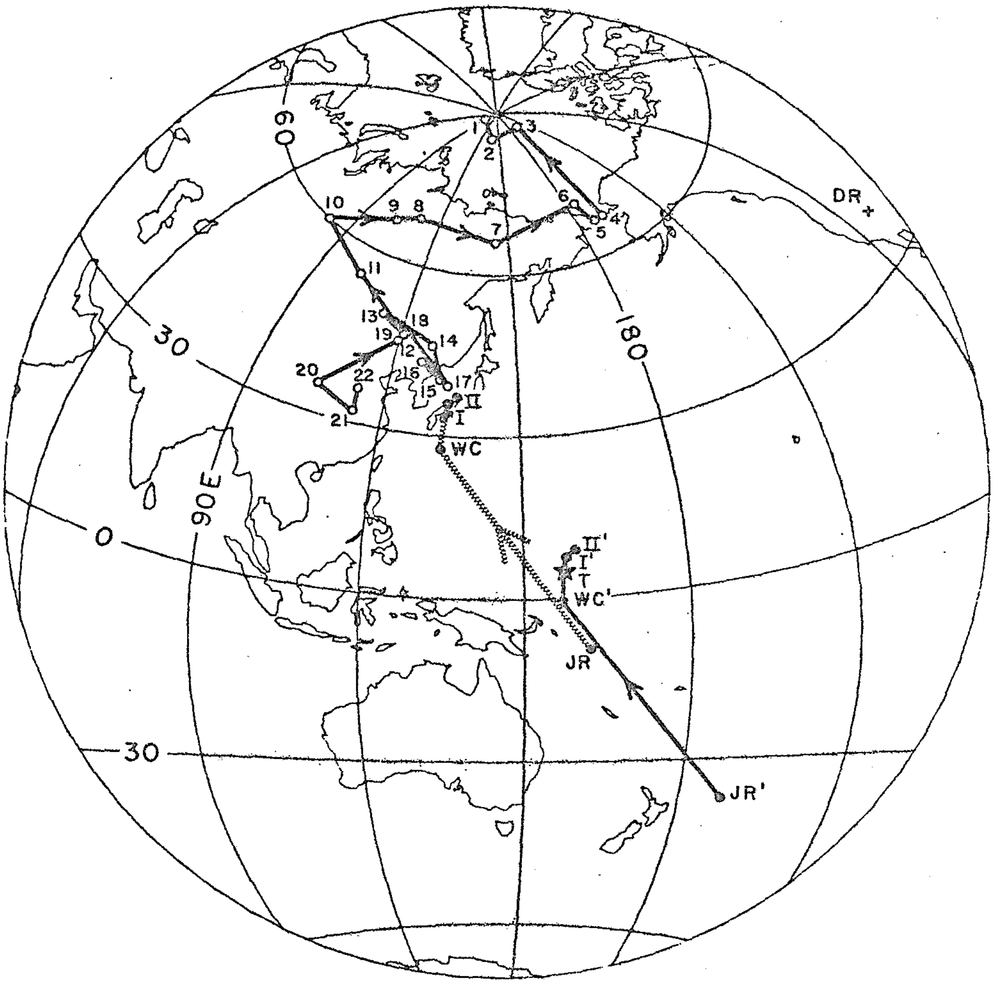

Figure 3.4 Late Precambrian and Cambrian paleomagnetic poles from the Desert Range before (unprimed symbols) and after (primed symbols) correction for the proposed 36° clockwise vertical-axis rotation of the Desert Range section with respect to the Colorado Plateau (Chapter 2). II = uppermost Carrara and lowermost Bonanza King Formations (Middle Cambrian), I = remainder of the Carrara Formation (Lower and Middle Cambrian), and WC = Wood Canyon Formation (Lower Cambrian), all from Chapter 2; JR = Rainstorm Member of the Johnnie Formation (upper Precambrian), from this study. The star labeled "T" is the pole from the Tapeats Sandstone (Lower and Middle Cambrian) of the Colorado Plateau (Elston and Bressler, 1977). The cross labeled "DR" marks the sampling site. Also shown is the post-Early Ordovician part of the apparent polar wander path for North America determined in Chapter 4. Numbers on the path refer to nearly uniform, 22-m.y. time intervals: 1-3, Cenozoic; 4-7, Cretaceous; 7-9, Jurassic; 10-11, Triassic; 12-13, Permian; 14-15, Pennsylvanian; 16-17, Mississippian; 18-19, Devonian; 20-21, Silurian; 21-22, Ordovician.

the post-Early Ordovician APW path for North America and from the pole for the Tapeats (Figure 3.4). This lack of correspondence between the paleomagnetic pole for the Rainstorm and Phanerozoic poles suggests that the characteristic magnetization of the Rainstorm was indeed acquired in the late Precambrian.

The regional extent of the distinctive units in the Rainstorm Member, together with the polarity reversals found in the Desert Range, suggest yet another test for the time of acquisition of its characteristic magnetization. If the magnetization were acquired soon after deposition, and if deposition rates were relatively uniform, the distinctive polarity zonation of the Rainstorm in the Desert Range would be reproducible at other localities. As an initial step in such a test, the authors and colleagues have sampled the Rainstorm in the southern Nopah Range, about 120 km southwest of the Desert Range section. The data obtained to date suggest that the polarity zonation of the Rainstorm in the Nopah Range is indeed similar to that in the Desert Range (Figure 3.5).

Stirling Quartzite

The Stirling Quartzite, which conformably overlies the Rainstorm Member of the Johnnie Formation, consists of light-colored quartzite with minor siltstone and dolomite (Figure 3.1). Because it lies below the earliest occurrence of an olenellid trilobite/archeocyathid fauna, the Stirling has been assigned a late Precambrian age (Stewart, 1970). However, Nelson (1978) has suggested that the Precambrian-Cambrian boundary may lie in the upper Stirling, on the basis of proposed litho-

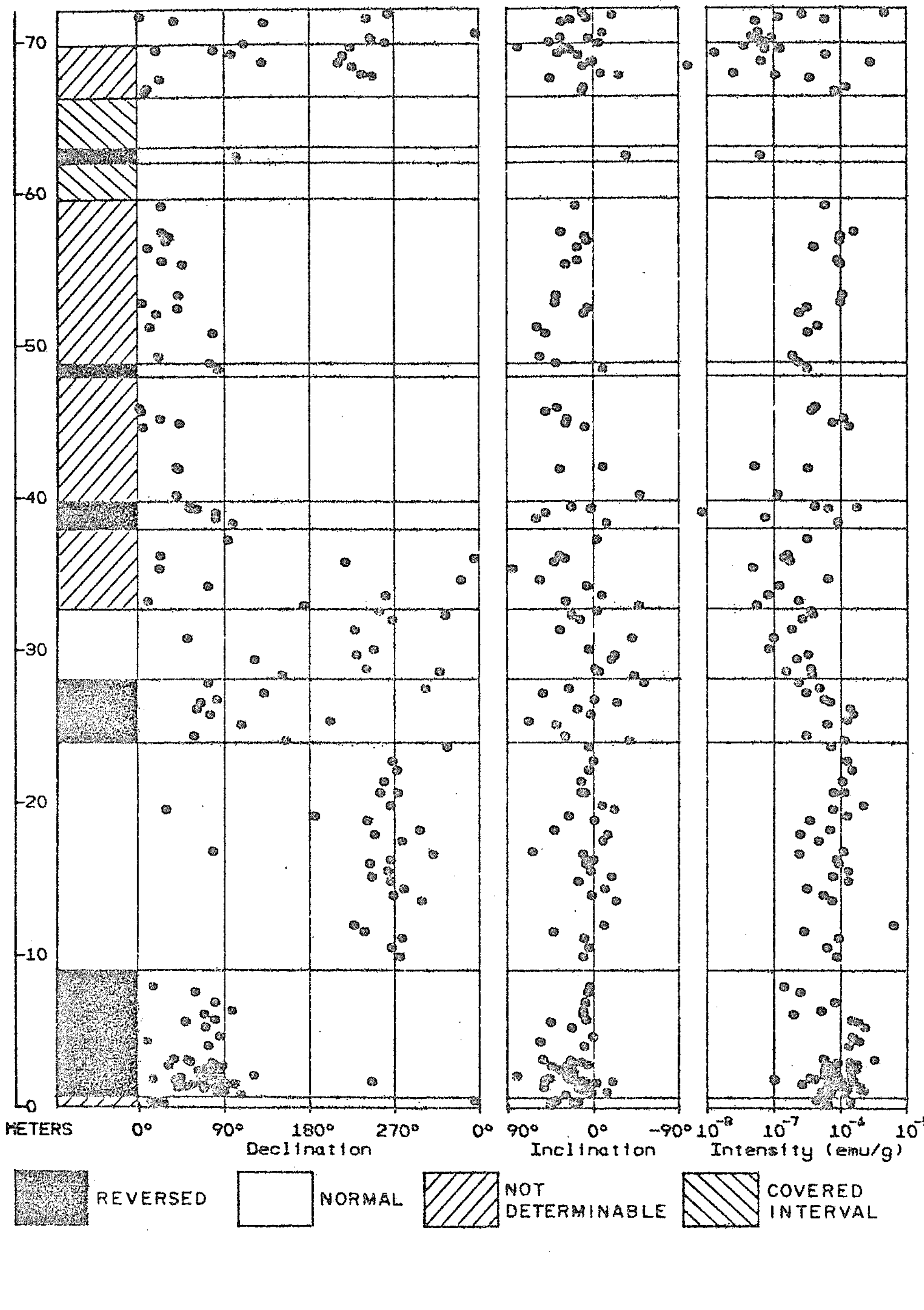

Figure 3.5 Apparent polarity zonation of the carbonate unit of the Rainstorm Member of the Johnnie Formation in the Nopah Range, California. Polarity assignments have been made by inspection of vector demagnetization diagrams of each sample. The base of the section is the top of the "Johnnie oolite"; the top of the section is the top of the carbonate unit.

logic correlations with more fossiliferous strata in the White-Inyo Mountains of eastern California. The Stirling has been divided into 5 informal members of wide extent, designated A, B, C, D, and E by Stewart (1970). The dolomitic D member is absent in the Desert Range, as it is in most of the more eastern sections of the miogeoclinal sequence (Stewart, 1970). At a locality 6.5 km south of the Desert Range section sampled in this study, Stewart and Barnes (1966) and Stewart (1970) have measured and described a total of 950 m of beds belonging to the A, B, C, and E members.

Of a total of 180 samples collected from the Stirling, 142 are from light-colored quartzite and 6 are from interbeds of dolomitic sandstone and purple siltstone from the A, C, and E members. The other 32 samples are from purple to red-purple siltstones and claystones from near the base of the C member. The siltstones and claystones from the C member are the only rocks from the Stirling in which a characteristic magnetization could be discerned.

All samples of the siltstone and claystone were subjected to progressive thermal demagnetization at 6 steps between 400 and 640°C. The NRM directions of most samples clustered near the direction of the present axial field (Figure 3.6). By 500°C, directions from about half of the samples had remained essentially stationary, whereas the other half had moved slightly away from the present axial field toward a direction with more easterly declination and shallower inclination. Bimodal distributions of directions were evident at the 610 and 625° demagnetization steps. At 640°C, however, samples representing each of the two modes displayed a viscous magnetization; hence, demagnetization

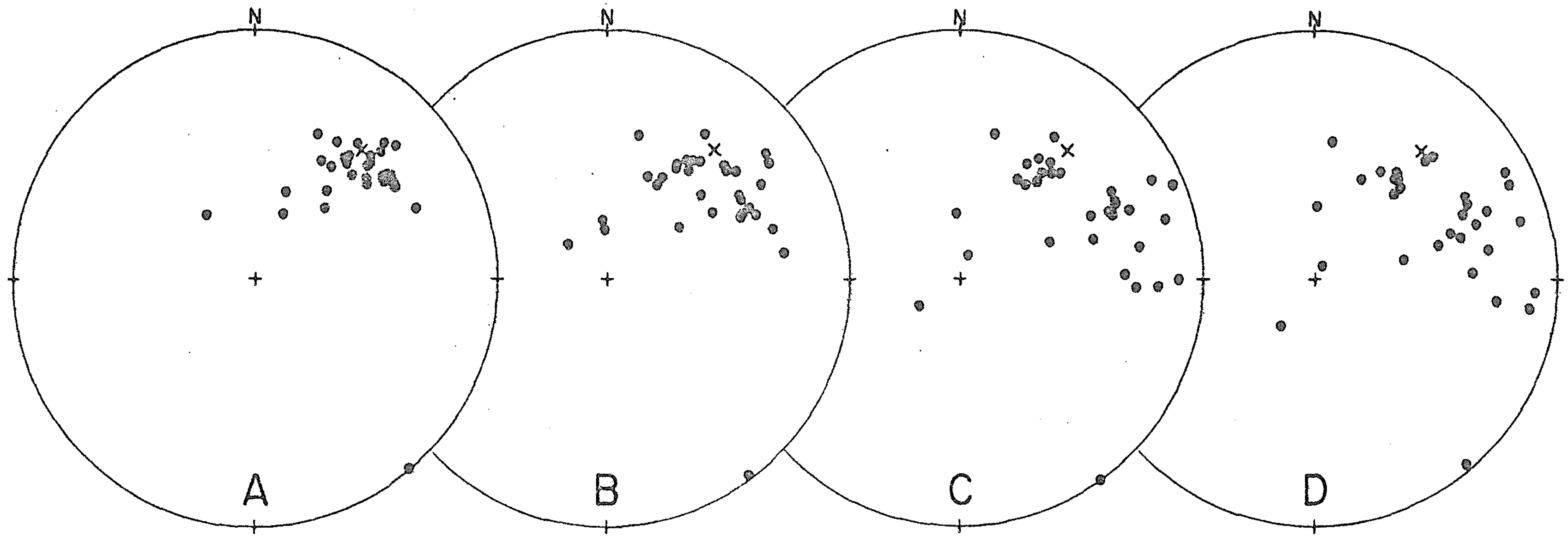

Figure 3.6 Directions of magnetization at selected steps from progressive thermal demagnetization of red-purple siltstones and claystones from the lower C member of the Stirling Quartzite. Symbols are as in Figure 3.2. (A) NRM; (B) 500°C; (C) 610°C; (D) 640°C.

of these samples at higher temperatures was not attempted.

These progressive thermal demagnetization experiments reveal the existence of two components of magnetization in about half these rocks: (1) a component of recent origin aligned roughly with the direction of the present axial field; and (2) a characteristic component, with a more shallow inclination and an easterly declination. The characteristic magnetization has only one polarity. Because the stability spectra of the two components overlap, the directions from samples in which both components are present never reach a stable endpoint. This, together with the single-polarity nature of the characteristic magnetization, precludes a precise determination of its direction. However, this magnetization appears to have a direction similar to the reversed-polarity directions from the Rainstorm Member of the upper Precambrian Johnnie Formation (D=78°, I=+1°) or the middle member of the Lower Cambrian Wood Canyon Formation (D=124°, I=-3°) (Chapter 2). This suggests that the characteristic magnetization in the C member of the Stirling was acquired in the Precambrian and is of reversed polarity.

All samples of red-purple siltstone and claystone were examined under a low-power, binocular microscope; in addition, two thin sections of red-purple claystone were examined in transmitted and reflected light. The two rock types are intergradational in lithology, and the variation in grain size is not correlated with the thermal demagnetization behavior. An opaque, red, hematitic cement is abundant in the fine-grained parts of these rocks, accounting for their color in hand specimen (5 P 4/2, purple, to 5 RP 4/2, red-purple). This cement is the dominant form of hematite in the rocks. Specular hematite was found only in one of the two thin sections, and some of this is probably authigenic, as suggested by its association with hematitic aggregates.

The hematitic cement probably carries the major part of the characteristic magnetization in the siltstones and claystones from the Stirling Quartzite. Although this cement must postdate deposition to some degree, textural evidence suggests that it was formed quickly, implying that any CRM associated with the cement was probably acquired penecontemporaneously with deposition. These textural relations are similar to those in red-purple mudstones in the middle member of the overlying Wood Canyon Formation (Chapter 2). The siltstones and claystones contain laminae; lenses; and irregular, crosscutting, infilled cracks or burrows, all of which are composed of coarser, quartzo-feldspathic material. The cement tends to be confined to the fine-grained areas, so that the coarser areas remain light-colored, and the contact of the two is commonly sharp. This suggests that at least the precursor to the cement was present shortly after deposition. One would expect hematite from late alteration (a) to show more diffuse contacts; and (b) to follow preferentially the coarse-grained areas because of their initially higher permeability.

Progressive thermal demagnetization did not reveal a direction of characteristic magnetization from other units of the Stirling Quartzite. The quartzitic and dolomitic samples from the Stirling were progressively thermally demagnetized at 5 steps between 400 and 640°C; about half were initially demagnetized at 200 and 300°C as well. Directions from the quartzites and from 4 brown dolomite interbeds were poorly grouped before demagnetization, and they remained dispersed at every demagnetization step (Figure 3.7). Directions from some of these samples, especially the dolomitic interbeds, changed through large angles in an erratic manner. However, directions from most samples, which were predominantly quartzite, were extremely stable to thermal demagnetization even by 640°C. This suggests that the Stirling quartzites contain multiple components of magnetization that reside in hematite and that cannot be separated by thermal demagnetization. These negative results discouraged us from sampling the quartzitic B member.

A thin section of pale reddish quartzite from the upper C member of the Stirling, representative of the quartzites with scattered but stable directions, was examined in transmitted and reflected light. The rock is composed predominantly of sand-sized ($\sim$150 μm) quartz grains, which are rounded to subangular. Occasional detrital white mica and feldspar are present. Discrete grains of specular hematite comprise less than 0.5% of the rock. They are commonly squarish, which suggests they may be martite; in addition, they show rounding and crisp contacts, which suggest they are detrital. Aggregates containing earthy and/or fine-grained specular hematite are also common; these aggregates apparently represent alteration "ghosts" of iron-bearing minerals. Veins containing earthy and specular hematite also occur. An orange cast pervades the rock.

The thin section observations may partly explain the magnetic behavior of the Stirling quartzites. Several distinct forms of hematite

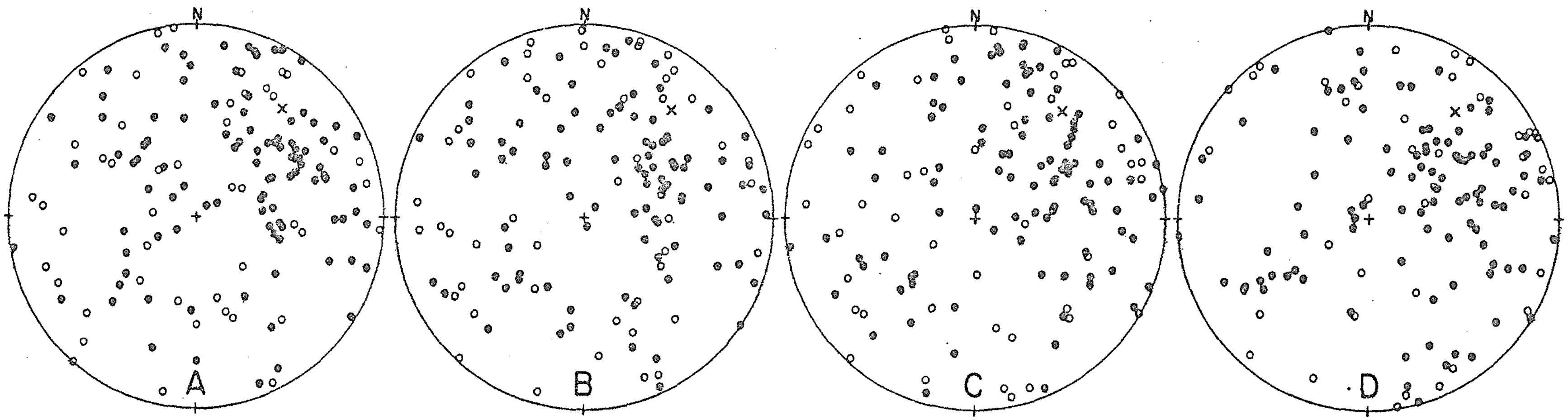

Figure 3.7 Directions of magnetization at selected steps from progressive thermal demagnetization of quartzite samples from the A, C, and E members of the Stirling Quartzite. Symbols are as in Figure 3.2. (A) NRM; (B) $400^{\circ}C$; (C) $550^{\circ}C$; (D) $640^{\circ}C$.

are present, and all probably contribute in varying degrees to the observed remanence. Roy and Robertson (1978) suggested that scattered, stable directions can arise from the presence of multiple components of magnetization, perhaps acquired during a time of frequent polarity reversals. Several components of magnetization could be present in the various forms of hematite observed in the Stirling, and this may account for the scattered directions of magnetization.

Discussion

The paleomagnetic data presented above suggest that both the Rainstorm Member of the Johnnie Formation and the C member of the Stirling Quartzite contain characteristic magnetizations acquired penecontemporaneously with deposition in the latest Precambrian. The precise direction of this characteristic magnetization could be determined only in the Rainstorm Member, in which two polarities of the magnetization are present. The corresponding north paleomagnetic pole from the Rainstorm is displaced about 47° from the north paleomagnetic pole from the stratigraphically higher Wood Canyon Formation of Early Cambrian age.

For purposes of paleogeographic reconstruction and stratigraphic correlation, it is important to place constraints on the timing of this 47° apparent polar shift. The absolute age of the Rainstorm Member is unknown, because of the absence of any direct isotopic age control in the late Precambrian miogeoclinal sequence in the southern Great Basin. Estimates of the absolute age of the Rainstorm can be made from sedimentation rates derived for the strata of the Desert Range. Early through Late Cambrian strata comprising the uppermost Stirling Quartzite and overlying Wood Canyon, Carrara, Bonanza King, and Nopah Formations total about 3000 m. The duration of the Cambrian has been estimated to be about 75 m.y. (Lambert, 1971). On this basis, the average deposition rate of Cambrian strata in the Desert Range is roughly 40 m/m.y. Essentially the same rate is obtained by arbitrarily dividing the Cambrian into thirds and computing the deposition rate from the Early Cambrian strata alone. This suggests that the change from dominantly terrigenous sedimentation in the late Precambrian and Early Cambrian was not accompanied by a gross change in deposition rate. From the faunal evidence summarized earlier, the Precambrian-Cambrian boundary probably lies in the upper Stirling Quartzite (Nelson, 1978). The pole from the Rainstorm Member of the Johnnie comes from beds approximately 800 m below the suggested Precambrian-Cambrian boundary. If the late Precambrian strata in the Desert Range were deposited at the rate of 40 m/m.y., estimated for the Cambrian, then the Rainstorm would be approximately 20 m.y. older than the Precambrian-Cambrian boundary. Assuming this boundary to have an age of about 590 m.y.B.P. (Lambert, 1971), the age of the Rainstorm would be about 610 m.y.B.P.

A second very rough estimate of the absolute age of the Rainstorm can be made from consideration of sedimentation rates off the Gulf Coast of the United States. In this region, thick near-shore accumulations of sediments are thought to represent a modern analogue of miogeoclinal sedimentation (Matthews, 1974, p. 82-88). During the Cenozoic, a time comparable in length to the Cambrian, miogeoclinal sediments were deposited at rates ranging between 20 m/m.y. off Florida to 200 m/m.y.

near the Mississippi delta (Matthews, 1974, p. 88). If these limits bracket the deposition rates of the late Precambrian strata of the Desert Range, then the Rainstorm would lie between 4 and 40 m.y. below the Precambrian-Cambrian boundary. Because the Precambrian-Cambrian boundary probably lies in the range 610 to 570 m.y.B.P. (Lambert, 1971), reasonable limits to the absolute age of the Rainstorm would be about 650 to 575 m.y.B.P.

A third estimate of the age of the Rainstorm can be made on the basis of lithologic correlations and isotopic age control from other parts of the Cordilleran Geosyncline. Stewart (1972) has emphasized the stratigraphic similarities of the belt of upper Precambrian to Lower Cambrian strata extending from Alaska to northern Mexico. The basal units of these successions characteristically contain diamictite (possibly of glacial origin) and tholeiitic basalt, overlain by thick sections of siltstone, shale, argillite, quartzite, and conglomerate. In the southern Great Basin, the diamictite and associated volcanics occur in the Kingston Peak Formation, which unconformably underlies the Noonday Dolomite; the Noonday is conformable with the overlying Johnnie Formation. Stewart proposed that the diamictite and associated volcanics mark the initiation of a late Precambrian continental separation, followed by deposition of a thick wedge of miogeoclinal sediments on a trailing-edge continental margin. This separation must be younger than the youngest rocks of the Belt Supergroup (about 850 m.y.B.P.; cf. Stewart, 1972). Assuming: (1) an 850 m.y.B.P. age for the base of the Noonday Dolomite; (2) a constant sedimentation rate for the overlying late Precambrian strata; and (3) a 590 m.y.B.P. age for the Precambrian-Cambrian boundary, the Rainstorm would have an age of about 675 m.y.B.P. This estimate is based on the thickness of the late Precambrian strata in the Nopah Range, where a complete sequence is exposed from the basal Noonday Dolomite to the Middle Cambrian Bonanza King Formation. The three assumptions made in this calculation, however, would indicate an average sedimentation rate of 7 m/m.y. in the Nopah Range, which seems rather slow for miogeoclinal sedimentation. Hence, the 675 m.y.B.P. estimate probably represents a maximum age for the Rainstorm.

The maximum and minimum age estimates set forth above are 675 and 575 m.y.B.P. We suggest that the age of the Rainstorm Member of the Johnnie Formation is about 625 ± 50 m.y.B.P.

Few paleomagnetic poles have been reported from rocks of Laurentia (i.e., North America, Greenland, and the Lewisian Platform of northwest Scotland) that might date from 675 to 575 m.y.B.P. (Table 3.2, Figure 3.8). Colorado intrusives studied by French et al. (1977) and by Larson and Mutschler (1971) are dated as 704 to 485 m.y.B.P. and give a multi-modal grouping of directions (represented by poles 1 through 5). Pole 6, from 3 flows of the late Precambrian to Early Cambrian (605(?) m.y.B.P.) Cloud Mountain Basalts, western Newfoundland, may have incompletely averaged secular variation; in addition, this pole may reflect a vertical-axis rotation of that island (Deutsch and Rao, 1977).

Perhaps the most reliable late Precambrian poles for Laurentia come from flows, dikes, and sills of the Canadian Arctic (Fahrig et al., 1971; Robertson and Baragar, 1972; Fahrig and Schwarz, 1973; and Palmer and

Table 3.2

PALEOMAGNETIC POLES FROM LAURENTIA THAT MAY DATE FROM 675-575 M.Y.B.P.

Pole No.	Pole Position	Rock Unit, Location	Age	Reference
1	37N, 122E	Intrusives, Colorado	Late Precambrian-Early Ordovician (704-485 m.y.B.P.)	Larson and Mutschler, 1971
2	44N, 100E			French *et al.*, 1977
3	15N, 142E			
4	5N, 174E			
5	48N, 107E			
6	5N, 172E	Cloud Mtn. Basalt, W. Newfoundland	Late Precambrian-Early Cambrian (605±10 m.y.B.P.)	Deutsch and Rao, 1977
7	4S, 161E	Franklin rocks, Can. Arctic (Group A)	Late Precambrian (675-625 m.y.B.P.)	Palmer and Hayatsu, 1975
8	8N, 166E	(Group B)		
9	4N, 199E*	Tillite Fm., East Greenland	∿600	Bidgood and Harland, 1961
10	13S, 219E*	Multicoloured Series, East Greenland	∿600-800	
11	9S, 159E	Mafic rock, Michigan	Late Precambrian-Cambrian(?)	Van der Voo and Watts, 1978
12	4N, 169E	Upper Keween. overprint, Michigan	Late Precambrian-Cambrian(?)	Henry *et al.*, 1977
13	17N, 137E	Tudor Gabbro, Ontario	(1200-670 m.y.B.P.)	Palmer and Carmichael, 1973 Hayatsu and Palmer, 1975

*Pole uncorrected for rotation of Greenland with respect to North America.

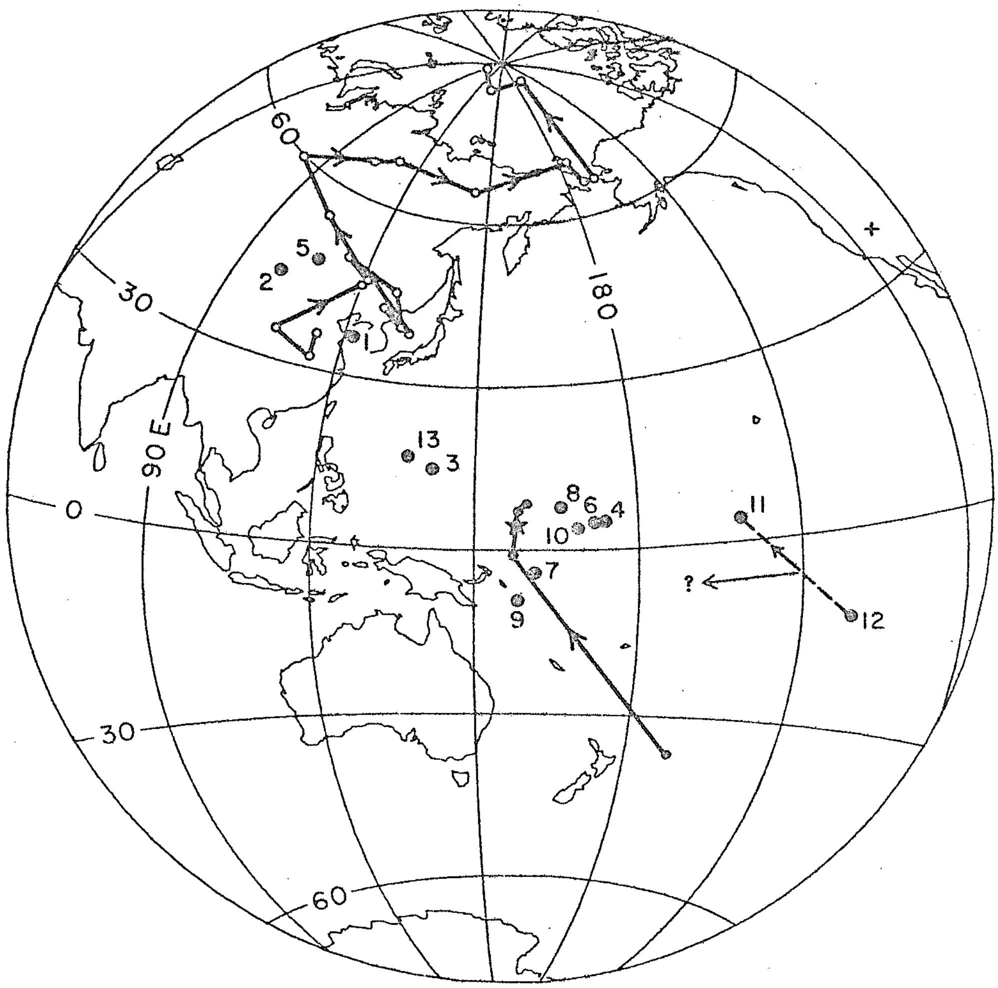

Figure 3.8 Paleomagnetic poles from North America and Greenland that may represent the time 675-575 m.y.B.P. Pole numbers are as in Table 3.2. Also shown are the late Precambrian and Cambrian apparent polar wander path derived from the Desert Range (this study and Chapter 2) and the post-Early Ordovician part of the apparent polar wander path for North America (Chapter 4).

Hayatsu, 1975). Palmer and Hayatsu (1975) have shown that lavas of the Natkusiak Formation and apparently related diabase sills from Victoria Island have a single-polarity (reversed?) magnetization with a pole (pole 7) similar to that from the Coronation sills of the mainland. The Natkusiak Formation disconformably overlies the upper Precambrian Shaler Group and is disconformably overlain by Cambrian strata (Fahrig et al., 1971). K-Ar age determinations on two sites in the least altered lavas gave a whole-rock isochron at 625 m.y.B.P. In contrast to these single-polarity results, possibly coeval intrusives of the central and eastern Arctic yield two-polarity magnetizations with a slightly more northerly pole position (pole 8). The mean of post-1963 K-Ar age determinations on these intrusives is in agreement with the isochron from the Natkusiak (Palmer and Hayatsu, 1975; Fahrig et al., 1971). However, Fahrig et al. preferred an age of 675 m.y.B.P. for the intrusives from the central and eastern Arctic.

Other latest Precambrian to earliest Cambrian poles come from reconnaissance paleomagnetic sampling by Bidgood and Harland (1961) of the thick, generally conformable succession of East Greenland. Pole 9, from the Tillite Group of latest Precambrian or earliest Cambrian age, is derived from the 16 most stable samples. Although no demagnetization studies were performed on these rocks, stability of the magnetization is suggested by the reversal and fold tests. In addition, 10 samples from the Multicoloured Series of the upper Precambrian Upper Eleonore Bay Group gave a single-polarity (reversed?) direction with a pole (pole 10) similar to that from the Tillite Group; however, no field or laboratory tests for stability were reported. The Multicoloured Series underlies the Limestone-Dolomite Series, which contains stromatolites comparable to those from Spitsbergen assigned a Vendian age (700-600 m.y.B.P.) (Henriksen and Higgins, 1976). If the poles from East Greenland were corrected for rotation of Greenland with respect to North America, they would become more consistent with the late Precambrian and Early Cambrian results from elsewhere in Laurentia.

Several paleomagnetic poles have recently been reported from upper Precambrian or Lower Cambrian rocks of the Lake Superior region. Most of these poles are from the poorly dated Jacobsville Formation (French, 1976; Roy and Robertson, 1978), for which estimated ages range from late Precambrian to Triassic (Hamblin, 1958). The most thorough paleomagnetic study of the Jacobsville (Roy and Robertson, 1978) supports an age greater than 1000 m.y.B.P., consistent with late Precambrian ages previously suggested for the Jacobsville by Dubois (1962) and Babcock (1975). Van der Voo and Watts (1978) reported a paleomagnetic pole (pole 11) similar to that of the Franklin Group A pole from a deep (5.32 km), fully-oriented core in a mafic unit. The regional stratigraphy suggests that this unit is an altered Keweenawan lava, but on the basis of its pole position, these authors suggest that it was completely remagnetized in the late Precambrian-Cambrian. Finally, Henry _et al._ (1977) have reported a secondary magnetization in the Nonesuch Shale (~1060 m.y.B.P.) and Freda Sandstone (~1050 m.y.B.P.); they suggested this magnetization might have been acquired during an episode of late Precambrian copper mineralization. However, this secondary magnetization, represented by pole 12, also may well be older than 675 m.y.B.P.

Pole 13, from the Tudor Gabbro in the Grenville Province (Palmer and Carmichael, 1973), may date from about 670 m.y.B.P. Although Palmer and Carmichael thought this gabbro to have been emplaced at about 1200 m.y.B.P. (on the basis of isotopic ages on possibly coeval intrusives), no isotopic ages had been obtained from the Tudor itself. Hayatsu and Palmer (1975) have since reported a 670 m.y. K-Ar isochron for the Tudor Gabbro. However, the relation of this K-Ar age to the age of the magnetization is unclear. In addition, a controversy exists about whether the Grenville Province has undergone significant rotation with respect to the rest of North America (e.g., Irving and McGlynn, 1976).

None of the late Precambrian poles that might be younger than 675 m.y.B.P. is close to the pole from the Rainstorm Member of the Johnnie Formation (Figure 3.8). This suggests that the Rainstorm pole is not contemporaneous with any of them, a hypothesis permitted by the imprecise absolute ages for all these poles.

Morris and Roy (1977) and Roy and Robertson (1978) presented evidence for a nearly 180° apparent polar shift with respect to North America between about 1000 and 675 m.y.B.P. The post-675 m.y.B.P. shift inferred from paleomagnetic investigations in the Desert Range may represent the later part of a long history of large-amplitude apparent polar wandering in the late Precambrian.

Conclusion

The 47° difference between the paleomagnetic pole from the uppermost Johnnie Formation (late Precambrian; 625 ± 50 m.y.B.P.) and the pole from the middle Wood Canyon Formation (Early Cambrian) suggests that at least 45° of apparent polar wandering occurred with respect to North America in latest Precambrian time.

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Chapter 4

A REVISED PHANEROZOIC APPARENT POLAR WANDER PATH FOR NORTH AMERICA

Apparent polar wander (APW) paths determined from paleomagnetic data are not only useful in making paleogeographic reconstructions, but also promise to be of value in stratigraphic correlation. The interest in interpretation of these paths, however, has not been matched with sufficient criticism of the methods by which they are constructed.

The validity of constructing APW paths is dependent on several assumptions: (1) that the configuration of the time-averaged geomagnetic field has been that of a geocentric axial dipole; (2) that the primary or characteristic component of remanent magnetization of a variety of rock types accurately records a direction of the ancient geomagnetic field at the sampling site; (3) that the time at which the characteristic component was acquired can be determined; and (4) that the post-magnetization tectonic history of the rocks can be deciphered. Failure of one or more of these assumptions can lead to spurious paleomagnetic poles and inaccurate APW paths.

Despite the many possible sources of error, even the earliest APW paths were accurate enough to show progressive motion of the paleomagnetic pole over a distance of 90° since the Cambrian. The causes of this relatively slow (0.2-2.0°/m.y.) apparent polar wander are the subject of current debate. Plate tectonics (differential motion of segments of the lithosphere), true polar wandering (net motion of the entire mantle with respect to the spin axis), and, perhaps, independent motion of the magnetic dipole with respect to the spin axis have all been considered as contributors to apparent polar wandering.

In this chapter, I will review North American paleomagnetic data in order to track the magnetic pole's long-term apparent motion with respect to North America during the Phanerozoic. Consideration of the causes of apparent polar wandering are deferred to Chapter 5. After a brief comment on the accuracy of previous APW paths, the techniques used to generate a revised North American Phanerozoic path will be described. This revised APW path is documented with a table and bibliography, supplemented with a period-by-period commentary.

Previous Work

The first paleomagnetically determined APW paths were published by Runcorn (1956), Creer, Irving, and Runcorn (1957), and Dubois (1958). The accuracy of these early paths was limited, because most of the data were obtained prior to the development of alternating-field, thermal, or chemical demagnetization techniques. The early paths were refined by Irving (1964), Creer (1967), Hospers (1967), and Strangway (1970), utilizing both the older data and incorporating a growing body of second-generation paleomagnetic data based on demagnetization procedures. Phillips and Forsyth (1972) computed a post-Devonian world mean APW path, and McElhinny (1973) compiled global paleomagnetic data to up-date the APW paths of all the major lithospheric plates. Recently, Van der Voo and French (1974) reviewed the APW for the Atlantic-bordering continents from the Late Carboniferous to Eocene; Irving (1977) employed a running-mean technique to revise the APW paths of the major continental blocks since the Devonian; and Van Alstine and de Boer (1978) presented a synopsis of the present study.

The common procedure for constructing APW paths for the Phanerozoic has been to connect mean poles calculated for entire geologic periods, giving unit weight to results listed in published compilations of poles. The disadvantages of this method are threefold. First, employing the geologic period as the standard division of time limits the time resolution of the pole-paths. Because Phanerozoic periods span, on the average, about 50 m.y., while rates of apparent polar wander are typically between 0.2-2.0°/m.y., significant details of apparent polar motion are averaged out. Moreover, because geologic periods are not of uniform length but vary between about 30 m.y. (Silurian) and about 75 m.y. (Cambrian), the time resolution is variable.

Second, if unit weight is given indiscriminately to entries in compiled tables, the mean pole position that is derived for a given period may not be representative of the time interval. Most compilations of poles include both true paleomagnetic poles, thought to represent sufficient intervals of time that geomagnetic secular variation has been satisfactorily averaged out, and virtual geomagnetic poles (VGP's), which represent instantaneous readings of the paleomagnetic field (McElhinny, 1973, p. 24-26). Averaging VGP's with paleomagnetic poles can generate inaccurate poles for geologic periods, especially in cases where a VGP represents an anomalous configuration of the geomagnetic field (e.g., a polarity transition or a short-period excursion of the field direction).

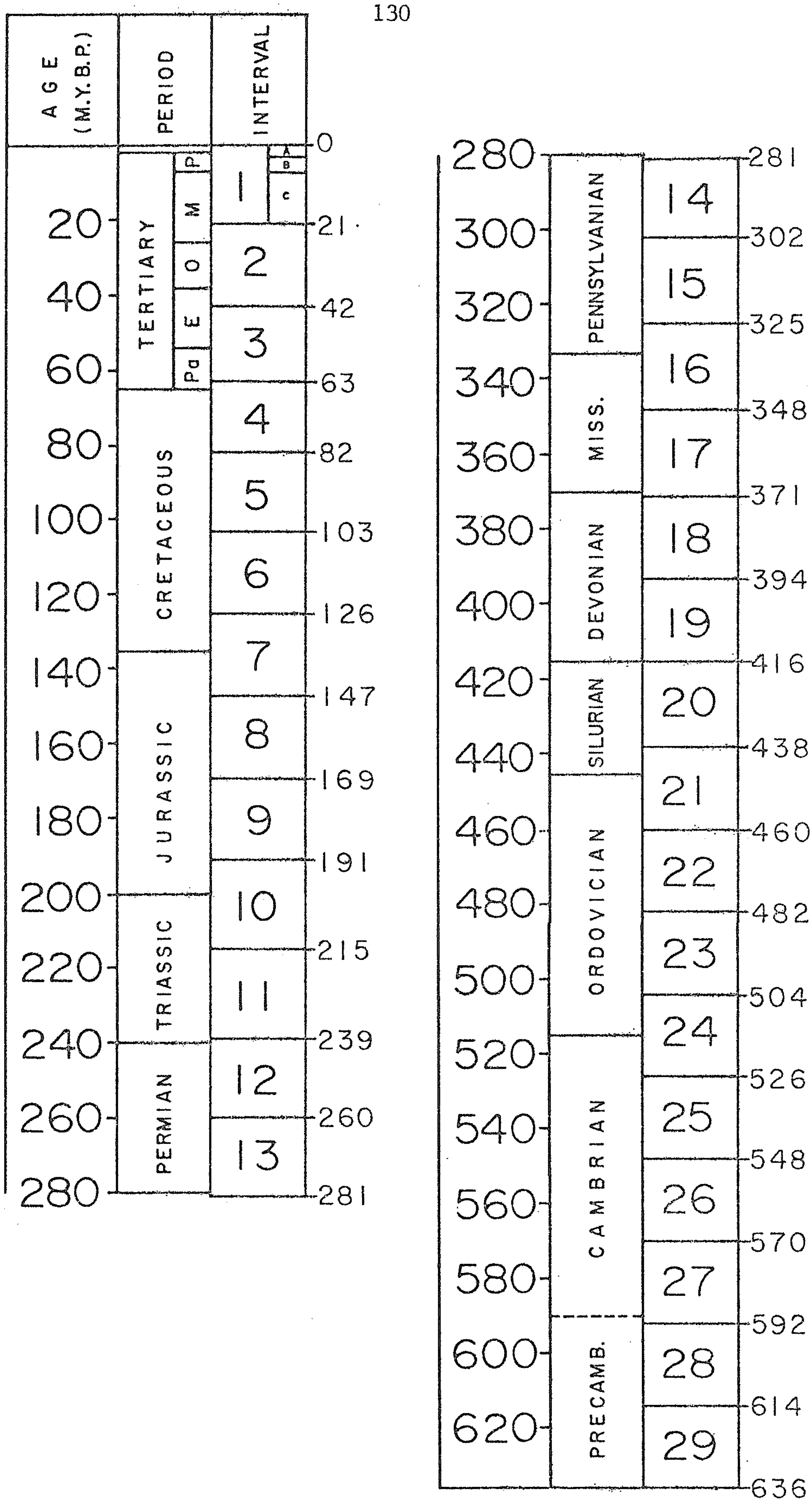
Third, the usual procedure of calculating means and other Fisherian statistics (Fisher, 1953) on small samples of paleomagnetic vectors may be misleading. Even the most reliable paleomagnetic poles are not necessarily drawn from Fisherian distributions for every time interval. In relatively small samples of poles (commonly with less than 10 unit vectors per period), one highly anomalous pole can significantly bias the estimated mean and distort estimates of its confidence limits.

Suggested Improvements for Constructing APW Paths

1. Demarcation of Uniform Time Intervals

To avoid the disadvantages inherent in computing poles for entire geologic periods of non-uniform length, the Phanerozoic and very latest Precambrian is here divided into 29 intervals, each about 22 m.y. long. Figure 4.1 shows the demarcation of these time intervals on the revised Geological Society of London Phanerozoic Time Scale recommended by Lambert (1971). Because paleomagnetic data are abundant and more precise age control is available for the late Cenozoic, the most recent time interval is further broken into three subgroups.

Figure 4.1 Diagram showing the relationships between the 29 time intervals referred to in this study, the geological periods, and corresponding absolute ages according to the revised Geological Society of London Phanerozoic Time Scale (Lambert, 1971).


AGE (M.Y.B.P.)
PERIOD
INTERVAL
TERTIARY
P
M
O
E
Pa
CRETACEOUS
JURASSIC
TRIASSIC
PERMIAN
A
B
C
1
2
3
4
5
6
7
8
9
10
11
12
13
0
21
42
63
82
103
126
147
169
191
215
239
260
281
20
40
60
80
100
120
140
160
180
200
220
240
260
280
PENNSYLVANIAN
MISS.
DEVONIAN
SILURIAN
ORDOVICIAN
CAMBRIAN
PRECAMB.
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560
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600
620

The time intervals adopted here represent a compromise between the desire to maintain a standard time interval and the desire to avoid ambiguity in assigning a given paleomagnetic pole to a particular interval. If every time interval were of equal duration, relative rates of apparent polar motion would be indicated by the spacing between successive interval poles. If strictly equal divisions of time were used, however, the time represented by a given paleomagnetic pole (commonly several million years) might straddle two intervals; in this case, the paleomagnetic pole should be included in the determination of the typical pole for both intervals. This would tend to smooth out possibly significant details of the APW path. To minimize this undesirable effect, the intervals were chosen to conform more closely with the ages assigned to existing paleomagnetic poles; the interval length ranges between 19 and 24 m.y. In a few instances, however, the age range of a pole was sufficiently large that it still had to be incorporated in two intervals.

2. Selection and Weighting of Paleomagnetic Poles

Evaluation of published paleomagnetic data must precede any attempt to reconstruct the polar path. In revising the North American APW path, poles were generally excluded for which (1) stability of the magnetization has not been demonstrated by alternating-field, thermal, or chemical demagnetization techniques; (2) significant post-magnetization tectonic rotations have been suggested; (3) ages are in doubt by more than half (about 25 m.y.) of a typical

Phanerozoic geologic period; or (4) the computed mean, which must have been based on at least 10 independently oriented samples, has an α_{95} (half-angle of the cone of 95% confidence; Fisher, 1953) greater than 15°. (This last criterion is not completely rigorous, because authors vary in assigning unit weight to specimen, sample, or site directions, which can greatly affect the reported confidence limits.)

In addition to an assessment of the reliability of the published data, it is imperative that only true paleomagnetic poles (i.e., poles that have averaged out secular variation) be given unit weight in calculating the interval poles. To be considered a paleomagnetic pole in this study, a pole must represent an average of directions over at least 30,000 years, in accordance with McElhinny and Merrill's (1975) conclusion that more than 10,000 years seems to be required to average out secular variation effects. For any interval, if the number of poles judged to have insufficiently averaged out secular variation was at least 5, these were combined to form a composite paleomagnetic pole that was then treated as a unit vector. Moreover, poles from two or more studies of the same rock unit representing the same interval of geologic time were averaged to form a single paleomagnetic pole. These procedures, while substantially reducing the number of vectors given unit weight, help prevent misrepresentation of an interval pole by some small portion of the total time or by some unrecognized local tectonic rotation.

3. Computing Interval Poles by Determining the Mode Rather than the Mean

To avoid bias caused by outliers in small samples of poles, interval poles were computed by determining the mode rather than the mean. Because traditional methods of determining the mode (e.g., Knopf and Ingerson, 1938, p. 245) are of little use when applied to samples as small as those encountered in this study, a new technique for determining the mode was devised and incorporated in a FORTRAN computer program. (A complete description of this technique and its applications in paleomagnetism has been given in Chapter 1, and a copy of the program is included as Appendix A.) Using the mode rather than the mean seems preferable for constructing APW paths, because the mode is unaffected by outliers; hence, its position is less likely to change as more data are accumulated.

Review of North American Phanerozoic Paleomagnetic Data

This section is an interval-by-interval review of Phanerozoic paleomagnetic data that relate to apparent polar wandering with respect to North America. All data are discussed except those derived from (1) studies made prior to the development of demagnetization techniques for testing the stability of magnetization, and (2) studies of a purely rock magnetic or magnetostratigraphic nature. Discussions of data from each interval proceed uniformly in the following order: (1) Introductory comments summarizing the number of paleomagnetic investigations, the rock types involved, the geographic distribution of sampling sites, the number of poles appearing in Table 4.1, and the sources from which they are derived; (2) Identification of any poles (e.g., composite paleomagnetic poles) that are not taken directly from the original references; (3) Enumeration of the poles given unit weight in determining the mode and/or mean for the interval (i.e., the "interval pole"); (4) Enumeration of any poles excluded from the determination of the interval pole, as well as the reasons for their exclusion; and (5) Concluding remarks on the rate or direction of apparent polar wandering implied by the relation of a given interval pole to poles from adjacent intervals.

Pole numbers cited in the text are cross-referenced to those listed in Table 4.1 and plotted in the figures. Entries in Table 4.1 are ordered roughly according to the ages of the rock units studied. The ages of the rock units are taken directly from the original references (or from references found therein), except where noted otherwise. In most cases, the magnetizations represented by poles listed in Table 4.1 are thought to have been acquired penecontemporaneously with deposition or emplacement of the rock units.

Table 4.1

PHANEROZOIC PALEOMAGNETIC POLES FROM NORTH AMERICA

Rock Unit, Location	Reference	Age (m.y.B.P.)†	Intrvl	Pole No.††	Position	Weight$^{\gamma}$
Aiyansh flow, B. C.	Symons, 1975	Q(1650±40 A.D.)	1A	(1)		0.6
Cascade lavas, Ore.	Heinrichs, 1973	Q(0.000-0.012)	1A	(1)		0.6
Mono Lake seds, Calif.	Denham & Cox, 1971	Q(0.013-0.030)	1A	(1)		0.6
Tills, Alberta	Barendregt _et al._ 1977	Q	1A	(1)		0.6
Port Stanley Till, Ont.	Gravenor _et al._ 1973 ; Stupavsky _et al._, 1974a	Q(0.013-0.015)	1A	(1)		0.6
Leaside & Sunnybrook tills, Ont.	Stupavsky _et al._ 1974b	Q(0.013-0.015)	1A	(1)		1.1
St. Joseph Till, Ont.	Gravenor & Stupavsky, 1976	Q(0.013)	1A	(1)		1.1
Seds, Baja Calif., Mex.	Strangway _et al._, 1971	a. Q b. TP	1A 1B	(1) -		0.6 se
Lake Tecopa seds, Calif.	Hillhouse & Cox, 1976	Q(0.7)	1A	(1)		1.1
Yellowstone tuffs, Wy., Mont., Ida.	Reynolds, 1977	Q(0.60,1.22)	1A	(1)		1.1

Table 4.1 - continued

Rock Unit, Location	Reference	Age (m.y.B.P.)	Intrvl	Pole No.	Position	Weight
Valley of Mex. volcs, Mex.	Mooser <u>et</u> <u>al</u>., 1974	Q	1A	2	84N, 50E*	9.1
Iztaccihautl volcs, Mex.	Steele, 1971	Q (-TP?)	1A	3	89N, 4E	9.1
Suttle Lake lavas, Ore.	Cameron & Stone, 1970	Q	1A	4	88N, 272E	9.1
Mt. Edgecumbe lavas, Alaska	Cameron & Stone, 1970	Q	1A	5	75N, 183E	9.1
Mt. Griggs volcs, Alaska	Cameron & Stone, 1970	Q	1A	6	77N, 151E	9.1
Valles Caldera volcs, New Mex.	Doell <u>et</u> <u>al</u>., 1968	Q(0.4-1.4)	1A	7	83N, 83E*	9.1
San Pedro V. seds, Ariz.	Johnson <u>et</u> <u>al</u>., 1975	Q-TP (0-3)	1A	8	82N, 36E*	9.1
Lousetown Fm. volcs, Nev.	Heinrichs, 1967	a. Q-TP (1-3) b. TP (7?)	1A 1B	9 --	67N, 357E 16N, 204E	9.1 ex
Volcs, Montana a. Norris b. Virginia City c. Bvrhd. V., H.P.C.	Hanna, 1967	 a. Q-TP b. Q-TP c. TP-ℓTM	 1A,1B 1A,1B 1B,1C	 (1),(13) 10 (13),(18)	 64N, 294E 60N, 241E 72N, 251E	 0.6,1.4 9.1,16.7 1.4,1.9
Brwn bslt dikes, B.C.	Symons, 1968	Q(-ℓTM)	1A,1B	(1),(13)		0.6,1.4

Table 4.1 - continued

Rock Unit, Location	Reference	Age (m.y.B.P.)	Intrvl	Pole No.	Position	Weight
Mt. Edziza volcs, B.C.	Souther & Symons, 1973	a. Q-TP(0-3)	1A	12	81N, 328E*	9.1
		b. TP (3-6)	1B	14	85N, 87E*	16.7
Lavas: a) N. Mex.	Ozima et al., 1967;	a. ℓTP (3.6-4.5)	1B	15	89N, 61E*	16.7
b) Arizona	Kono et al., 1967	b. *m*TP	1B	(13)		5.6
Sonoma volcs, Calif.	Mankinen, 1972	TP (2.9-5.3)	1B	16	79N, 83E	16.7
Wrangell volcs, Alaska	Bingham & Stone, 1976	TP (3.4)	1B	(13)		2.8
Chamita Fm. seds, N. Mex.	MacFadden, 1977	TP-TM (4.5-6.0)	1B	17	76N, 139E*	16.7
R. Gr. de Sant. volcs, Mex	Watkins et al., 1971	a. TP (4.7-5.4)	1B	(13)		4.2
		b. TM (8.8-9.3)	1C	(18)		7.6
Abert Rim lavas, Ore.	Watkins, 1965a	TM (~15)	1C	--		inc.(19,20)
Columbia Plat. lavas, Ore.	Watkins, 1965b	TM (15-25)	1C	--		inc.(19,20)
Steens Mtn. bslt, Ore.	Watkins, 1969	TM (~15)	1C	--		inc.(19,20)
Santa Rosa volcs, Nev.	Larson et al., 1971	TM (15)	1C	--		inc.(19-20)
Basalts, Ore., Wash., Ida.	Watkins & Baksi, 1974	TM (13-16)	1C	19	82N, 178E*	25.0
			1C	20	79N, 319E*	tc

Table 4.1 - continued

Rock Unit, Location	Reference	Age (m.y.B.P.)	Intrvl	Pole No.	Position	Weight
Cariboo Plat. volc plugs, B.C.	Symons, 1969a	TM (10-13)	1C	21	85N, 217E*	25.0
Cariboo Plat. bslts, B.C.	Symons, 1969b	TM (10-14)	1C			
Lavas & tuff , Ariz.	Young & Brennan, 1974	TM (17-18)	1C	22	84N, 34E	25.0
Mt. Barr Cmplx, B.C.	Symons, 1973c	TM (16-21)	1C	(18)		5.7
Batholiths, Wash.	Beske *et al.*, 1973					
a. Snoqualmie		a. TM (16-18)	1C	(18)		7.6
b. Grotto		b. TM (25-26)	2	(23)		7.1
Lavas & tuffs, Mex.	Nairn *et al.*, 1975a	TM-TO (23-31)	2	24	71N, 157E	14.3
Lavas & tuffs, Ut., Ore., Calif.	Grommé & McKee, 1971	TM-TO (22-37)	2	25	78N, 146E	14.3
Ignimbrites, Nev.	Gose, 1970	a. TM (17.8)	1C	(18)		1.9
		b. TM-TO (21-30)	2			inc.(25)
Ash flows, Nev. & Utah	Grommé *et al.*, 1972	TM-TO (22-35)	2			inc.(25)
Volcs, Nev. & Utah	Nairn *et al.*, 1975b	TM-TO (20-31)	2			inc.(25)
Needles R. Fm., Utah	Best *et al.*, 1973; Shuey *et al.*, 1976	TO (30±1)	2			inc.(25)

Table 4.1 - continued

Rock Unit, Location	Reference	Age (m.y.B.P.)	Intrvl	Pole No.	Position	Weight
Lovejoy bslt, Calif.	Grommé, 1965	TM (22-24)	2	--		st
Yarmony Mtn. flows, Colo.	York _et al._, 1971	TM (21-24)	2	(23)		4.8
Spanish Peak dikes, Colo.	Larson & Strangway, 1969	TM (22-26)**	2	26	81N, 211E	14.3
Volcs, Ariz.	Strangway _et al._, 1971	TO (25-30)	2	27	75N, 11E	ab
Buck Hill volcs, Tex.	Gilliland _et al._, 1969	TO (27-36)	2	28	81N, 89E	14.3
San Juan volcs, Colo.	Beck _et al._, 1977; Diehl _et al._, 1974; Tanaka & Kono, 1973	TO (23-30)	2	29	85N, 114E	14.3
Marys Peak sill, Ore.	Clark, 1969	TO (30)	2	30	63N, 16E	ex,tc?
Mistastin Lake volcs, Labr.	Currie & Larochelle, 1969	TE (38±4)**	2	(23)	86N, 118E	2.4
East Sooke stock, Vanc. I, B.C.	Symons, 1973d	TE (39)	2	31	70N, 151E	tc
Hope Pluton, B.C.	Symons, 1973c	TO-TE (35-41)	2	32	88N, 208E	14.3
Intrusions, Vanc. I, B.C.	Symons, 1971c	a. TO (37±5) b. TE,TPA (50-59)	2 3	-- --		n.p., tc n.p., tc

Table 4.1 - continued

Rock Unit, Location	Reference	Age (m.y.B.P.)	Intrvl	Pole No.	Position	Weight
Younger plutons, B.C.	Hicken & Irving, 1977	TO-TPA (26-62)	2, 3	32.5	70N, 206E	tc,dt
Granite Falls stock, Wash.	Beske *et al.*, 1973	TE (43±1)	3	(34)		2.4
Seds & volcs, Ore.	Simpson & Cox, 1977					
a. Yachats		a. ℓTE	2	33	58N, 308E	tc
b. Tyee-Flournoy		b. mTE	3	35	49N, 305E	tc
c. Siletz R.		c. eTE	3	36	48N, 306E	tc
Tolstoi Fm., Alaska	Stone & Packer, 1977	TE (∼45)	3	36.5	69N, 303E	tc
Quottoon & Kasiks pltns, B.C.	Symons, 1974a	TE (47±6)	3	37	76N, 216E	14.3
Intrusions, Virg.	Løvlie & Opdyke, 1974	TE (47±1)	3	38	88N, 46E	14.3
Green R. seds, Colo.	Strangway & McMahon, 1973	TE (45-49)	3	39	19N, 225E	ex, n.p.
Rattlesnake Hills intr., Wyo	Shive *et al.*, 1977	TE (44-49)	3	40	83N, 208E	14.3
Absaroka volcs, Wyo.	Shive & Pruss, 1977	TE (48)	3	41	84N, 177E	14.3
Lavas, Baffin Isl., Nthwst. Terr.	Deutsch *et al.*, 1971	TPA (58±2)	3	42	83N, 305E	14.3

Table 4.1 - continued

Rock Unit, Location	Reference	Age (m.y.B.P.)	Intrvl	Pole No.	Position	Weight
Nacimiento Fm. seds, N. Mex.	Butler & Taylor, 1978	mTPA (∼59-63)	3	43	76N, 148E	14.3
Flows & sill, Colo.	Hoblitt & Larson, 1975; Larson et al. 1969	TPA (59-64)	3	(34)		9.6
Intrusion, Colo.	McMahon & Strangway, 1968 a,b	TPA (∼60)	3	--		n.p.,tc?
Twin Sisters Dunite, Wash.	Beck, 1975	TE-TPA	3	44	28N, 298E	tc
Idaho Batholith, Ida.	Beck et al., 1972	a. TE (42-49)	3	--		n.p.,tc?
		b. ℓK (73)	4	--		n.p.,tc?
Masset Fm.,B.C.	Hicken & Irving, 1977	TPA (62)	3	45	72N, 263E	tc
Dikes & intr., B.C.	Symons, 1968	TE-ℓK (47-77)	3,4	--		dt, tc?
Boulder Bath., Mont.	Hanna, 1973a	ℓK (68-78)	4	46	73N, 249E	25.0
Elkhorn Mtn. volcs., Mont.	Hanna, 1967; Hanna, 1973b					
a. Bvrhd Valley		a. TE	3	(34)	66N, 239E	2.4
b. Bull Mtn.		b. ℓK (78±2)	4	47	69N, 189E	25.0
Ecstall Pluton, B.C.	Symons, 1974a	ℓK (75±12)	4	48	68N, 313E	tc

Table 4.1 - continued

Rock Unit, Location	Reference	Age (m.y.B.P.)	Intrvl	Pole No.	Position	Weight
Chignik Fm., Alaska	Stone & Packer, 1977	ℓK (~70-75)	4	48.5	39N, 343E	tc
Difunta Grp., Mex.	Nairn, 1976	ℓK	4	48.7	58N, 169E	tc?
Mesaverde Grp. seds, Wy. & Ut.	Kilbourne, 1969	ℓK (70-76)	4	49	65N, 198E	25.0
Seds, Wy., Colo., Kan.	Shive & Frerichs, 1974					
a. Hilliard Fm. (Kemmerer)		a. ℓK (~69-82)	4	50	62N, 178E	25.0
b. Niobrara Fm. (Pueblo)		b. ℓK (~82-88)	5	51	66N, 187E	33.3
Sierra Nevada plutons, Calif.	Currie _et al._, 1963; Grommé & Merrill, 1965	ℓK (79-89)	5	52	69N, 195E	33.3
Stevens Pass granite, Wash.	Beck & Noson, 1972	ℓK (~85)	5	53	73N, 60E	tc
Howe Sound plutons, B.C.	Symons, 1973b	ℓK (94±6)	5	54	83N, 129E	tc
Alkalic cmplxs., Ark.	Scharon & Hsu, 1969	K (88-105)	5	55	65N, 187E	33.3
Idaho Bath., Ida.	Beck _et al._, 1972	eK (115-124)	6	--		n.p.,tc
S. Calif. Bath., Calif.	Teissere & Beck, 1973	eK (100,121)	6	56	86N, 23E	tc

Table 4.1 - continued

Rock Unit, Location	Reference	Age (m.y.B.P.)	Intrvl	Pole No.	Position	Weight
Coast plut. cmplx, B.C.	Symons, 1977					
a. Stephens Isl.		a. ℓK (102±8)	6	57	67N, 21E	tc
b. Captain Cove		b. ℓK (109±6)	6	58	72N, 351E	tc
Isachsen diab., Nthw. Terr.	Larochelle & Black, 1963	ℓK (102-110)	6	59	69N, 180E	50.0
Monteregian Hills, Que.	Larochelle, 1968, 1969	ℓK (100-120)	6	60	71N, 189E	50.0
White Mtn. magma ser., N.H. & Ver.						
a. sites 1,2,3,9,10, A,B	a. Opdyke & Wensink, 1966	a. ℓK (113-121)	6	(61)		an
b. sites dB1	b. de Boer, in Bouley, 1971	b. ℓK (114-118)	6	(61)		an
c. sites B8, B10	c. Bouley, 1971	c. ℓK (118-121)	6	(61)		an
Topley Intr., B.C.	Symons, 1973a	a. ℓK (112-114)	6	--		n.p.
		b. ℓJ (136-144)	7	63	70N, 129E	25.0
		c. ℓ-mJ (138-163)	8	--		n.p.
Lamp. dikes, Newf.	Deutsch & Rao, 1977b	ℓK-ℓJ (115-144)	6,7	62	83N, 195E	an,dt
Gil Isl. pluton, B.C.	Symons, 1977	ℓJ (136±3)	7	64	70N, 3E	tc
Guad. & Bucks baths, Calif.	Grommé *et al.*, 1967	ℓK-ℓJ (129-142)	7	65	54N, 186E	25.0

Table 4.1 - continued

Rock Unit, Location	Reference	Age (m.y.B.P.)	Intrvl	Pole No.	Position	Weight
Morrison Fm., Colo.	Steiner & Helsley, 1975	ℓJ (∿143)				
a. upper			7	66	68N, 162E	25.0
b. lower			7	67	61N, 142E	25.0
Todos Santos Fm., Mex.	Guerrero & Helsley, 1974	ℓJ	7	68	70N, 160E	tc?
Red Mtn. intr., Calif.	Saad, 1969	(ℓK)-ℓJ	8	69	56N, 310E	dt,tc
			8	70	66N, 228E	dt,tc
			8	71	3N, 324E	dt,tc
Franciscan Fm. ign., Calif.	Grommé & Gluskoter, 1965	eK-ℓJ	8	72	29N, 316E	dt,tc
Idaho Bath., Ida.	Beck <u>et al</u>., 1972	ℓJ (156)	8	--		n.p.,tc
Sed. rocks, Alaska	Packer & Stone, 1972, 1974; Stone & Packer, 1977	ℓ-mJ (∿140-160)	7,8	73	48N, 295E	tc
Isl. Intr., Vanc. I, B.C.	Symons, 1970	ℓ-mJ (159±10)	8	74	79N, 240E	tc
Summerville Fm., Utah	Steiner, 1978; Steiner & Helsley, 1972	ℓJ (∿153)	8	75	68N, 111E	100.0 (-50.0)
White Mtn. magma ser. N.H. & Ver.						
a. sites 4,5	a. Opdyke & Wensink, 1966	a. mJ (∿157-164)	8	(76)		0.0(-12.5)

Table 4.1 - continued

Rock Unit, Location	Reference	Age (m.y.B.P.)	Intrvl	Pole No.	Position	Weight
b. sites dB8,dB9,dB10	b. de Boer, in Bouley, 1971	b. *mJ* (157-168)	8	(76)		0.0(-18.8)
c. sites B2,B4,B6	c. Bouley, 1971	c. *mJ* (157-164)	8	(76)		0.0(-18.8)
White Mtn. magma ser., N.H. & Ver.						
a. sites 6,7,8,11,12	a. Opdyke & Wensink, 1966	a. *m-eJ* (~171-185)	9	(77)		an
b. sites dB4,dB5,dB6	b. de Boer, in Bouley, 1971	b. *m-eJ* (168-180)	9	(77)		an
Tulameen Cmplx., B.C.	Symons, 1974b	*m-e* J (172-188)	9	78	46N, 119E	tc
Anticosti I. dike, Que.	Larochelle, 1971	*m-eJ* (178±8)	9	79	76N, 85E	n.p.,tc?
Carmel Fm., Utah	Johnson, 1972; Johnson & Nairn, 1972	*mJ*	9	80	76N, 307E	se?
Navajo Ss., Utah	Johnson, 1972; Johnson & Nairn, 1972; Johnson, 1976	*eJ*	9	81	80N, 71E	se
Appalachian dikes, east U.S.	de Boer, 1967	*m-eJ* (~180**)	9	---		su
Appal. dikes, Ct. & Md.	Smith, 1976a	*m-eJ* (~180**)	9	82	69N, 101E	50.0
Diabase intr., Penn.	Beck, 1965, 1972	*m-eJ* (~180**)	9	83	62N, 105E	50.0

Table 4.1 - continued

Rock Unit, Location	Reference	Age (m.y.B.P.)	Intrvl	Pole No.	Position	Weight
North Mtn. bslt., N. Sc.	Carmichael & Palmer, 1968; Bouley 1969; Larochelle, 1967b	eJ-ℓT	10	(84)		3.1
Granby & Holyoke lavas, Mass.	Irving & Banks, 1961	eJ-ℓTR (193±6)	10	(84)		3.1
Newark Grp. ign & sed, N.J. & Penn.	Opdyke, 1961	eJ-ℓTR (193±6)	10	(84)		4.6
Newark Grp. ign, Ct.	de Boer, 1968	eJ-ℓTR	10	(84)		9.2
Shelburne dike, N. Sc.	Larochelle & Wanless, 1966	eJ-ℓTR (197±32)	9,10	--		dt,n.p.
Copper Mt. intrs., B.C.	Symons, 1973e	eJ-ℓTR (194±8)	10	85	68N, 351E	tc
Guichon Bath., B.C.	Symons, 1971b	eJ-ℓTR (198±8)	10	86	66N, 13E	tc
Karmutsen bslts, Vanc. I, B.C.	a. Symons, 1971a	ℓTR	10	87	67N, 36E	tc
			10	88	13N, 45E	tc
	b. Irving & Yole, 1972	ℓTR	10	89	70N, 205E	tc
			10	90	21N, 61E	tc
Kayenta Fm., Utah	a. Johnson, 1972a, b, 1976	eJ-ℓTR	10	--		su
	b. Steiner & Helsley, 1974a	eJ-ℓTR	10	91	62N, 74E	20.0
	c. Steiner & Helsley, 1974b, 1972	eJ-ℓTR	10	91.1	47N, 83E	an
			10	91.2	62N, 354E	an
			10	91.3	34N, 101E	an
			10	91.4	76N, 104E	an

Table 4.1 - continued

Rock Unit, Location	Reference	Age (m.y.B.P.)	Intrvl	Pole No.	Position	Weight
Wingate & Chinle Fms. Utah	Reeve, 1975	ℓTR	10	92	58N, 68E	20.0
Chinle Fm., New Mex.	Reeve & Helsley, 1972	ℓTR	10	93	58N, 79E	20.0
Manicouagan ign, Que.	a. Larochelle & Currie, 1967 b. Robertson, 1967	ℓTR (210±4**) ℓTR (210±4**)	10	94	61N, 89E 57N, 89E	10.0 10.0
Nikolai Grnst., Alaska	Hillhouse, 1977	ℓ-*m*TR	10,11	95	2N, 146E	dt,tc
Nazas Fm., Mex.	Nairn, 1976	*e*J-*m*TR	10,11	95.5	76N, 119E	dt,tc?
Chugwater Fm., Wy.	Grubbs & Van der Voo, 1976; Van der Voo & Grubbs, 1977	*e*TR (?)	11	96	49N, 112E	dt,tc?
Upper Moenkopi Fm., Colo.	Helsley & Steiner, 1974	*e*TR	11	97	55N, 103E	33.3
Lower Moenkopi Fm., Colo.	a. Helsley, 1969 b. Baag & Helsley, 1974	*e*TR	11 11	98	57N, 98E 57N, 100E	16.7 16.7
State Bridge Fm., Colo.	Christensen & Helsley, 1974a	*e*TR	11	99	52N, 106E	33.3
Upper Maroon Fm., Colo	McMahon & Strangway, 1968a,b	*e*TR(?) (Post-Kiaman)	11	100	61N, 101E	n.c.?

Table 4.1 - continued

Rock Unit, Location	Reference	Age (m.y.B.P.)	Intrvl	Pole No.	Position	Weight
Hoskinnini Mbr., Utah	a. Farrell & May, 1969	eTR (-ℓP?)	11	101	50N, 121E	su
	b. Helsley & Scott, in Scott, 1975		11	102	56N, 104E	ab
Upper Maroon Fm., Colo.	Christensen & Helsley, 1974b	ℓP (Kiaman)	12	103	42N, 119E	50.0
Malpeque Bay sill, P.E.I.	Larochelle, 1967a	ℓ(?)P	12	104	51N, 111E	n.p.,dt,M
Pegmatites, Ct.	de Boer & Brookins, 1972	ℓ-mP (255±5)	12	105	35N, 126E*	n.p.?,M
Red beds, sthwst U.S.	Peterson & Nairn, 1971	a. ℓ-mP (Och. & Guad.)	12	106	52N, 124E*	50.0
		b. eP (Leon. & Wolf.)	13	(107)		24.2
Red beds, Colo.	McMahon & Strangway, 1968a,b	ℓP-m(?)PN	12-15(?)			n.c.?
Halgaito Tongue, Utah	Farrell & May, 1969	eP	13	(107)		3.0
Cutler Fm., Colo.	Helsley, 1971	eP	13	(107)		6.1
Cutler Grp., Utah	Gose & Helsley, 1972	eP	13	108	51N, 111E*	33.3
Upper Casper Fm., Wyo.	Diehl & Shive, 1976	eP (Wolf.)	13	109	52N, 125E	33.3

Table 4.1 - continued

Rock Unit, Location	Reference	Age (m.y.B.P.)	Intrvl	Pole No.	Position	Weight
Honaker Tr. Fm., Utah	Scott & Helsley, 1975	(*e*P?-)*ℓ*PN	13,14	110	55N, 106E	se?
Dunkard Series, W. Va.	Helsley, 1965	*ℓ*PN**	14	111	49N, 122E	33.3
Lower Maroon Fm., Colo.	Christensen & Helsley, 1974b	*ℓ*PN	14	112	41N, 133E	33.3
Red beds, P.E.I.	Black, 1964	*e*P-*ℓ*PN	14	113	40N, 126E	M
Red beds, eastern Canada	Roy, 1966					
a. Red beds, P.E.I.		*e*P-*ℓ*PN	14	114	42N, 133E	M
b. Pictou Grp., N.Sc.		*ℓ*PN	14	115	41N, 132E	M
c. Bonaventure Fm., Que.		*ℓ*PN	14	116	38N, 133E	M
Hurley Crk. Fm., N. Bruns.	Roy <u>et al</u>., 1968	*ℓ*PN	14	117	39N, 125E	M
Casper Fm., Wyo.	Diehl & Shive, 1976					
a. middle		*ℓ*PN	14	118	47N, 131E	33.3
b. lower		*m*PN (Desm.)	15	119	42N, 131E	50.0
Coal Pole, Ohio & W. Va.	Kopacz & Noltimier, 1976	*m*PN	15	120	37N, 132E	50.0
Cumberland Grp., N. Sc.	Roy, 1969	*m*PN	15	121	36N, 125E	M
Hopewell Grp., N. Bruns.	a. Roy & Park, 1969	*e*PN-*ℓ*M	16	122	34N, 118E	M
	b. Roy & Park, 1974		16	123	36N, 123E	M

Table 4.1 - continued

Rock Unit, Location	Reference	Age (m.y.B.P.)	Intrvl	Pole No.	Position	Weight
Maringouin Fm., N. Bruns.	Roy & Robertson, 1968	ℓM	16	124	34N, 117E	M
Mauch Chunk Fm., Penn.	Knowles & Opdyke, 1968	ℓM	16	125	43N, 127E	100.0
Codroy Grp., W. Newf.	Black, 1964	M	16,17	126	30N, 127E	WN
St. Joe Ls., Ark.	Scott, 1975	eM	17	127	41N, 132E	50.0
			17	128	37N, 135E	50.0
Mafic cmplx, Mass.	Schutts _et al_., 1976	eM-ℓD (370±10)	17,18	129	23N, 126E	M
Perry Fm., ign & seds, N. Bruns.	Black, 1964	ℓD	18	130	26N, 109E	M
			18	131	35N, 121E	M
Perry Fm.,ign & seds, N. Bruns.	Robertson _et al_., 1968	ℓD	18	132	32N, 118E	st,M
Perry Fm. volcs, Maine	Phillips & Heroy, 1966	ℓD	18	133	24N, 128E	M
Belchertown pluton, Mass.	Ashwal & Hargraves, 1977	ℓD (-eD?) (380±5)	18,19	134	48N, 147E	M,tc
Catskill Fm., Penn	Phillips, 1966	ℓD	18	135	43N, 130E	20.0
Catskill Fm., Penn, Md., W. Va.	French, 1976	ℓD	18	136	45N, 128E	20.0

Table 4.1 - continued

Rock unit, Location	Reference	Age (m.y.B.P.)	Intrvl	Pole No.	Position	Weight
Catskill Fm., N.Y.	Kent & Opdyke, 1977a	ℓ-mD	18	137	47N, 117E	20.0
Temple Butte Ls., Ariz.	Elston & Bressler, 1977	ℓ-mD	18	138	53N, 115E	20.0
Martin Fm., Ariz.	Elston & Bressler, 1977	ℓ-mD	18	139	56N, 109E	20.0
Upper Columbus & Delaware Ls., Ohio	Martin, 1975	mD	19	140	46N, 120E	100.0
Lower Columbus & Raisin River Dol., Ohio	Martin, 1975	eD	19	141	24N, 164E	an
Clam Bank Grp., W. Newf.	Black, 1964	eD	19	142	28N, 146E	WN
Bloomsburg Fm., Penn. & Md.	Roy et al., 1967	ℓS	20	143	32N, 102E	50.0
Tymochtee Fm., Ohio	McMahon, 1974	ℓS	20	144	37N, 108E*	50.0
Rose Hill Fm., Va. & W. Va.	French & Van der Voo, 1977	mS	20	145	26N, 116E	n.c.?
Castanea Fm., Penn.	Opdyke in McElhinny & Opdyke, 1973	eS	21	146	21N, 105E	ab
Juniata Fm., Va., W. Va., Penn.	Van der Voo & French, 1977	ℓO	21	147	32N, 114E	100.0

Table 4.1 - continued

Rock Unit, Location	Reference	Age (m.y.B.P.)	Intrvl	Pole No.	Position	Weight
Beemerville Cmplx, N.J.	Proko & Hargraves, 1973	ℓO (435±20)	21	148	35N, 126E	st
Trenton Ls., N.Y.	McElhinny & Opdyke, 1973	ℓ-mO	22	149	36N, 114E	100.0
Volcanics, Que.	Seguin, 1977	mO				
a. A.F.			23	149.2	19N, 164E	tc? M?
b. Thermal			23	149.4	6N, 191E	tc? M?
Wabana Grp., E. Newf.	Deutsch & Rao, 1970	m-eO	23	150	28N, 192E	tc
West Spr. Crk. Fm., Okla.	Steiner, 1973	eO	23	151	46N, 129E	se?
Moreton's H. bslt., C. Newf.	Deutsch & Rao, 1977a	m-eO	23,24	152	32N, 130E	st,tc
St. George Ls., W. Newf.	Deutsch & Rao, 1977a	eO	23,24	153	30N, 119E	WN
St. George Ls., W. Newf.	Beales _et al._, 1974	eO	23,24	154	29N, 125E	WN
Intrusives, Colo.	Larson & Mutschler, 1971	eO-ℓP€ (485-704)				
a. A.F.			23+	155	37N, 122E	dt
b. Thermal			23+	156	44N, 100E	dt

Table 4.1 - continued

Rock Unit, Location	Reference	Age (m.y.B.P.)	Intrvl	Pole No.	Position	Weight
Intrusives, Colo.	French *et al.*, 1977	eO-ℓPЄ (485-704)				
a. Group I			23+	157	15N, 142E	dt
b. Group II			23+	158	5N, 174E	dt
c. Group III			23+	159	48N, 107E	dt,st
Wichita granites, Okla.	Ku *et al.*, 1967	ℓЄ (525±25)	24,25	160	30N, 142E*	st,n.c.?
Wichita granites, Okla.	Spall, 1968a	ℓЄ (525±25)				
a. A.F.			24,25	161	34N, 118E	an?
b. Thermal			24,25	162	2N, 147E	an?
Wichita granites, Okla.	Spall, 1970	ℓЄ (525±25)				
a. A.F.			24,25	163	13N, 147E	st,an?
b. Thermal			24,25	164	4N, 164E	st,an?
Wichita granites, Okla.	Vincenz *et al.*, 1975	ℓЄ (525±25)				
a. A.F.			24,25	165	16N, 149E	st,an?
b. Thermal			24,25	166	31N, 138E	st,an?
Wilberns Fm., Texas	Van der Voo *et al.*, 1976	ℓЄ	24	167	6N, 159E	an?
Wilberns & Riley Fm., Texas	Reeve, 1975	ℓ-mЄ				
a. Point Peak			24,25	168	10N, 163E	an?
b. Lion Mtn.			24,25	169	41N, 127E	an?
Bonneterre dls, Mo.	Beales *et al.*, 1974	ℓЄ	25	170	35S, 190E	an?

Table 4.1 - continued

Rock Unit, Location	Reference	Age (m.y.B.P.)	Intrvl	Pole No.	Position	Weight
Lamotte Fm., Mo.	Al-Khafaji & Vincenz, 1971	ℓ€				
a. Group IIa			25	171	38S, 192E	an?
b. Group IIb			25	172	12S, 155E	an?
c. Group I			25	173	30N, 187E	an?
d. Locality E			25	174	32N, 137E	an?
Abrigo Fm., Ariz.	Elston & Bressler, 1977	ℓ-m€	25	175	59N, 89E	an?
Muav Ls., Ariz.	Elston & Bressler, 1977	ℓ-m€	25	176	55N, 110E	an?
Ophiolite cmplx., Que.	Seguin, 1976	m(?)€ (550)	25,26	177	13N, 146E	tc? M?
Lower Bonanza King Fm., Nev.	Gillett & Van Alstine, in prep.	m€	26	178	9N, 160E$^{\delta}$	33.3
Carrara Fm., Nev.	Gillett & Van Alstine, in prep.	m-e€	26	179	7N, 158E$^{\delta}$	33.3
Rome Fm., Tenn.	French, 1976	m-e€	26,27	180	38N, 144E	tc?
Tapeats Ss., Ariz.	Elston & Bressler, 1977	(m?)-e€	26,27	181	5N, 158E	33.3,50.0
Middle Wood Cnyn. Fm., Nev.	Gillett & Van Alstine, in prep.	e€	27	182	1S, 157E$^{\delta}$	50.0
Bradore Fm., W. Newf.	Rao & Deutsch, 1976	e€	27	183	29N, 167E	WN
Ratcliffe Brook Fm., N. Bruns.	Black, 1964	e€	27	184	10N, 124E	M

Table 4.1 - concluded

Rock Unit, Location	Reference	Age (m.y.B.P.)	Intrvl	Pole No.	Position	Weight
Cloud Mtn. bslt., W. Newf.	Deutsch & Rao, 1977a	ℯЄ-ℓPЄ (605±10)	28	185	5N, 172E	WN
Franklin ign., Canad. Arctic	Palmer & Hayatsu, 1975; Fahrig & Schwarz, 1973; Robertson & Baragar, 1972; Fahrig *et al.*, 1971	ℓPЄ (625-675)				
a. Group A			29+	186	4S, 161E	dt
b. Group B			29+	187	8N, 166E	dt

Notes:

† Q (Quaternary); TP (Pliocene); TM (Miocene); TO (Oligocene); TE (Eocene); TPA (Paleocene); K (Cretaceous); J (Jurassic); TR (Triassic); P (Permian); PN (Pennsylvanian); M (Mississippian); D (Devonian); S (Silurian); O (Ordovician); Є (Cambrian); PЄ (Precambrian). Subdivisions of periods are denoted by: ℯ (early); *m* (middle); ℓ (late).

†† Pole number in parentheses indicates that a result has probably not averaged secular variation and hence was entered into a "composite pole" designated by the number in parentheses.

γ Numerical entry indicates % contribution to interval mean of Table 4.2. Alphabetical entry indicates pole excluded from interval pole calculations for the following reasons:

ab data in abstract form or preliminary
an anomalous
dt imprecisely dated (±>25 m.y.)
ex excursion of geomagnetic field
inc incorporated in pole (#)
M pole from "Maritime block" (Scott, 1975)
n.c. insufficient cleaning
n.p. not considered a paleomagnetic pole (probably represents <30,000 years)
se secondary magnetization

Table 4.1 Notes - continued

st $\alpha_{95} > 15°$
su superseded by later studies
tc tectonic rotation of sampling locality
WN pole from Western Newfoundland

* Pole not reported in original reference (see text).

** Age of pole differs from that reported in original reference (see text).

δ Pole corrected for 36° clockwise rotation with respect to Colorado Plateau.

Cenozoic

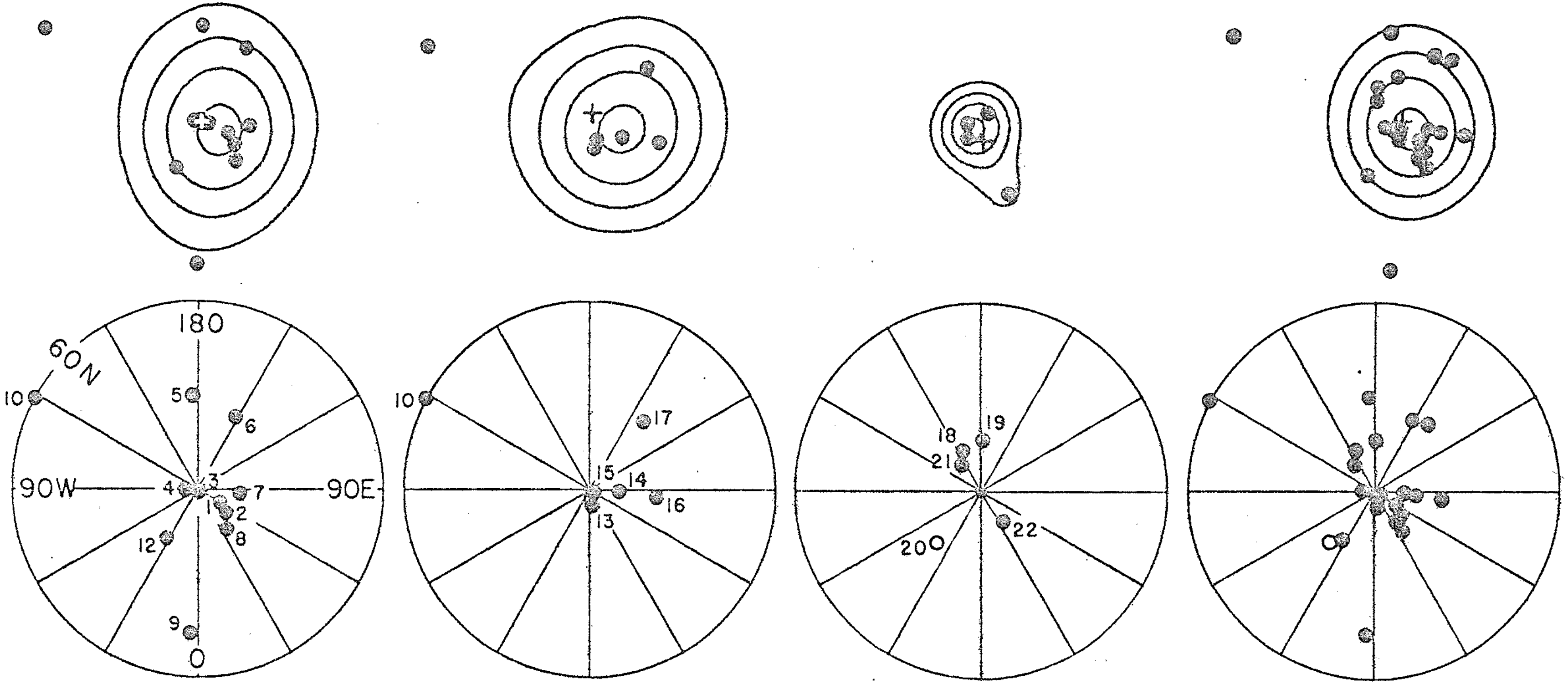
Interval 1A

In Table 4.1 are listed 20 paleomagnetic investigations of rocks less than 3 m.y. old (ca. late Pliocene or younger). These studies are predominantly of sediments and volcanic rocks from western U.S., Alaska, and Mexico. From these studies, 11 poles have been derived (Table 4.1, Figure 4.2), 5 of which do not appear in the original references.

Sixteen published poles for interval 1A probably represent less than 30,000 years, the minimum length considered in this paper to be adequate for averaging out effects of secular variation. These 16 poles have been combined to form composite pole 1, with a mode at 86°N, 60°E. Ten of these poles are from lavas and sediments less than 30,000 years old; two more were computed for beds above and below the Brunhes-Matuyama transition zone in sediments from Lake Tecopa, California (Hillhouse and Cox, 1976); two are from tuffs in the Yellowstone Park area (Lava Creek and Mesa Falls tuffs; Reynolds, 1977); one is from a single flow near Norris, Montana (Hanna, 1967); and one is from normally-magnetized brown basaltic dikes, British Columbia (Symons, 1968). (Most of the poles from igneous rocks studied by Symons (1968) are not considered further in this paper because of their imprecise ages and structurally complex settings.)

In addition to composite pole 1, four other poles from interval 1A do not appear in the original references. Pole 2, apparently from the Sierra del Chichinautzin and Sierra de Santa

Figure 4.2 North American paleomagnetic poles for interval 1 (0-21 m.y.B.P.; approximately post-early Miocene). Poles represented by solid symbols were given unit weight in the contour plots shown above the azimuthal equidistant projections. Poles represented by open symbols were excluded from the determination of the interval pole for reasons cited in Table 4.1 and the text. For intervals 1 to 11, numbers beside poles refer to the complete pole designations of Table 4.1. For enhanced clarity, however, the first digit of the Table 4.1 pole numbers has not been plotted for the Paleozoic (pre-interval 11). The contour interval is in 20% increments, ranging from 35% to 95% of the maximum probability at the mode. The cross on contour plots represents the mean of poles given unit weight and marks the pole of the projection on which contouring was performed.



INTERVAL IA (0–3 m.y.B.P.)

INTERVAL IB (3–7 m.y.B.P.)

INTERVAL IC (7–21 m.y.B.P.)

INTERVAL I (0–21 m.y.B.P.)

Catarina sites in volcanic rocks of central Mexico (Mooser et al., 1974), is pole 14.033 in the catalog of McElhinny and Cowley (1977). Pole 7, from volcanic rocks of the Valles Caldera, New Mexico (Doell et al., 1968), is pole 12136 in the catalog of Hicken et al. (1972). Pole 8, from sediments of the San Pedro Valley, Arizona (Johnson et al., 1975), is pole 14.060 of McElhinny and Cowley (1977). Pole 12, from volcanic rocks of Mt. Edziza, British Columbia (Souther and Symons, 1973), was calculated from those flows probably younger than 3 m.y. (sites A14 to A28; B45 to B53; and C1 to C32), excluding a few flow VGP's with $\alpha_{95} > 15°$.

Eleven poles were given unit weight in determining a pole most representative of interval 1A. These include poles 1, 2, 7, 8, and 12 (defined above), as well as pole 3 (Iztaccihautl volcanic rocks, Mexico; Steele, 1971), pole 4 (Suttle Lake lavas, Oregon; Cameron and Stone, 1970), pole 5 (Mt. Edgecumbe lavas, Alaska; Cameron and Stone, 1970), pole 6 (Mt. Griggs volcanic rocks, Alaska; Cameron and Stone, 1970), pole 9 (Lousetown Formation volcanic rocks, Nevada; Heinrichs, 1967), and pole 10 (volcanic rocks near Virginia City, Montana; Hanna, 1967). No paleomagnetic poles from interval 1A were excluded from the determination of the interval pole.

Except for pole 10 (which represents only 5 flows), the poles given unit weight in interval 1A cluster fairly well around a mode at 87°N, 50°E (Figure 4.2). This pole, which represents the past 3 m.y., is "far-sided" by 3° with respect to the present geographic pole. Wilson and McElhinny (1974) calculated a far-sidedness of 4°

based on a data set representing western North American poles for the past 25 m.y.

Interval 1B

In Table 4.1 are listed 11 paleomagnetic investigations of rocks with ages between 3 and 7 m.y. (ca. middle and early Pliocene). These studies are predominantly of volcanic rocks from western U.S., Alaska, and Mexico. From these studies, 6 poles have been derived (Table 4.1, Figure 4.2), 4 of which do not appear in the original references.

Twelve published poles for interval 1B were judged to have insufficiently averaged out secular variation. These 12 poles have been combined to form composite pole 13, with a mode at 88°N, 11°E. Two of these poles are from igneous rocks in Montana studied by Hanna (1967) (the single Norris flow and the Beaverhead Valley (Horse Prairie Creek) units); four are from middle Pliocene lavas from Arizona studied by Kono et al. (1967); two are from sections in the Wrangell Volcanics, Alaska (Bingham and Stone, 1976) that were not thought to represent excursions of the geomagnetic field (Wait Creek(c) and Air II sections); three are from the study by Watkins et al. (1971) of the Rio Grande de Santiago Volcanics, Mexico (ca. 5 m.y.B.P. sites 5, 6, and 7); and one is from normally-magnetized brown basaltic dikes, British Columbia (Symons, 1968).

In addition to composite pole 13, three other poles from interval 1B do not appear in the original references. Pole 14, from volcanic rocks of Mt. Edziza, British Columbia (Souther and Symons, 1973), was calculated from those flows with ages predominantly between 3 and 6 m.y. (sites A1 to A18; B1 to B49; and C1 to C32), excluding a few sites with $\alpha_{95} > 15°$. Pole 15 is the mean of 12 sites ($N \geq 5$, $\alpha_{95} \leq 15°$) in lavas from the Rio Grande Gorge, New Mexico (Kono et al., 1967). Pole 17 was computed from the mean direction reported by MacFadden (1977) from sediments of the Chamita Formation, New Mexico.

Six poles were given unit weight in determining a pole most representative of interval 1B. These include poles 13, 14, 15, and 17 (defined above), as well as pole 10 (from volcanic rocks near Virginia City, Montana; Hanna, 1967) and pole 16 (from the Sonoma Volcanics, California; Mankinen, 1972). No paleomagnetic poles from interval 1B were excluded from the determination of the interval pole.

Except for pole 10 (which represents only 5 flows), the poles given unit weight in interval 1B cluster around a mode at 85°N, 103°E (Figure 4.2). The mode for interval 1B is nearly identical to that for interval 1A and is similarly far-sided of the geographic pole by about 5°. This suggests that the far-sided tendency may have persisted since at least 7 m.y.B.P. (although many more poles would be required to test this hypothesis at the 95% confidence level).

Interval 1C

In Table 4.1 are listed 13 paleomagnetic investigations of rocks with ages between 7 and 21 m.y. (ca. late and middle Miocene). These studies are predominantly of intrusive and extrusive igneous rocks from western U.S. and Mexico. From these studies, 5 poles have been derived (Table 4.1, Figure 4.2), 4 of which do not appear in the original references.

Thirteen published poles for interval 1C were judged to have insufficiently averaged out secular variation. These 13 poles have been combined to form composite pole 18, with a mode at 83°N, 205°E. Four of these poles are from the study by Watkins et al. (1971) of the Rio Grande de Santiago Volcanics, Mexico (ca. 9 m.y.B.P. sites 1 to 4); three are from the Mount Barr pluton, British Columbia (sites 6, 8, and 10; Symons, 1973c); four are from the Snoqualmie Batholith, Washington (sites S1, S3, S4, and S6; Beske et al., 1973); one is from the Hiko ignimbrite, Nevada (Gose, 1970); and one is from the Beaverhead Valley (Horse Prairie Creek), Montana, igneous units studied by Hanna (1967).

In addition to composite pole 18, three other poles from interval 1C do not appear in the original references. Poles 19 and 20 represent results of intensive paleomagnetic investigations of basalts from the Columbia Plateau in the Pacific Northwest; these results have been summarized and discussed in detail by Watkins and Baksi (1974). Pole 19 has been derived from the Snake River, Lapwai, and Asotin poles of their Table 14; pole 20 is from the Imnaha, Owyhee, and Steens poles of that table. Pole 21 is the mean of the pole (Symons, 1969b) for basalts of the Interior Plateau in the Cariboo region, south-central British Columbia, and the nearly identical pole (Symons, 1969a) for nearby volcanic plugs thought to be coeval with the plateau basalts.

Four poles (18, 19, 21, and 22) were given unit weight in determining a pole most representative of interval 1C; pole 20 was excluded. Watkins and Baksi (1974) suggested that a pole near pole 19 represented the unrotated middle Miocene pole for North America, whereas basalts represented by pole 20 had experienced at least 15° of post-magnetization clockwise rotation. This interpretation might seem to be supported by the proximity of pole 19 to composite pole 18 (derived from rocks in Mexico, British Columbia, Washington, Montana, and Nevada) and to pole 21 (from the Interior Plateau of British Columbia)(Figure 4.2). It is curious, however, that poles 18, 19, and 21 differ significantly from the middle Miocene pole perhaps least likely to have experienced vertical-axis rotation: pole 22 (Young and Brennan, 1974) from the western Colorado Plateau, representing 15 distinct cooling units (the Peach Springs Tuff and 14 overlying and underlying flows) of two polarities (11 normal and 4 reversed). Clearly, many more paleomagnetic studies of middle Miocene rocks from cratonic North America are needed to determine whether pole 22 or the cluster formed by poles 18, 19, and 21 is more typical of the middle Miocene for North America. The mode of poles 18, 19, 21, and 22 (85°N, 220°E) is provisionally accepted as being the best estimate of the pole for interval 1C.

A mode determined for all 20 paleomagnetic poles given unit weight in intervals 1A through 1C is at 88°N, 74°E. Because 16 of these poles are from rocks less than 7 m.y. old, the mode for interval 1 is heavily weighted toward its youngest third.

Interval 2

In Table 4.1 are listed 22 paleomagnetic investigations of rocks with ages between 21 and 42 m.y. (ca. early Miocene, Oligocene, and late Eocene). These studies are predominantly of intrusive and extrusive igneous rocks and ignimbrites from western U.S. and Mexico. From these studies, 13 poles have been derived (Table 4.1, Figure 4.3), 3 of which do not appear in the original references.

Six published poles for interval 2 were judged to have insufficiently averaged out secular variation. These 6 poles have been combined to form composite pole 23, with a mode at 89°N, 159°E. Three of these poles are from the Grotto Batholith, central Washington, studied by Beske et al. (1973); two are from the non-transition-zone flows from Yarmony Mtn., Colorado (one, the average of normal-polarity poles 1 and 2 of York et al. (1971), and the other the average of their reversed poles 4 and 5); and one is from the Mistastin Lake volcanic rocks (Labrador), no longer considered Late Triassic as reported in Currie and Larochelle (1969), in light of the 38±4 m.y. age recently reported by Mak et al. (1976).

In addition to composite pole 18, three other poles from interval 2 do not appear in the original references. Pole 24, from ignimbrites and lavas near Durango, Mexico (Nairn et al., 1975a) is taken from the catalog of McElhinny and Cowley (1978) (their pole 15.027). Pole 28, from the Buck Hill Volcanic Series, Texas, is computed from the mean directions of magnetization reported in Gilliland et al. (1969). Pole 30, from the Marys Peak sill, Oregon,

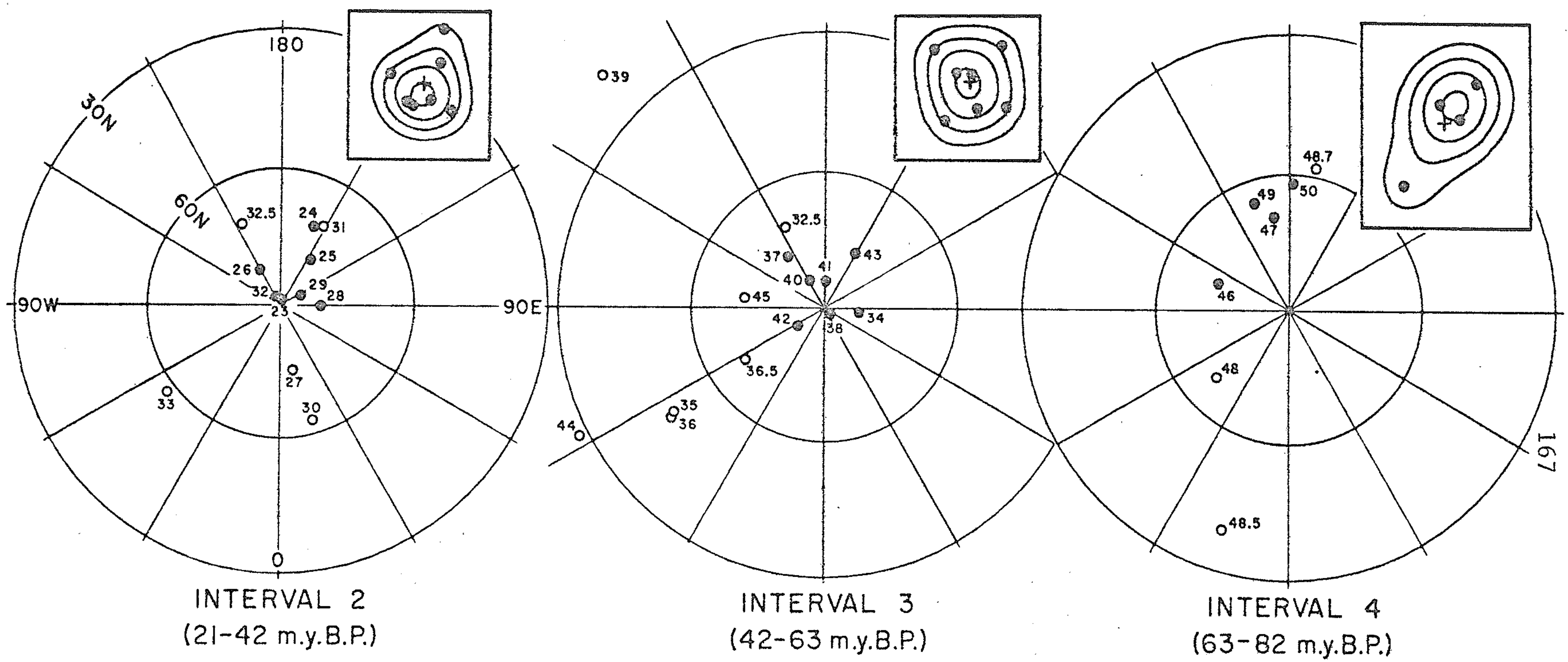


Figure 4.3 North American paleomagnetic poles for intervals 2, 3, and 4 (21 to 82 m.y.B.P.; approx. early Miocene to late Cretaceous). Symbols, notation, and contouring procedures are explained in the caption for Fig. 4.2.

is calculated from mean site directions reported in Clark (1969).

Seven poles were given unit weight in determining a pole most representative of interval 2. These include poles 23, 24, and 28 (defined above), as well as poles 25, 26, 29, and 32. Pole 25 is that reported by Grommé and McKee (1971) as the mean of 56 VGP's from flows and tuffs from Utah, Nevada, Oregon, and California; it is probably a more representative average than any that could be computed from the other, less extensive studies of this sequence of rocks listed in Table 4.1. Pole 26 is the mean of 6 sites from the Spanish Peak dikes, Colorado (Larson and Strangway, 1969), which are now considered to be 22 to 25 m.y. old (Stormer, 1972). Pole 29, recently determined by Beck et al. (1977), incorporates paleomagnetic results from all previous investigations of the San Juan volcanic field, Colorado. Pole 32, which is taken directly from the original reference (Symons, 1973c), is from the Hope plutonic complex, British Columbia.

Poles from 7 studies listed in Table 4.1 were excluded from the determination of the pole for interval 2. A pole (not shown in Figure 4.3) from the Lovejoy basalt, California, has an α_{95} of 25° and may reflect anomalous behavior of the geomagnetic field (Grommé, 1975). Pole 27 is taken from Table 3 of Strangway et al. (1971); however, the confidence limits for this pole and sampling details have apparently not yet been published. Pole 30, from Marys Peak sill, Oregon, probably represents less than 6000 years and may reflect a time of anomalous behavior of the geomagnetic field (Clark, 1969) or perhaps 30° of clockwise rotation of western Oregon (Simpson and Cox, 1977). Pole 31, from the East Sooke stock, Vancouver Island, British Columbia, was thought by Symons (1973d) to reflect a 20° counterclockwise rotation of the southern part of that island relative to cratonic North America. Similarly, site poles reported by Symons (1971c) from some minor Oligocene igneous bodies on other parts of Vancouver Island may well have had a complex tectonic history; moreover, the individual site poles probably have not sufficiently averaged out secular variation to be regarded as paleomagnetic poles. Pole 32.5, from the Younger Plutons, Queen Charlotte Islands, British Columbia, has an imprecise age (Paleocene to Oligocene); in addition, these plutons may have been rotated through large angles about a vertical axis (Hicken and Irving, 1977). Pole 33, from the Yachats basalt, Oregon, may reflect about 50° of clockwise rotation of western Oregon since the late Eocene (Simpson and Cox, 1977).

The mode determined for interval 2 (85°N, 138°E) differs by about 4° from the Oligocene reference pole (81°N, 132.5°E) determined by Beck _et al._ (1977). Their Oligocene pole was computed by giving unit weight to site poles incorporated in poles 25, 26, and 29 of this study.

Interval 3

In Table 4.1 are listed 19 paleomagnetic investigations of rocks with ages between 42 and 63 m.y. (ca. middle Eocene into early Paleocene). About 90% of these studies are of intrusive and extrusive igneous rocks from western U.S. and Alaska. From these studies, 13 poles have been derived (Table 4.1, Figure 4.3), 2 of which do not appear in the original references.

Six published poles for interval 3 were judged to have insufficiently averaged out secular variation. These 6 poles have been combined to form composite pole 34, with a mode at 82°N, 82°E. Three of these poles are from intrusives and extrusives of the Front Range, Colorado (North Table Mountain flows 1 and 2 and Lyons Quarry sill; Hoblitt and Larson, 1975). Also included in composite pole 34 is one pole from the North Table Mountain flow 1 of Larson et al. (1969); one pole from volcanic rocks in the Beaverhead River Valley, Montana (Hanna, 1967); and one pole from the Granite Falls stock, Washington (Beske et al., 1973).

The other interval 3 pole that does not appear in the original reference is pole 38, derived from 6 sites in felsite and basic intrusions from Virginia (Løvlie and Opdyke, 1974). Pole 38 represents the mean of these 6 site poles and was taken from the catalog of McElhinny and Cowley (1977) (their pole 14.146).

Seven poles were given unit weight in determining a pole most representative of interval 3. These include poles 34 and 38 (defined above), as well as pole 37 (from the Quottoon and Kasiks plutons, British Columbia; Symons, 1974a), pole 40 (from the Rattlesnake Hills intrusives, Wyoming; Shive et al., 1977), pole 41 (from the Absaroka Volcanic Supergroup, Wyoming; Shive and Pruss, 1977), pole 42 (from lavas of Baffin Island, Canada; Deutsch et al., 1971), and pole 43 (from sedimentary rocks of the Nacimiento Formation, New Mexico;

Butler and Taylor, 1978).

Several poles from interval 3 were excluded from the determination of the mode. Five of these are paleomagnetic poles that probably reflect significant vertical-axis rotations (for reasons cited in the original references): pole 35 (from marine sediments of the Tyee and Flournoy Formations, Oregon; Simpson and Cox, 1977), pole 36 (from the Siletz River Volcanic Series, Oregon; Simpson and Cox, 1977), pole 36.5 (from volcanic rocks from the Tolstoi Formation, Alaska; Stone and Packer, 1977), pole 44 (from dunite of the Twin Sisters Intrusion, Washington; Beck, 1975), and pole 45 (from volcanic rocks of the Masset Formation, British Columbia; Hicken and Irving, 1977). In addition, pole 32.5 (from the Younger Plutons, British Columbia; Hicken and Irving, 1977) was excluded not only because of the possibility of a vertical-axis rotation but also because of its imprecise age (Paleocene to Oligocene). Eocene site poles from the Idaho Batholith, Idaho (Beck et al., 1972) and a pole from intrusive rocks on Flagstaff Mountain, Colorado (McMahon and Strangway, 1968a,b) are probably not true paleomagnetic poles and may well have experienced post-magnetization vertical-axis rotations. A pole from annually-banded shales of the Green River Formation, Colorado, was thought by Strangway and McMahon (1973) to reflect an anomalous configuration of the geomagnetic field during about 10,000 years of Eocene time.

The seven poles from interval 3 given unit weight yield a mode at 86°N, 195°E. The modes determined for intervals 1, 2, and 3 are all within 5° of the present geographic pole. This suggests that little apparent polar wandering occurred with respect to North America during most of the Cenozoic.

Mesozoic

Interval 4

In Table 4.1 are listed 9 paleomagnetic investigations of rocks with ages between 63 and 82 m.y. (ca. early Paleocene and late Cretaceous). Of these, 5 studies are of igneous rocks from northwestern U.S. and British Columbia, and 4 studies are of sedimentary rocks from western U.S., Alaska, and Mexico. From these studies, 7 poles have been derived (Table 4.1, Figure 4.3), all of which are taken directly from the original references.

Four poles were given unit weight in determining a pole most representative of interval 4. These include pole 46 (from the Boulder Batholith, Montana; Hanna, 1973a), pole 47 (from the Elkhorn Mountains volcanic field, Montana; Hanna, 1967), pole 49 (from sedimentary rocks of the Mesaverde Group, Wyoming and Utah; Kilbourne, 1969), and pole 50 (from sedimentary rocks of the Hilliard Formation near Kemmerer, Wyoming; Shive and Frerichs, 1974). Judging from Figure 2 of Shive and Frerichs (1974), pole 50 is based on rocks that are predominantly younger than Santonian (<82 m.y.B.P.).

Several poles from interval 4 were excluded from the determination of the mode. Pole 48 (from the Ecstall pluton, British Columbia; Symons, 1974a) and pole 48.5 (from sedimentary rocks of the Chignik Formation, Alaska; Stone and Packer, 1977) are thought to reflect post-magnetization tectonic rotations through large angles (for reasons discussed in the original references). Pole 48.7 (from red beds of the Difunta Group, northern Mexico; Nairn, 1976) was excluded because it is not yet certain to what extent Mexico can be regarded as part of cratonic North America in the Mesozoic. Poles from one site in the Idaho batholith, Idaho (Beck _et al._, 1972) and from two sites in gabbroic intrusives from British Columbia (Symons, 1968) were excluded because they probably do not represent enough time to have averaged out secular variation, and because they may well reflect post-magnetization tectonic rotations. A pole determined by Shive and Frerichs (1974) from the Campanian part of the Niobrara Formation (their section in the Centennial Valley, Wyoming) was excluded because it is based on only 8 samples with scattered directions of magnetization.

The four paleomagnetic poles given unit weight have a mode at 66°N, 191°E. This mode for interval 4 is displaced by about 20° from the mode for interval 3. Curiously, there appears to be no well-defined streak between the modes for the two intervals. In fact, connecting pole 46 from the Boulder Batholith (probably the youngest pole of interval 4) with pole 43 from the Nacimiento Formation (probably the oldest pole of interval 3) would produce a highly irregular APW path. A conservative interpretation of the data presently available from these two intervals would suggest an episode of rapid (nearly 1°/m.y.) apparent polar wander with respect to North America between about 75 m.y.B.P. (poles 50, 49, and 47) and 50 m.y.B.P. (poles 41, 40, and 38).

Interval 5

In Table 4.1 are listed 6 paleomagnetic investigations of rocks with ages between 82 and 103 m.y. (parts of late and middle Cretaceous). Of these, 5 studies are of intrusive igneous rocks from western and central U.S. and British Columbia, and 1 study is of sedimentary rocks of western and central U.S. From these studies, 5 poles have been derived (Table 4.1, Figure 4.4), all of which are taken directly from the original references.

Three poles were given unit weight in determining a pole most representative of interval 5. These include pole 52 (from plutons in the Sierra Nevada, California; Currie et al., 1963; Grommé and Merrill, 1965), pole 55 (from alkalic complexes, Arkansas; Scharon and Hsu, 1969), and pole 51 (from sedimentary rocks of the Niobrara Formation near Pueblo, Colorado; Shive and Frerichs, 1974). Judging from Figure 2 of Shive and Frerichs (1974), pole 51 is based on rocks that are predominantly older than Campanian (>82 m.y.B.P.) (from the Fort Hays Member and lower Smoky Hill Member).

Several poles from interval 5 were excluded from the determination of the mode. Pole 53 (from granites near Stevens Pass, Washington; Beck and Noson, 1972) and pole 54 (from felsic plutons along Howe Sound, British Columbia; Symons, 1973b) were excluded because of the strong possibility of post-magnetization tectonic rotations of these rocks. A pole determined by Shive and Frerichs (1974) from the Coniacian part of the Niobrara Formation (the Fort Hays Member sampled near Trego, Kansas) was excluded because it is based on only 5 stably magnetized samples (although this pole is nearly identical to pole 51 from the Niobrara near Pueblo, Colorado).

The three paleomagnetic poles given unit weight are very tightly grouped (Figure 4.4) about a mode at 66°N, 187°E. This mode for interval 5 is about 3° from the mode determined for interval 4.

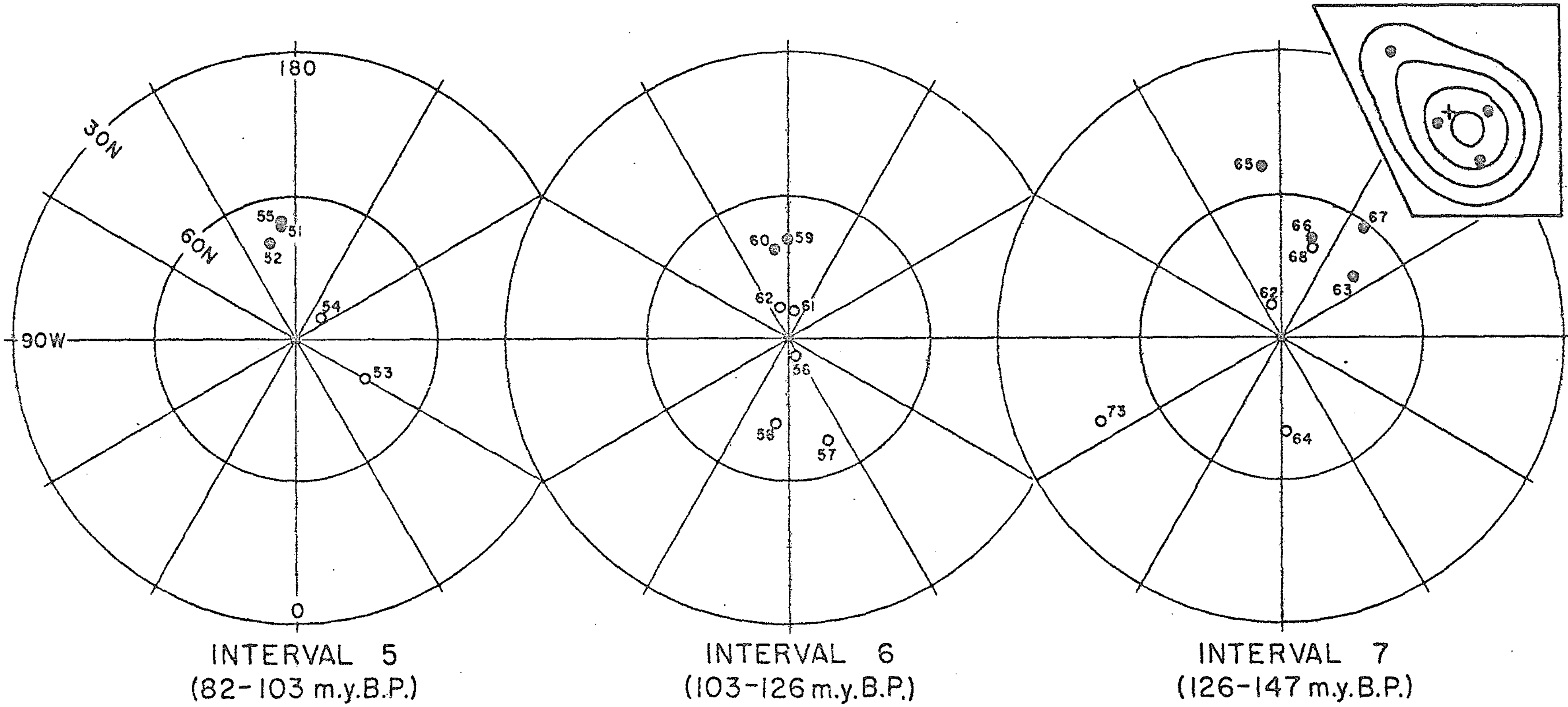


Figure 4.4 North American paleomagnetic poles for intervals 5, 6, and 7 (82 to 147 m.y.B.P.; approx. late Cretaceous to late Jurassic). Symbols, notation, and contouring procedures are explained in the caption for Fig. 4.2. Contours for interval 5 were too peaked to show at the scale of this figure.

Interval 6

In Table 4.1 are listed 11 paleomagnetic investigations of rocks with ages between 103 and 126 m.y. (parts of middle and early Cretaceous). All of these studies are of intrusive igneous rocks, distributed over most of Canada and northeastern and western U.S. From these studies, 7 poles have been derived (Table 4.1, Figure 4.4), only one of which (pole 61) is not taken directly from the original references.

Pole 61 is a composite paleomagnetic pole, based on three paleomagnetic studies of the White Mountain magma series, New England: Opdyke and Wensink (1966), Bouley (1971), and de Boer (cited in Bouley, 1971). Age assignments for individual site poles listed in these studies have been reinterpreted here on the basis of extensive isotopic age determinations on the magma series by Foland et al. (1971) and Foland and Faul (1977). Only site poles based on at least 5 stably-magnetized samples and with an α_{95} less than 15° are included in pole 61. Ten site poles, representing both normal and reversed polarities, meet these criteria: 7 from Opdyke and Wensink (1966), 2 from data of Bouley (1971), and 1 from data of de Boer (in Bouley, 1971). Pole 61 (84°N, 168°E) is the mode of these 10 site poles.

Two poles were given unit weight in determining a pole most representative of interval 6: pole 59 (from diabase near Isachsen in the Canadian Arctic; Larochelle and Black, 1963) and pole 60 (from intrusives of the Monteregian Hills, Quebec; Larochelle, 1968, 1969).

Other poles from interval 6 were considered to be unrepresentative of the position of the paleogeographic pole with respect to the North

American craton. Pole 56 (from the Southern California Batholith; Teissere and Beck, 1973) and poles 57 and 58 (from the Stephens Island and Captain Cove plutons, westernmost British Columbia; Symons, 1977) are thought to reflect post-magnetization tectonic rotations (for reasons cited in the original references). Pole 61 (the composite pole derived from the White Mountain magma series, New England) and pole 62 (from 12 lamprophyre dikes in Newfoundland; Deutsch and Rao, 1977b) are from structurally simpler settings; hence, their discordance from poles 59 and 60 of interval 6 and from modes determined for intervals 5 and 4 probably requires some explanation other than post-magnetization tectonic rotation. Pole 61 is suspect in light of the failure of older site poles from the White Mountains to agree with supposedly coeval poles from the craton (cf. discussions of data from intervals 8 and 9). The age of pole 62 is not well known and could be as old as 144 m.y. (Deutsch and Rao, 1977b). Poles from one site in the Idaho batholith, Idaho (Beck et al., 1972) and from two sites in the Fraser unit of the Topley Intrusions, central British Columbia (Symons, 1973a) probably do not represent enough time to have averaged out secular variation. Moreover, the Idaho batholith may have experienced post-magnetization internal deformation (Beck et al., 1972).

The best estimate of the pole for interval 6 is at 70°N, 184°E (i.e., the mean of poles 59 and 60). The poles determined for intervals 4, 5, and 6 are all within about 5° of one another. This suggests that the paleogeographic pole was relatively stationary ("quasi-static"; Briden, 1967) with respect to North America during most of Cretaceous time.

Interval 7

In Table 4.1 are listed 8 paleomagnetic investigations of rocks with ages between 126 and 147 m.y. (early Cretaceous and late Jurassic). About half these studies are of intrusive igneous rocks from California, British Columbia, and Newfoundland; the other half are of sedimentary rocks (predominantly red beds) from Colorado, southeastern Mexico, and the Alaska Peninsula. From these studies, 8 poles have been derived (Table 4.1, Figure 4.4), all of which are taken directly from the original references.

Four poles were given unit weight in determining a pole most representative of interval 7. These include pole 63 (from 13 sites representing 6 of the Topley Intrusions, British Columbia; Symons, 1973a), pole 65 (from 13 sites in the Guadalupe and Bucks intrusions in the Sierra Nevada, California; Grommé et al., 1967), and poles 66 and 67 (from red beds of the upper and lower Morrison Formation, Colorado; Steiner and Helsley, 1975).

Four poles from interval 7 were excluded from the determination of the mode. Pole 62 (from lamprophyre dikes in Newfoundland; Deutsch and Rao, 1977b) has an uncertain age and could be as young as 115 m.y. Pole 64 (from the Gil Island pluton, westernmost British Columbia; Symons, 1977) is thought to reflect a post-magnetization tectonic rotation (for reasons cited in the original reference). Pole 68 (from red beds of the Todos Santos Formation, southeastern Mexico; Guerrero and Helsley, 1974) is nearly identical to pole 66 from the upper Morrison Formation; however, pole 68 was excluded from the mode because it is not yet certain to what extent southern Mexico can be regarded as part of cratonic North America in the Mesozoic. Pole 73 (from sedimentary rocks of the Alaska Peninsula; Packer and Stone, 1974) was excluded in light of paleomagnetic evidence (Stone and Packer, 1977) that southern Alaska has moved north and rotated clockwise with respect to cratonic North America during Mesozoic and/or early Tertiary time.

The 4 paleomagnetic poles given unit weight have a mode at 66°N, 145°E. This pole for interval 7 is displaced about 15° from the pole for interval 6. Moreover, Steiner and Helsley (1975) noted significant apparent polar motion during Morrison time. These observations suggest that the Late Jurassic-Early Cretaceous was a time of comparatively rapid apparent polar wander with respect to North America (about 0.7°/m.y.). Evidently, the Cretaceous quasi-static interval did not begin until about 120 m.y.B.P. (earliest Monteregian Hills time).

Interval 8

In Table 4.1 are listed 11 paleomagnetic investigations of rocks with ages between 147 and 169 m.y. (late to middle Jurassic). About 75% of these studies are of intrusive igneous rocks from British Columbia, western U.S., and New England; the other 25% are of sedimentary rocks from Utah and the Alaska Peninsula. From these studies, 8 poles have been derived (Table 4.1, Figure 4.5), only one of which (pole 76) is not taken directly from the original references.

Pole 76 is a composite paleomagnetic pole, based on paleomagnetic studies of the White Mountain magma series, New England, by Opdyke and Wensink (1966), Bouley (1971), and de Boer (cited in Bouley, 1971). As in interval 6, age assignments for individual site poles listed in

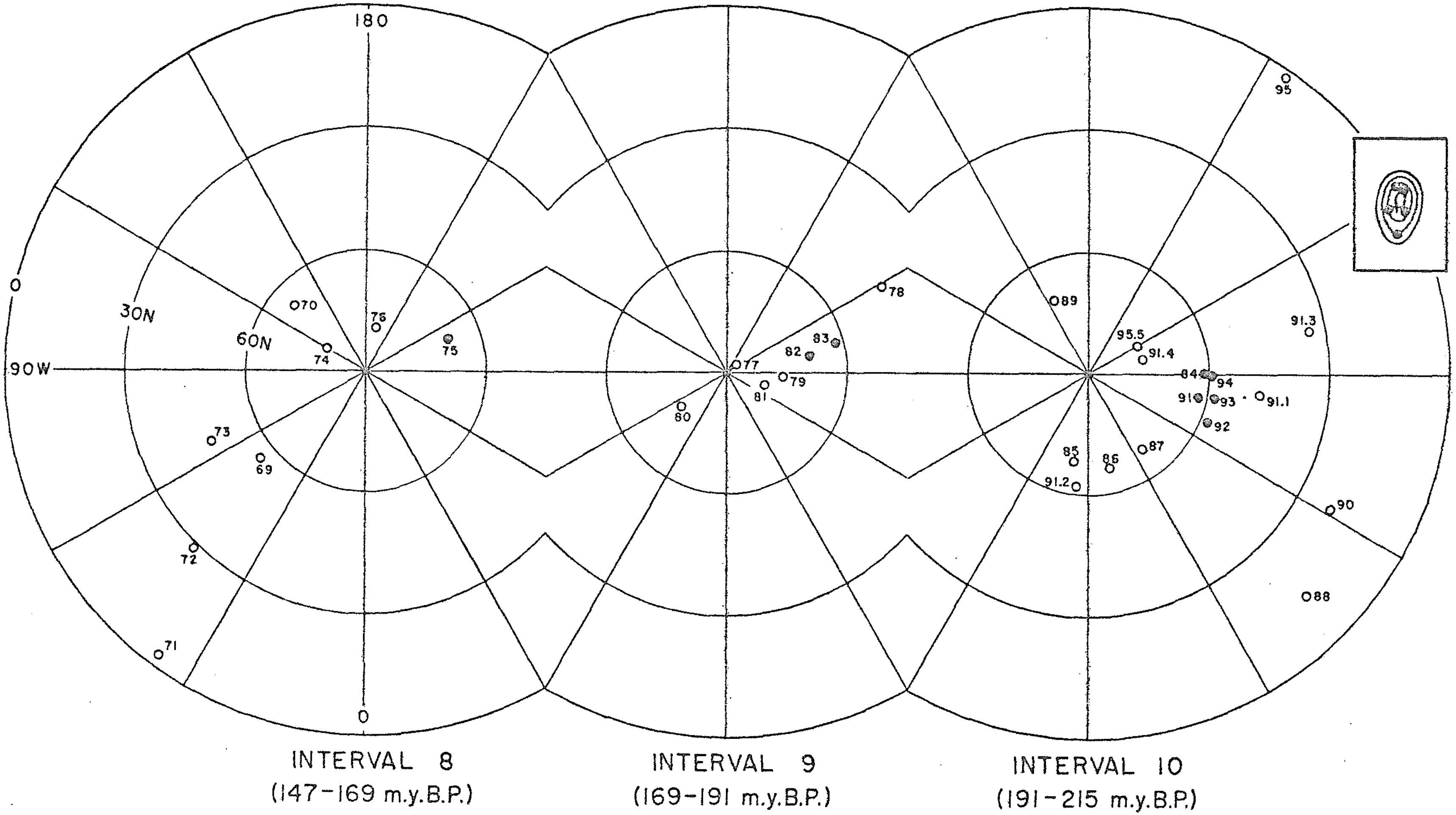


Figure 4.5 North American paleomagnetic poles for intervals 8, 9, and 10 (147 to 215 m.y.B.P.; approx. late Jurassic to middle Triassic). Symbols, notation, and contouring procedures are explained in the caption for Fig. 4.2.

these studies have been reinterpreted here on the basis of extensive isotopic age determinations on the magma series by Foland et al. (1971) and Foland and Faul (1977). Only site poles based on at least 5 stably-magnetized samples and with an α_{95} less than 15° are included in pole 76. Eight site poles, representing only normal-polarity directions, meet these criteria: 7 are from different intrusive phases of the Belknap stock, and 1 is from the Gore Mountain stock. Pole 76 (79°N, 167°E) is the mode of these 8 site poles.

Of all the poles for interval 8, only pole 76 from the White Mountain magma series and pole 75 from the Summerville Formation, Utah (Steiner, 1978) are from regions with a reasonably well-known and uncomplicated tectonic history. Of these two poles, that from the Summerville is probably the more reliable indicator of the position of the paleogeographic pole with respect to North America during at least part of interval 8 time. The Summerville section sampled by Steiner (1978) yielded almost perfectly antipolar normal and reversed-polarity directions; this pole is nearly identical to an earlier pole obtained from the Summerville at another locality in Utah (Steiner and Helsley, 1972). In contrast, the White Mountain pole 76 represents only normal-polarity directions. Moreover, the White Mountain data are suspect, because modes computed separately from sites with ages falling in intervals 9, 8, and 6 are very similar (consistently near the position of the present geographic pole), despite the minimum of 25° of apparent polar wandering that seems to have occurred between Early Jurassic and Middle Cretaceous time (cf. discussions of data from intervals 9 and 6). It is certainly possible that the discordance of the White Mountain and

Summerville poles resulted from their averaging different parts of interval 8 time. If this explanation were correct, then interval 8 must have been characterized by very rapid apparent polar wandering.

Interval 8 contains an abundance of apparently anomalous poles, which diverge by up to 90° from the White Mountain and Summerville poles (Figure 4.5). The discordance of most of these poles is perhaps not surprising in view of the probably complex tectonic histories of the rocks from which they are derived. Poles 69, 70, and 71 are from different phases of the Red Mountain ultramafic intrusion (Saad, 1969), and pole 72 is from igneous rocks of the Franciscan Formation (Grommé and Gluskoter, 1965); all four of these poles are from west-central California. Pole 73 is from sedimentary rocks of the Naknek and Chinitna Formations, southern and southeastern Alaska (Packer and Stone, 1974; Stone and Packer, 1977). Pole 74 is from the Island Intrusions, Vancouver Island, British Columbia (Symons, 1970). Beck (1976) has reviewed these and other Mesozoic and Tertiary anomalous poles from the westernmost Cordillera on a case-by-case basis; he concluded that the anomalous poles reflect regional right-lateral shear, since their distribution can be explained by clockwise rotation of declinations and flattening of inclinations. More recently, Stone and Packer (1977) have demonstrated the anomalous nature of not only Jurassic poles from southern Alaska, but Cretaceous (pole 48.5) and Eocene (pole 36.5) ones as well. Thus, regional tectonic rotations in parts of the western Cordillera provide plausible explanations for most of the anomalous poles of interval 8. However, until more Middle and Late Jurassic paleomagnetic poles are available from cratonic North America, suggestions by Grommé and Gluskoter (1965), Spall (1968), and Saad (1969) of an anomalous configuration of the geomagnetic field during parts of the middle Mesozoic cannot be totally dismissed (cf. discussions of anomalous results from intervals 11 and 24-25).

In view of the discrepancy between the White Mountain and Summerville poles, two possible interpretations of the middle Jurassic segment of the North American APW path are shown in Figure 4.12. A minimum of apparent polar wandering (the author's preferred interpretation, for reasons discussed above) would be required if the White Mountain pole were anomalous, and only the Summerville pole represented a valid paleomagnetic pole for interval 8. If the White Mountain pole were also a valid paleomagnetic pole, then significant apparent polar wandering would be indicated during interval 8, at least as much as that shown by the dashed segment connecting intervals 9 and 7 in Figure 4.12. (The point labeled 8' on this dashed segment is the mean of White Mountain pole 76 and Summerville pole 75.) Additional middle to late Jurassic paleomagnetic poles from cratonic North America are badly needed to ascertain which of these two poles is more representative of interval 8 time.

Interval 9

In Table 4.1 are listed 13 paleomagnetic investigations of rocks with ages between 169 and 191 m.y. (parts of middle and early Jurassic). These studies are predominantly of intrusive igneous rocks from eastern U.S. and eastern Canada, but they also include three studies of red beds from Utah. From these studies, 7 poles have been derived (Table 4.1,

Figure 4.5), only one of which (pole 77) is not taken directly from the original references.

Pole 77 is a composite paleomagnetic pole, based on paleomagnetic studies of the White Mountain magma series, New England, by Opdyke and Wensink (1966), Bouley (1971), and de Boer (cited in Bouley, 1971). As in intervals 6 and 8, age assignments for individual site poles listed in these studies have been reinterpreted here on the basis of extensive isotopic age determinations on the magma series (Foland et al., 1971; Foland and Faul, 1977). Only site poles based on at least 5 stably-magnetized samples and with an $\alpha_{95} < 15°$ are included in pole 77. Eight site poles, representing both normal and reversed polarities, meet these criteria (6 from the White Mountain batholith and 2 from the Mt. Monadnock complex). Pole 77 (87°N, 128°E) is the mode of these 8 site poles.

A pole most representative of interval 9 is probably that which can be derived from 139 VGP's from diabase dikes and sills in Connecticut, Pennsylvania, and Maryland (Beck, 1972; Smith, 1976a). These rocks intrude the youngest sedimentary rocks (or structures involving the youngest rocks) in the Hartford and Gettysburg Basins of the Newark Group (Smith, 1976a). The Newark Group contains intercalated dikes and sills, dated at 193±6 m.y. Although the dikes studied by Smith (1976a) must be younger than the Newark sedimentary rocks, which they cut, isotopically determined ages for the dikes range from 197 to 244 m.y. Armstrong and Besancon (1970) noted this discrepancy and considered all Newark Group ages suspect, attributing ages less than 200 m.y. to Ar loss during zeolite facies metamorphism. However, a younger age for the Newark is indicated by the recent discovery of Early to early Middle

Jurassic pollen assemblages near the top of the Group (Cornet _et al._, 1973; Cornet and Traverse, 1975). An age less than 200 m.y. for the dikes studied by Smith (1976a) can also be supported on other grounds. These intrusives are the North American part of a dike swarm radiating from a point near the present Blake plateau-Bahama platform. This dike swarm was thought by May (1971) to reflect the stress field at the initiation of the break-up of Pangea. Recently, Dalrymple _et al._ (1975) have reported isotopic ages and paleomagnetic poles from the African part of the dike swarm in Liberia. The Liberian dikes yield ages of 180 ±10 m.y. and a paleomagnetic pole nearly identical to that from their North American counterparts, after closure of the North Atlantic Ocean (cf. Dalrymple _et al._, 1975; Smith, 1976a). Thus, the average of VGP's from the studies by Smith (1976a) and Beck (1972) probably provides a reliable point on the North American APW path representative of the interval 180 ±10 m.y.B.P. (cf. Smith, 1976b).

Other poles from interval 9 are probably less representative of the paleomagnetic pole with respect to North America in the Early Jurassic. De Boer (1967) has also reported poles from Appalachian dikes cross-cutting the Newark Group. Although several of the units studied by de Boer (1967) were also studied by Smith (1976a), the mean poles from the two investigations differ by about 20°. This difference has been attributed by de Boer (cited in Smith, 1976a) to excess alternating-field demagnetization of de Boer's samples. Composite pole 77, determined from the White Mountain magma series, is suspect because of the consistent position (near the present geographic pole) of White Mountain poles from rocks spanning a very long time interval (from about 185 to

115 m.y.B.P.; Early Jurassic to Early Cretaceous). Pole 78 (from the Tulameen complex, British Columbia; Symons, 1974b) is thought to reflect post-magnetization tectonic rotation (for reasons cited in the original reference). Pole 79, from a single dike on Anticosti Island, Quebec, probably represents only a spot reading of the magnetic field and might even reflect a post-magnetization vertical-axis rotation of that island (Larochelle, 1971). Poles 80 and 81, from the Carmel Formation and Navajo Sandstone in Utah, are close to the present geographic pole and were considered by Johnson (1976) to reflect Brunhes-age remagnetization, partly on the basis of a fold test in the Navajo. A pole was determined by Larochelle and Wanless (1966) for a single dike near Shelburne, Nova Scotia; however, this dike may not represent enough time to have averaged out secular variation and has an imprecise age, although it is generally thought to be related to the intrusives crosscutting the Newark Group (e.g., May, 1971; Beck, 1972).

The pole determined for interval 9 (65°N, 103°E) is the mean of pole 82 (68.6°N, 100.9°E; Smith, 1976a) and pole 83 (62.0°N, 104.5°E; Beck, 1972), giving unit weight to each. (Note that the interval 9 pole computed here differs by about 1° from a composite pole (65.4°N, 101.6°E) computed by Smith (1976a), which he claimed to have derived by giving unit weight to each of the 139 VGP's determined in the two studies; however, Smith's composite pole cannot be correct, since it should be weighted slightly toward the study of Beck (1972), from which 78 of the 139 VGP's were taken.) The proximity of the pole determined for interval 9 to pole 75 from the Summerville Formation suggests that only a few degrees of apparent polar wander occurred with respect to

North America during most of early and middle Jurassic time (if the minimum APW interpretation of the interval 8 results is correct).

Interval 10

In Table 4.1 are listed 23 paleomagnetic investigations of rocks with ages between 191 and 215 m.y. (ca. earliest Jurassic and late Triassic). About half these studies are of intrusive and extrusive igneous rocks from northeastern U.S., eastern and western Canada, and Alaska. The other half are of red beds from eastern and western U.S. and Mexico. From these studies, 17 poles have been derived (Table 4.1, Figure 4.5), two of which (poles 84 and 94) are not taken directly from the original references.

In terms of sheer number, poles from igneous rocks associated with the Newark Group of the northeastern U.S. dominate most published compilations of Late Triassic poles for North America. It must be emphasized, however, that most of these poles are VGP's from a few flows or dikes that individually have probably not averaged out secular variation. Consequently, in this study, unit weight is given to a composite Newark pole based on the mode of the following 13 poles: 2 poles from the North Mountain basalt, Nova Scotia (Carmichael and Palmer, 1968; Bouley, 1969; Larochelle, 1967b), one representing the upper and one the lower flows; 2 poles from Irving and Banks (1961); 3 poles from Opdyke (1961); and 6 poles from de Boer (1968). These 13 poles yield a mode at 61°N, 90°E (pole 84) (cf. the Newark mean of Smith (1976b) at 64°N, 84°E).

The other pole from interval 10 that is not taken directly from the original references is pole 94 (59°N, 89°E), the mean of two poles (which differ by only 4°) from investigations of igneous rocks from the Manicouagan structure, Quebec (Larochelle and Currie, 1967; Robertson, 1967). Monzonite of the Manicouagan structure has been dated as 210±4 m.y. by Wolfe (1971).

Five poles were given unit weight in determining a pole most representative of interval 10. These include poles 84 and 94 (defined above), as well as pole 91 (from the Kayenta Formation, Utah; Steiner and Helsley, 1974a), pole 92 (from the Wingate and Chinle Formations, Utah; Reeve, 1975), and pole 93 (from the Chinle Formation, New Mexico; Reeve and Helsley, 1972). (A pole for the Kayenta determined by Johnson (1972, 1976) is based on only 7 samples; this pole should probably be regarded as having been superseded by the pole of Steiner and Helsley (1974a), which is based on directions from 105 samples representing 7 polarity zones.) Poles from each of the two, widely-spaced Chinle localities were given unit weight because they probably represent a different part of Late Triassic time, as indicated by different observed polarity structure in each section sampled.

Other poles from interval 10 are probably less representative of the paleomagnetic pole with respect to North America in the Late Triassic. Many of these poles are from rocks of the western Cordillera and are discordant from poles from elsewhere in North America. These anomalous poles include pole 85 (from the Copper Mountain intrusions, British Columbia; Symons, 1973e), pole 86 (from the Guichon batholith, British Columbia; Symons, 1971b), and poles 87 through 90 (from the Karmutsen Volcanic Group, Vancouver Island, British Columbia; Symons, 1971a; Irving and Yole, 1972). Beck (1976) has reviewed each of these and other Mesozoic and Tertiary anomalous poles from the westernmost Cordillera on a case-by-case basis; he concluded that the distribution of these poles was suggestive of a history of right-lateral shear throughout much of this region (cf. discussion of anomalous poles for interval 8). Similarly, anomalous pole 95 (from the Nikolai Greenstone, south-central Alaska) was considered by Hillhouse (1977) to reflect a complex tectonic history, a hypothesis supported by other paleomagnetic data from this region (e.g., Stone and Packer, 1977). Pole 95.5 from red beds of the Nazas Formation, west-central Mexico (Nairn, 1976) was not included in determining the mode for interval 10, partly because of its imprecise age (Middle Triassic to Early Jurassic), and partly because it is not yet certain to what extent Mexico can be regarded as part of cratonic North America in the Mesozoic.

Although post-magnetization tectonic rotations provide the most plausible explanations for the discordance of many of the poles of interval 10, some anomalous poles have been reported for rocks from tectonically stable regions. Steiner and Helsley (1972, 1974) found anomalous magnetizations (represented by poles 91.1 through 91.4 of Table 4.1) in the Kayenta Formation of the Colorado Plateau. Three of these anomalous poles were reproducible in sections 6.5 km apart, leading them to suggest that these magnetizations could have recorded an anomalous configuration of the Late Triassic geomagnetic field.

The mode of the 5 poles given unit weight for interval 10 is at 60°N, 83°E. This mode is about 11° southwest of the pole determined for interval 9. Together, the poles for intervals 10, 9, and 8 (minimum APW interpretation) indicate an APW rate of about 0.3°/m.y. with respect to North America from late Triassic through middle Jurassic time.

Interval 11

In Table 4.1 are listed 11 paleomagnetic investigations of rocks with ages between 215 and 239 m.y. (ca. middle and early Triassic). These studies are predominantly of red beds from Colorado, Utah, and Wyoming. From these studies, 9 poles have been derived (Table 4.1, Figure 4.6), only one of which (pole 98) is not taken directly from the original references. Pole 98 (57°N, 99°E) is the mean of two poles (which differ by about 1°) determined for the lower Moenkopi Formation, Colorado, by Helsley (1969) and by Baag and Helsley (1974).

In computing a mode for interval 11, unit weight was given to well-grouped poles 97 (from the upper Moenkopi Formation, Colorado; Helsley and Steiner, 1974), 98 (from the lower Moenkopi, as described above), and 99 (from the State Bridge Formation, Colorado; Christensen and Helsley, 1974a). Paleomagnetic results from these formations document at least 10 polarity zones probably spanning most of Early Triassic time.

The concordance of poles 97, 98, and 99 (Figure 4.6) hides a curious phenomenon first noted by Helsley and Steiner (1974) in the upper Moenkopi Formation: an "undulatory" variation of magnetization directions as a function of stratigraphic position. These directional variations, which are primarily declination swings, seem to be independent of geomagnetic reversals (cf. Figure 6 of Helsley and Steiner, 1974). Moreover, the directional variations correspond to apparent

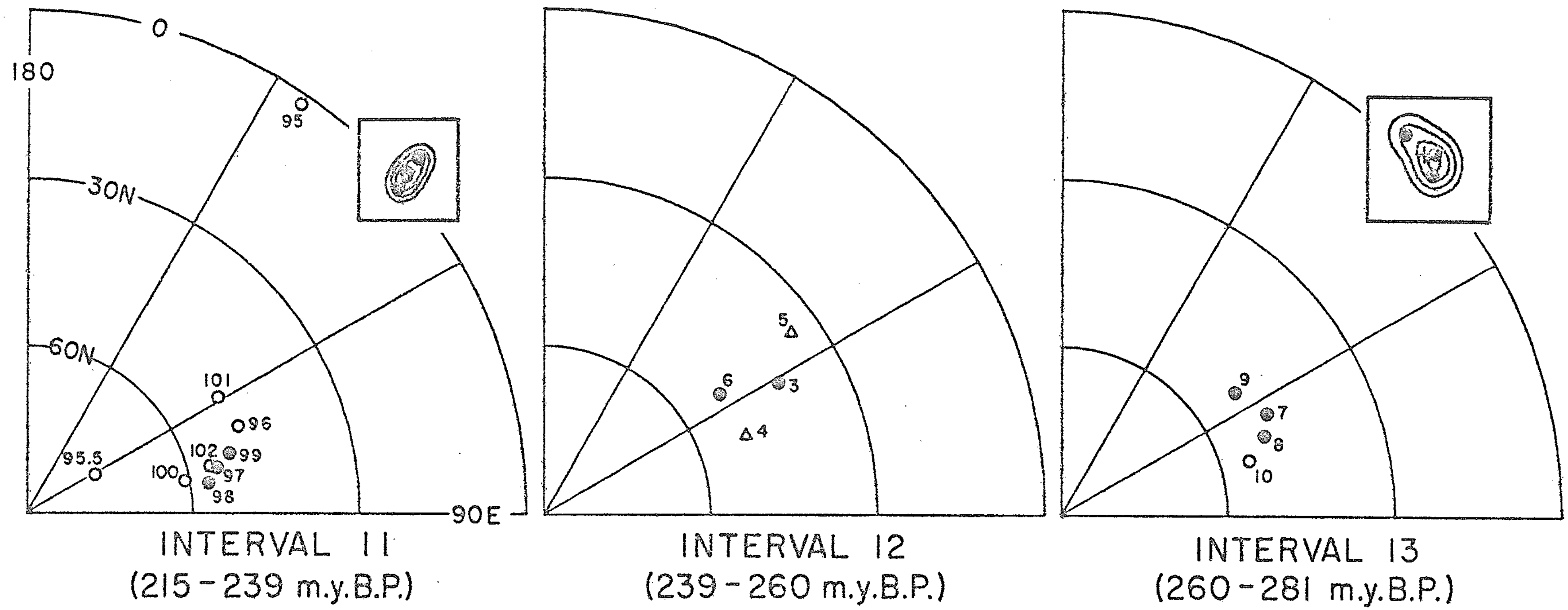


Figure 4.6 North American paleomagnetic poles for intervals 11, 12, and 13 (215 to 281 m.y.B.P.; approx. middle Triassic to latest Pennsylvanian). Symbols, notation, and contouring procedures are explained in the caption for Fig. 4.2. Poles represented by open triangles are from the "Maritime block" (Scott, 1975).

polar oscillations up to 30°, with estimated periodicities of between 10^5 and 10^6 years (Helsley, 1977). The undulations were thought by Helsley and Steiner (1974) and Helsley (1977) to record real features of geomagnetic field behavior, especially since the undulations were reproducible at widely-spaced sections: two in the Moenkopi Formation (Helsley, 1969; Helsley and Steiner, 1974) and one about 200 km to the northeast in the State Bridge Formation (Christensen and Helsley, 1974a).

Recognition of possible long-term directional variations in the geomagnetic field during the Early Triassic may be important in interpreting pole 96, from the Chugwater Formation, Wyoming (Van der Voo and Grubbs, 1977). The mean pole from the Chugwater (pole 96) differs by about 10° from the mode defined by the Moenkopi and State Bridge poles. Even more important, however, is the observation that site poles from the Chugwater, when plotted sequentially as a function of stratigraphic position, move progressively from a position near poles typical of the Permian (intervals 12 and 13) toward the mean poles from the Moenkopi and State Bridge Formations. Steiner (in Van der Voo and Grubbs, 1977) suggested that the linear distribution of site poles from the Chugwater (which represent only two polarity zones) might record a part of the undulatory directional variation noted in the Moenkopi and State Bridge Formations. Alternatively, the discordance of the Chugwater pole from the Moenkopi and State Bridge poles might be attributable to a slight vertical-axis rotation of the sample locality or to a possible pre-Moenkopi age for the Chugwater, as discussed by Van der Voo and Grubbs (1977). In view of these complexities, the Chugwater pole 96 was excluded from the interval 11 pole determination.

Several other poles from interval 11 were also excluded from the determination of the mode. Pole 100 (representing post-Kiaman directions from the Maroon Formation sampled near Horn Ranch, Colorado; McMahon and Strangway, 1968a,b) was excluded because stability of the magnetization of these red beds was tested only by AF demagnetization in peak fields of 500 Oe. The near antiparallelism (161° difference) between normal and reversed directions from the Horn Ranch locality, and the proximity of the resultant pole 100 to those from the Moenkopi and State Bridge Formations (Figure 4.6) suggest that the upper Maroon pole 100 may not grossly misrepresent the Early Triassic pole. However, other mixed-polarity results from the investigations of Colorado red beds by McMahon and Strangway (1968a,b) show differences of 144° (Red Sandstone Creek locality) and 130° (Squaw Creek locality) between normal and reversed-polarity directions. Thus, most of the poles of McMahon and Strangway (1968a,b) are probably unreliable because of inadequate cleaning, and they are not considered further in this study. Pole 101 (from the Hoskinnini Member, basal Moenkopi Formation; Farrell and May, 1969) was excluded because it is derived from only 18 samples from a single site and because a more complete sampling of the Hoskinnini by Helsley and Scott (reported in Scott, 1975) yields a preliminary pole (102) more typical of the reliable poles 97, 98, and 99 from interval 11. Pole 95 (from the Nikolai Greenstone, south-central Alaska; Hillhouse, 1977) and pole 95.5 (from the Nazas Formation, west-central Mexico; Nairn, 1976) were excluded because of their imprecise ages and the possibility that they reflect post-magnetization tectonic rotations, as discussed in interval 10.

The mode of the 3 poles given unit weight for interval 11 is at 55°N, 103°E. This mode is about 12° southeast of the pole determined for interval 10, reflecting an abrupt change in the direction of apparent polar wandering with respect to North America during the late Triassic or early Jurassic.

Late Paleozoic

Interval 12

In Table 4.1 are listed 6 paleomagnetic investigations of rocks with ages between 239 and 260 m.y. (ca. late Permian into middle Permian). Four of these studies are of red beds from Colorado, New Mexico, and Oklahoma, and two are of intrusive igneous rocks from Connecticut and Prince Edward Island. From these studies, 4 poles have been derived (Table 4.1, Figure 4.6), two of which (poles 105 and 106) are not taken directly from the original references.

Pole 105 is a recalculated mean of 5 site poles from Connecticut pegmatites studied by de Boer and Brookins (1972). The mean pole for these rocks listed in the catalogue of McElhinny and Cowley (1977) is incorrect, because the longitude of the VGP for site 5 of de Boer and Brookins (1972) should be 114°E, rather than 144°E as originally reported. Pole 106 is a composite pole, derived from site poles determined by Peterson and Nairn (1971) from red beds of New Mexico and Oklahoma. Pole 106 (52°N, 124°E) is the mode of poles from 6 sites for which the number of stably magnetized samples was at least 5 and the α_{95} was less than 15° (sites M, SR3, D, EC1, EC2, and EC3 of their Table 1). Rocks from all 6 of these sites are of Guadalupian and Ochoan age.

Only two poles from interval 12 are considered here to be true paleomagnetic poles representative of the late Permian for cratonic North America. These are pole 106 (defined above) and pole 103 (a preliminary pole from the Kiaman part of the upper Maroon Formation, Colorado; Christensen and Helsley, 1974b). Both of these poles are from sedimentary rocks that probably represent enough time to have averaged out secular variation.

Other poles from interval 12 may not provide reliable points on the North American APW path. Directions of magnetization from Colorado red beds studied by McMahon and Strangway (1968a,b) are probably biased by a significant component of recent origin (cf. discussion of interval 11 data). Pole 104 (from the Malpeque Bay sill, Prince Edward Island; Larochelle, 1967a) probably represents only a spot reading of the magnetic field and is only imprecisely dated as Permian. Similarly, pole 105 may not have averaged out secular variation, since it is derived from only 5 sites in 2 pegmatites. In addition, both poles 104 and 105 are from New England and the Canadian Maritime provinces, which may have been in motion relative to cratonic North America during at least part of the Late Paleozoic (cf. discussion of poles from interval 14).

The mean of poles 103 and 106 is probably the best estimate of the pole for interval 12 on the North American APW path. This mean pole is nearly 15° southeast of the pole for interval 11, suggesting significant apparent polar wander during the late Permian. However, many more reliable late Permian poles are needed to substantiate this trend.

Interval 13

In Table 4.1 are listed 8 paleomagnetic investigations of rocks with ages between 260 and 281 m.y. (ca. middle Permian to latest Pennsylvanian). All of these studies are of sedimentary rocks (predominantly red beds) of southwestern and western U.S. From these studies, 4 poles have been derived (Table 4.1, Figure 4.6), two of which (poles 107 and

108) are not taken directly from the original references.

Pole 107 is a composite pole, based on the following 11 site poles: 8 poles from southwestern U.S. red beds (Peterson and Nairn, 1971; their Wolfcampian and Leonardian sites UCM2, A5, A6, A7, A8, W2, W5, and AT1, with $N \geq 5$, $\alpha_{95} \leq 15°$); 1 pole from the Halgaito Tongue (Farrell and May, 1971); and 2 poles from the Cutler Formation (Helsley, 1971). Pole 108 is derived from an extensive sampling of the Cutler Group, Utah, by Gose and Helsley (1972); this pole has been recomputed by Scott and Helsley (1975), using the procedure of Helsley (1973) for determining the direction of characteristic magnetization from a distribution biased by an unremoved secondary component.

Three poles were given unit weight in determining the mode for interval 13. These are poles 107 and 108 (defined above) and pole 109, from the Wolfcampian part of the upper Casper Formation, Wyoming (Diehl and Shive, 1976).

Other poles from interval 13 were excluded from the determination of the mode. Poles computed by McMahon and Strangway (1968a,b) are considered to be biased by a secondary component of magnetization, as discussed for interval 11. Pole 110, from the Honaker Trail Formation, Utah, was excluded because of uncertainties (discussed by Scott and Helsley, 1975) concerning the time of acquisition of the characteristic magnetization of these rocks.

The pole determined for interval 13 is within about 5° of that for interval 12.

Interval 14

In Table 4.1 are listed 9 paleomagnetic investigations of rocks with ages between 281 and 302 m.y. (ca. late to middle Pennsylvanian). All of these studies are of sedimentary rocks (predominantly red beds) of western U.S. and the Canadian Maritime provinces. From these studies, 9 poles have been derived (Table 4.1, Figure 4.7), all of which are taken directly from the original references. It should be noted that the age of the Dunkard Formation, from which pole 111 is derived, is now thought to be Late Pennsylvanian (Helsley and Scott, 1975), rather than Early Permian as reported in Helsley (1965).

Five of the poles from interval 14 (poles 113 through 117) are from red beds of the Canadian Maritime provinces (Black, 1964; Roy, 1966; Roy et al., 1968). These poles are very well grouped around a mode near 40°N, 130°E (Figure 4.7). However, Scott (1975) has called attention to the tendency of pre-Middle Permian poles from the Maritimes to lie closer to African and South American poles (after closure of the Atlantic Ocean) than to their time-equivalents from cratonic North America. Indeed, the 3 Late Pennsylvanian poles from the craton (pole 111 from the Dunkard Formation, West Virginia (Helsley, 1965); pole 112 from the lower Maroon Formation, Colorado (Christensen and Helsley, 1974b); and pole 118 from the middle Casper Formation, Wyoming (Diehl and Shive, 1976) do have latitudes typically about 6° higher than poles from the Maritimes (Figure 4.7). This supports the hypothesis of Scott (1975) that the Canadian Maritime provinces and parts of New England (designated the "Maritime block") formed part of an Africa-South America-Maritime plate prior to the Permian collision (Appalachian orogeny)

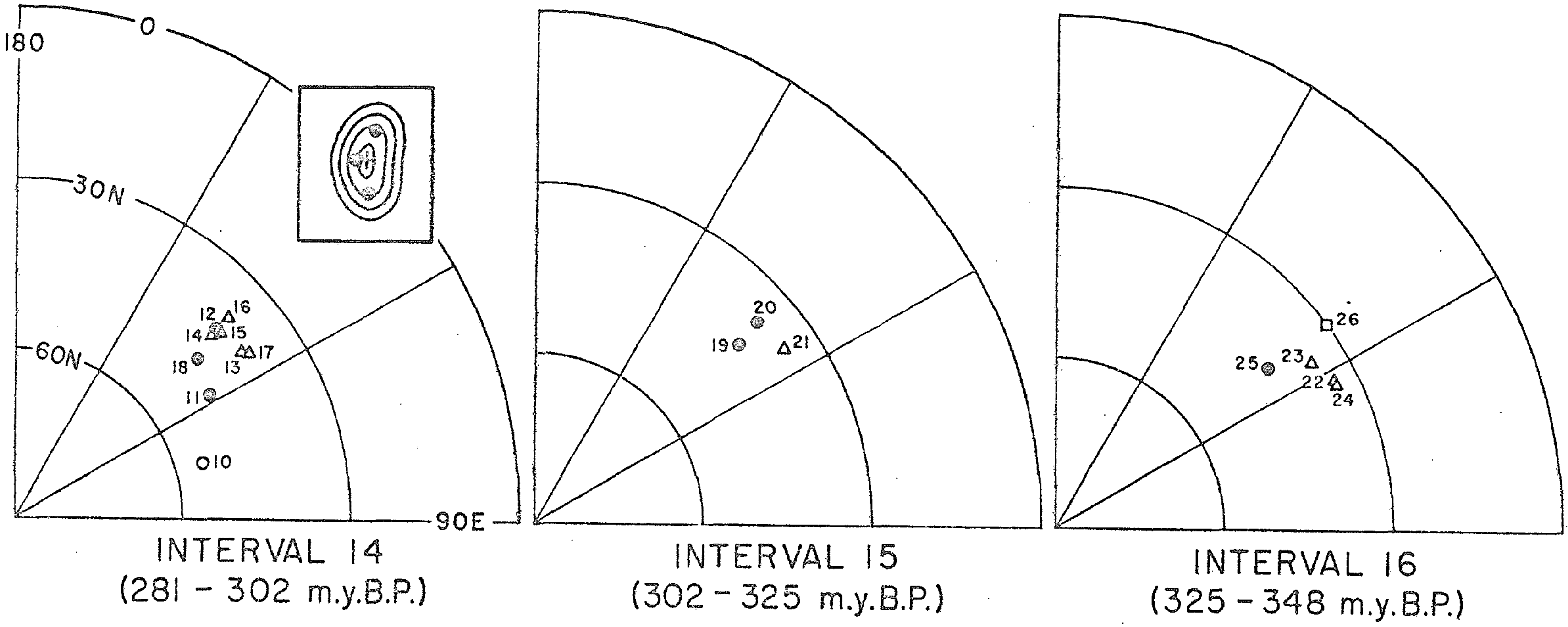


Figure 4.7 North American paleomagnetic poles for intervals 14, 15, and 16 (281 to 348 m.y.B.P.; approx. latest Pennsylvanian to middle Mississippian). Symbols, notation, and contouring procedures are explained in the caption for Fig. 4.2. Poles represented by triangles are from the "Maritime block" (Scott, 1975), and those represented by squares are from western Newfoundland.

with North America. Thus, in this study, Paleozoic poles from the "Maritime block" have been excluded from interval pole determinations. While this drastically reduces the number of poles given unit weight, it should result in a Paleozoic APW path more representative of cratonic North America.

In view of the foregoing discussion, only poles 111, 112, and 118 from cratonic North America were included in the determination of the mode for interval 14. (Poles from McMahon and Strangway (1968a,b) and from Scott and Helsley (1975) were excluded for reasons cited in the intervals 11 and 13 discussions.) The mode for interval 14 (46°N, 129°E) is about 12° southeast of that for interval 13.

Interval 15

In Table 4.1 are listed 5 paleomagnetic investigations of rocks with ages between 302 and 325 m.y. (ca. middle to early Pennsylvanian). These studies are predominantly of hematite-bearing sedimentary rocks from western and eastern U.S. and the Canadian Maritime provinces. From these studies, 3 poles have been derived (Table 4.1, Figure 4.7), all of which are taken directly from the original references.

Pole 119 (from the lower Casper Formation, Wyoming; Diehl and Shive, 1976) and pole 120 (from 9 coal and overclay units, Ohio and West Virginia; Kopacz and Noltimier, 1976) probably best represent the paleomagnetic pole for cratonic North America during interval 15 time. Poles computed by McMahon and Strangway (1968a,b) are probably biased by an unremoved secondary component of magnetization, as discussed in interval 11. Pole 121 (from the Cumberland Group, Nova Scotia; Roy,

1969) is about 5° southwest of the two poles from the craton, consistent with its having been derived from part of the "Maritime block" (cf. interval 14 discussion).

The mean of poles 119 and 120 is considered to be the best estimate of the interval 15 pole for cratonic North America. This pole for interval 15 is about 7° southeast of the pole determined for interval 14. Together, the poles for intervals 15 to 10 (early Pennsylvanian into early Jurassic) lie on an arc with a length of about 40°. Determining whether this apparent polar wandering was smooth or somewhat erratic should be one goal of future paleomagnetic studies of Late Paleozoic and Early Mesozoic rocks from cratonic North America.

Interval 16

In Table 4.1 are listed 5 paleomagnetic investigations of rocks with ages between 325 and 348 m.y. (ca. earliest Pennsylvanian and late Mississippian). These studies are all of sedimentary rocks (predominantly red beds) from eastern U.S. and the Canadian Maritime provinces. From these studies, 5 poles have been derived (Table 4.1, Figure 4.7), all of which are taken directly from the original references.

Pole 125, from the Mauch Chunk Formation, Pennsylvania (Knowles and Opdyke, 1968), is probably the only interval 16 pole representative of "cratonic" North America. The Mauch Chunk study was undertaken partly to test various hypotheses for the origin of the bend (salient) in the Pennsylvania Appalachian Mountains. Results of a fold test, and the consistency of directions from localities on both limbs of the salient, indicate that little if any differential vertical-axis rotation has occurred as a result of the Appalachian orogeny.

The Mauch Chunk pole 125 is about 10° northeast of the 3 well-grouped interval 16 poles from the "Maritime block": poles 122 and 123 from the Hopewell Group, New Brunswick (Roy and Park, 1969, 1974) and pole 124 from the Maringouin Formation, New Brunswick (Roy and Robertson, 1968). This discordance is another example of the systematic discrepancy between pre-middle Permian poles from cratonic North America and poles from the Canadian Maritime provinces and New England (cf. discussions by Scott (1975) and Kent and Opdyke (1977b)). Pole 126, from the Mississippian Codroy Group, western Newfoundland (Black, 1964) also differs somewhat from the Mauch Chunk pole and from the cluster of poles from the Canadian Maritimes. This may reflect a post-Mississippian rotation of western Newfoundland with respect to cratonic North America (cf. discussions of poles from western Newfoundland for intervals 19, 23, 24, 27, 28, and 29).

The pole from the Mauch Chunk Formation is slightly northwest of the pole determined for interval 15. Thus, the 40° apparent polar shift of the Late Paleozoic and Early Mesozoic does not seem to have begun until after the late Mississippian.

Interval 17

In Table 4.1 are listed 3 paleomagnetic investigations of rocks with ages between 348 and 371 m.y. (ca. middle and early Mississippian). These studies are of sedimentary and igneous rocks from central and northeastern U.S. and western Newfoundland. From these studies, 4 poles have been derived (Table 4.1,Figure 4.8), all of which are taken directly from the original references.

Poles 127 and 128, from two sections of the St. Joe Limestone, Arkansas (Scott, 1975), are probably the only interval 17 poles representative of cratonic North America. Pole 129, from a mafic complex in eastern Massachusetts, was thought by Schutts _et al._ (1976) to date from a thermal and metasomatic event at about 370±10 m.y.B.P.; its position more than 15° southwest of poles from the St. Joe Limestone is consistent with eastern Massachusetts' having been part of the "Maritime block" (Scott, 1975) in the late Devonian to early Mississippian. Pole 126, from the Codroy Group of western Newfoundland (Black, 1964), also is southwest of the poles from the St. Joe Limestone, suggesting a possible post-Mississippian rotation of western Newfoundland with respect to the North American craton.

The best estimate of the interval 17 pole for cratonic North America is taken to be the mean of poles 127 and 128. These poles from the two sections of the St. Joe Limestone are about 5° from each other, and their mean is about 5° from poles for intervals 16 and 15. Thus, the pole appears to have been nearly stationary with respect to North America for most of Mississippian through middle Pennsylvanian time.

Interval 18

In Table 4.1 are listed 9 paleomagnetic investigations of rocks with ages between 371 and 394 m.y. (ca. late Devonian to middle Devonian). About half these studies report results from hematitic sedimentary rocks from southwestern and eastern U.S. and New Brunswick.

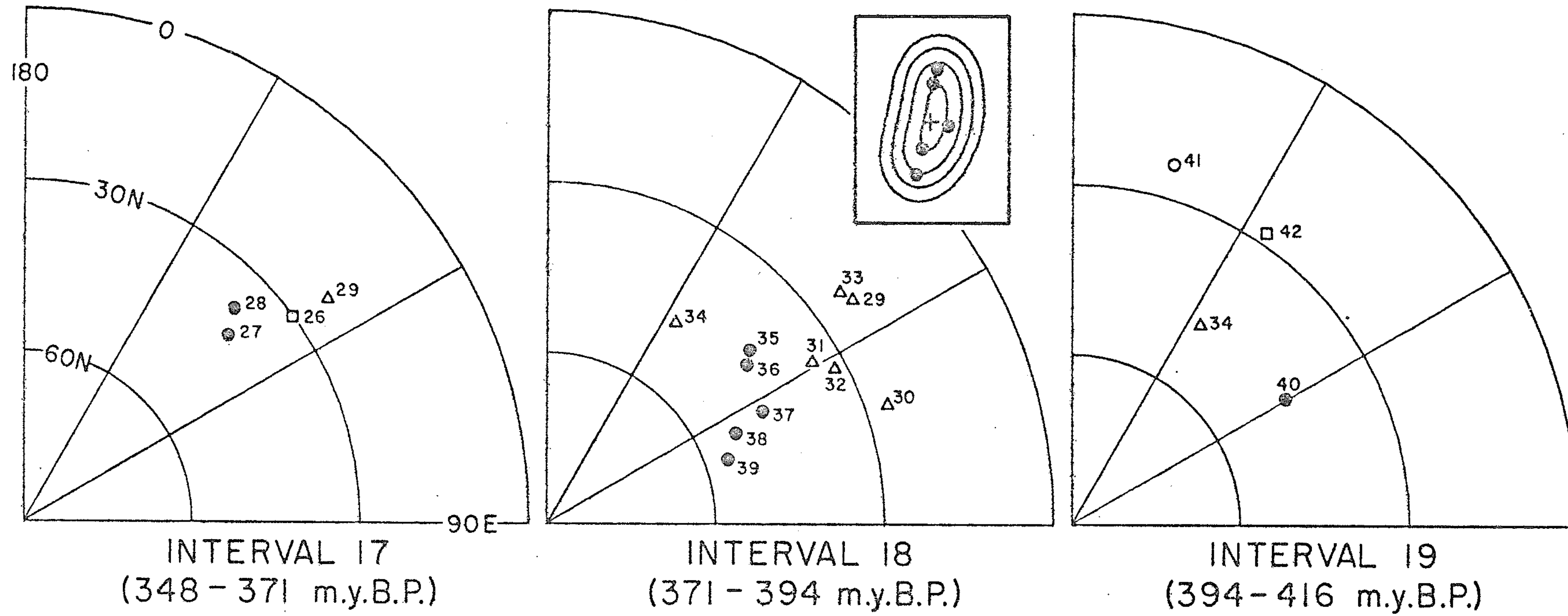


Figure 4.8 North American paleomagnetic poles for intervals 17, 18, and 19 (348 to 416 m.y.B.P.; approx. middle Mississippian to latest Silurian). Symbols, notation, and contouring procedures are explained in the captions for Figs. 4.2 and 4.7.

The other half report results from intrusive and extrusive igneous rocks from New England and New Brunswick. From these studies, 11 poles have been derived (Table 4.1, Figure 4.8), all of which are taken directly from the original references.

If the base of interval 18 is defined as middle Middle Devonian (ca. early Givetian), then there are 5 paleomagnetic poles for this interval from cratonic North America. Three of these poles are from the "Catskill Formation", a time-transgressive, deltaic sequence of the Appalachian basin. Poles 135 (Phillips, 1966) and 136 (French, 1976) are from localities in West Virginia, Maryland, and Pennsylvania, where the Catskill red beds are pre-middle Frasnian (Oliver _et al_., 1967). Pole 137, from the Catskill sequence in New York (Kent and Opdyke, 1977a), may be slightly older than poles 135 and 136 and is about 10° northwest of them. Pole 137 is very close to a preliminary pole from the early Late Devonian (early Frasnian) Genesee Group, western New York (Van Alstine and Shoemaker, unpublished data). Poles 138 and 139 are from the Temple Butte Limestone and Martin Formation, Arizona (Elston and Bressler, 1977), which are of late Middle(?) (latest Givetian?) and early Late (Frasnian) Devonian age. These two poles from the Colorado Plateau are northwest of poles 135, 136, and 137 from the Catskill Formation of eastern U.S. (Figure 4.8).

Six poles (129 - 134) for interval 18 have been obtained from the "Maritime block" (Scott, 1975). Pole 129, from a mafic complex in eastern Massachusetts (Schutts _et al_., 1976), and poles 130 to 133, from lavas and red beds of the Perry Formation in New Brunswick and Maine (Black, 1964; Robertson _et al_., 1968; Phillips and Heroy, 1966), form a cluster about 20° south of coeval poles from the craton (Figure 4.8). This difference again supports Scott's (1975) contention that New England and the Canadian Maritime provinces were in motion relative to cratonic North America in Late Paleozoic time (cf. Schutts *et al.*, 1976; Kent and Opdyke, 1977b). Pole 134, from the Belchertown pluton, west-central Massachusetts (Ashwal and Hargraves, 1977) diverges from the clusters formed by the cratonic and Maritime poles. However, this pluton was thought by Ashwal and Hargraves (1977) to have experienced a post-magnetization rotation about a *horizontal* axis.

The mode of the 5 poles from interval 18 representing cratonic North America is at 48°N, 121°E, about 15° northwest of the pole determined for interval 17. This difference suggests that the late Devonian was a time of rapid apparent polar wandering with respect to North America, which might also account for the linear distribution formed by the 5 poles from the craton (Figure 4.8). Apparent polar wandering in a southeasterly direction during the late Devonian is consistent with the relative ages of the strata from which these poles are derived.

Interval 19

In Table 4.1 are listed 3 paleomagnetic investigations of rocks with ages between 394 and 416 m.y. (middle Devonian to latest Silurian). Two of these studies are of sedimentary rocks from Ohio and western Newfoundland, and one is of a pluton from Massachusetts. From these studies, 4 poles have been derived (Table 4.1, Figure 4.8), all of which are taken directly from the original references.

Only two of the poles for interval 19 are from cratonic North America. In a study of limestones from Ohio, Martin (1975) reported what he considered to be a polarity transition between the Raisin River Dolomite and lower Columbus Limestone pole 141 (late Emsian) and the upper Columbus and Delaware Limestone pole 140 (middle to late Eifelian). Whereas pole 140 is based on 119 samples of reversed polarity, pole 141 is derived from "normal"-polarity directions from only 10 samples. Clearly, if poles 140 and 141 do represent reversed and normal-polarity directions, the inferred polarity transition must have been asymmetric. Alternatively, pole 141 may reflect an anomalous or metastable state of the geomagnetic field, like that suggested by Sallomy and Piper (1973) to explain anomalous directions of magnetization in some Lower Devonian rocks from Scotland.

The other two poles for interval 19 are from the "Maritime block" (Scott, 1975) and western Newfoundland. Pole 134, from the Belchertown pluton, west-central Massachusetts, was thought by Ashwal and Hargraves (1977) to have experienced a post-magnetization rotation about a horizontal axis. (Because Ashwal and Hargraves cite the Early Devonian and 380±5 m.y.B.P. as the time of emplacement of this pluton, pole 134 has been included in both intervals 18 and 19 of this study.) Pole 142 is from the Early Devonian (Gedinnian?) Clam Bank Group of western Newfoundland (Black, 1964), which might also have rotated with respect to cratonic North America during the Paleozoic (cf. discussions of poles from western Newfoundland for intervals 16, 23, 24, 27, 28, and 29). It may be significant that pole 142 is similar to pole 141 from Ohio limestones, and that both these poles are derived from "normal"-

polarity directions.

The proximity of pole 140 from the upper Columbus and Delaware Limestones to the pole determined for interval 18 suggests that a pole near 46°N, 120°E is typical of the early Middle Devonian for North America.

Early Paleozoic

Interval 20

In Table 4.1 are listed 3 paleomagnetic investigations of rocks with ages between 416 and 438 m.y. (ca. late and middle Silurian). Two of these studies are of hematitic sedimentary rocks from eastern U.S. and one is of carbonates from Ohio. From these studies, 3 poles have been derived (Table 4.1, Figure 4.9). All of these poles are taken directly from the original references, except pole 144, which is cited in Reeve (1975).

Two of the poles for interval 20 are from sedimentary rocks of the central Appalachian Mountains. Pole 143, from the Late Silurian Bloomsburg Formation, Pennsylvania and Maryland (Roy et al., 1967), is derived from sites with normal-polarity directions; reversed directions may also be present at other sites, but they could not be distinguished from a reversed-polarity secondary magnetization attributed to the Permian Appalachian orogeny. Similarly, effects of this orogeny apparently have obscured reversed-polarity directions in the Middle Silurian Rose Hill Formation, studied by French and Van der Voo (1977). However, directions from 12 samples of the Rose Hill (with a mean pole 145) are of normal polarity and seem to have been acquired prior to folding of the beds.

The other pole from interval 20 is derived from the Late Silurian Tymochtee Formation, Ohio (McMahon, 1974), which contains a reversed-polarity magnetization represented by pole 144. This pole from the Tymochtee is about 8° from pole 143 from the Bloomsburg

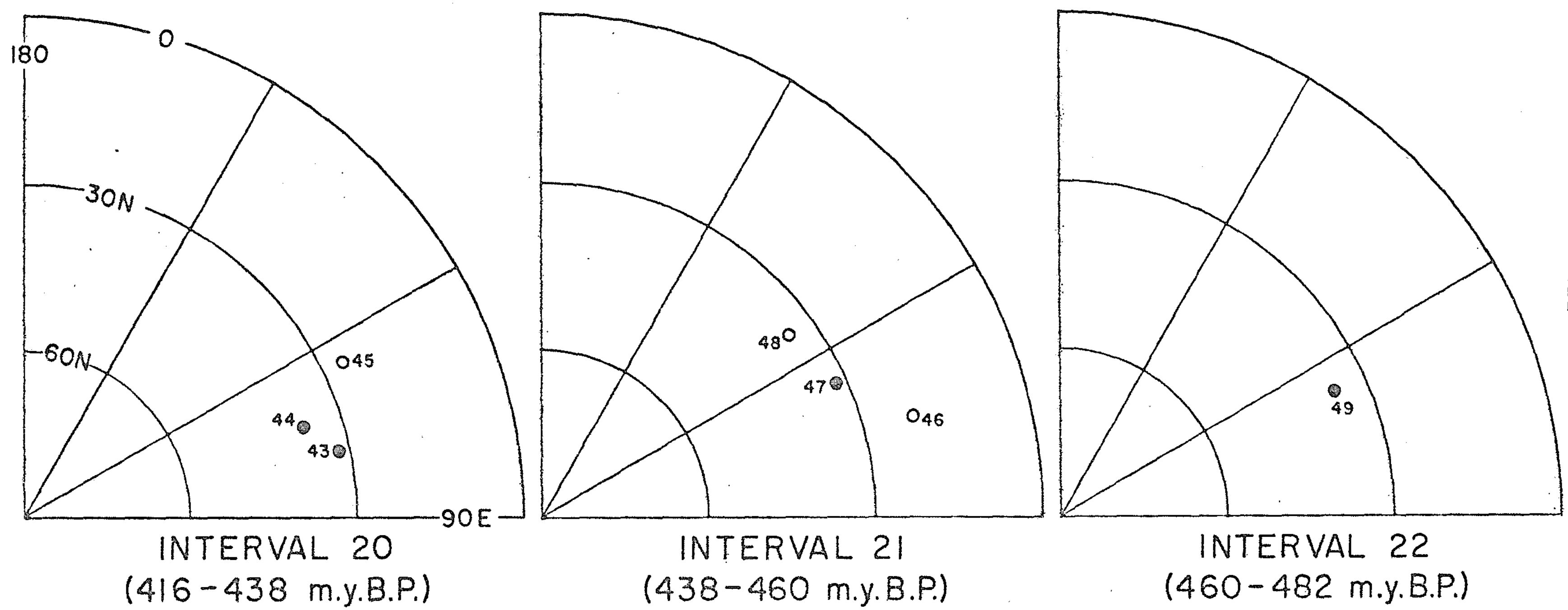


Figure 4.9 North American paleomagnetic poles for intervals 20, 21, and 22 (416 to 482 m.y.B.P.; approx. late Silurian to middle Ordovician). Symbols and notation are explained in the caption for Fig. 4.2.

Formation. The concordance of these two poles, which together represent both polarities of the geomagnetic field, suggests that the central Appalachian Mountains have not rotated appreciably about a vertical axis with respect to the craton (McMahon, 1974).

The best estimate of the pole for interval 20 is taken to be the mean of poles 143 (Bloomsburg Formation) and 144 (Tymochtee Formation). Pole 145 (Rose Hill Formation) was excluded from the interval pole determination, because French and Van der Voo (1977) did not consider it as firmly established (in view of the limited number of samples).

The pole determined for interval 20 (35°N, 105°E) is about 20° southwest of the pole for interval 19. This suggests that the late Silurian through early Devonian was a time of comparatively rapid apparent polar wandering with respect to North America.

Interval 21

In Table 4.1 are listed 3 paleomagnetic investigations of rocks with ages between 438 and 460 m.y. (ca. early Silurian and late Ordovician). Two of these studies are of sedimentary rocks and one is of a plutonic complex, all from the central Appalachian Mountains. From these studies, 3 poles have been derived (Table 4.1, Figure 4.9), all of which are taken directly from the original references.

Only one of these poles (pole 147) is considered here to be representative of the pole for interval 21. Pole 147 (32°N, 114°E) is from hematitic sedimentary rocks of the Late Ordovician Juniata Formation, sampled in Virginia, West Virginia, and Pennsylvania by Van der Voo and French (1977). This pole is the mean of 17 site poles, has an associated α_{95} of 5°, and represents both normal and reversed-polarity directions that are within 8° of being antiparallel. Results of a fold test indicate that the characteristic magnetization of the Juniata was acquired prior to folding of the beds in the Late Paleozoic.

The other two poles from interval 21 were excluded from the determination of the interval pole. Pole 146, which has an α_{95} of about 10°, is from the Early Silurian Castanea Formation, Pennsylvania. This pole is taken from Table 4 of McElhinny and Opdyke (1973), although no details are given from which the reliability of this result can be judged. Pole 148, from the Beemerville complex, New Jersey (Proko and Hargraves, 1973) is rather imprecise (α_{95} = 17°) and may not have averaged out secular variation, since it is based on VGP's from only 4 sites.

Interval 22

In Table 4.1 is listed only one paleomagnetic investigation of rocks with ages between 460 and 482 m.y. (ca. late to middle Ordovician). This study is of the Middle Ordovician "Trenton" (actually Denley and Steuben) Limestone, New York (McElhinny and Opdyke, 1973). The pole derived from the Trenton, pole 149 (Figure 4.9), is based on reversed-polarity directions of magnetization from 29 samples. These samples are from flat-lying rocks in which the characteristic magnetization seems to reside in titanomagnetite (McElhinny and Opdyke, 1973). The characteristic magnetization direction from the Trenton Limestone is nearly identical to reversed-polarity magnetization directions obtained from the nearly coeval Middle Ordovician Canajoharie Shale (Van Alstine, unpublished data). As in the Trenton, the characteristic magnetization of the nearly black Canajoharie Shale also appears to reside in magnetite.

Pole 149 (36°N, 114°E) from the Trenton is considered here to represent interval 22 on the North American APW path. This pole for interval 22 is within 4° of the pole derived for interval 21. In fact, poles 149 (Trenton), 147 (Juniata), 144 (Tymochtee), and 143 (Bloomsburg) form a distinct cluster, about 15° removed from Late Paleozoic poles. This suggests that the paleogeographic pole was quasi-static with respect to North America from middle Ordovician through late Silurian time.

<u>Interval 23</u>

In Table 4.1 are listed 8 paleomagnetic investigations of rocks with ages between 482 and 504 m.y. (ca. middle to early Ordovician). These studies are of igneous and sedimentary rocks from west-central U.S., Quebec, and Newfoundland. From these studies, 12 poles have been derived (Table 4.1, Figure 4.10), all of which are taken directly from the original references.

It is quite possible that none of the poles listed in Table 4.1 is representative of the pole for interval 23 with respect to cratonic North America. Indeed, only three studies of rocks of

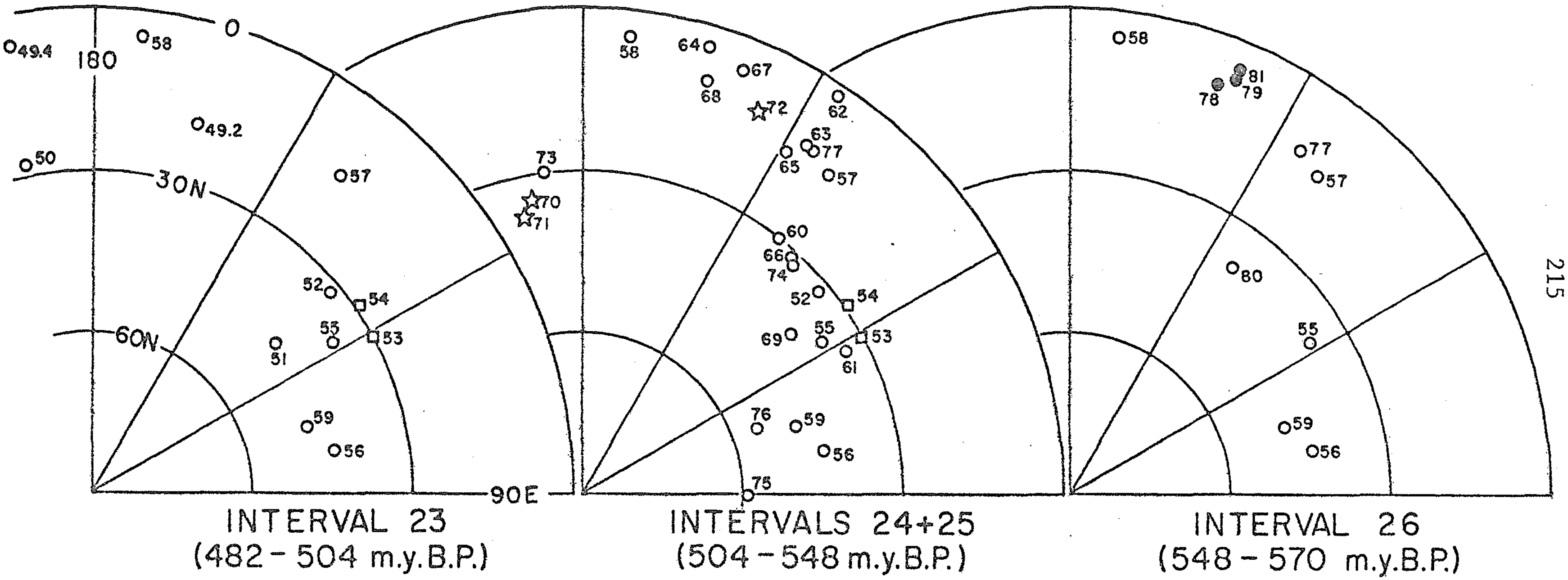


Figure 4.10 North American paleomagnetic poles for intervals 23, 24, 25, and 26 (482 to 570 m.y.B.P.; approx. middle Ordovician to early Cambrian). Symbols and notation are explained in captions for Figs. 4.2 and 4.7. Stars represent poles in the southern hemisphere. Contours for interval 26 were too peaked to show at the scale of this figure.

this age are demonstrably from the craton. Steiner (1973) reported reversed-polarity directions of magnetization (represented by pole 151) from hematitic shale and limestone of the West Spring Creek Formation, Oklahoma; however, results of a fold test indicate that this magnetization was probably acquired during the Arbuckle orogeny, in the late Pennsylvanian to early Permian. Larson and Mutschler (1971) and French et al. (1977) reported multimodal distributions of directions (represented by poles 155 through 159) from intrusives in central Colorado. In view of the multiple components of magnetization in these rocks, as well as the wide range in their reported isotopic ages (704 to 485 m.y.B.P.), French et al. (1977) suggested that the poles derived from these intrusives may well span the entire period from the late Precambrian into the Ordovician.

The other five paleomagnetic investigations of rocks from interval 23 are from regions that may have been in motion relative to cratonic North America during the Paleozoic. Seguin (1977) studied volcanic rocks near the structurally complex Appalachian front of southern Quebec. Poles derived from AF (pole 149.2) and thermally (pole 149.4) cleaned directions are about 30° apart (Figure 4.10), which confuses interpretation of their significance. Moreover, this part of southern Quebec might have been part of the "Maritime block" (Scott, 1975) during the Ordovician. The remaining four studies are of rocks from different parts of Newfoundland: carbonates of the St. George Formation in the west (Beales et al., 1974; Deutsch and Rao, 1977a); basalt of the Moreton's Harbour Group in the center (Deutsch and Rao, 1977a); and hematitic sandstone of the Wabana Group in the east (Deutsch and Rao, 1970). The two poles derived from the western part of the island are within about 10° of the Middle Ordovician pole from the Trenton Limestone of New York; the pole from central Newfoundland is displaced about 10° farther to the east; and the pole from eastern Newfoundland is 60° even farther to the east (Figure 4.10). The divergence of poles from eastern and western Newfoundland is to be expected if Newfoundland had been split by a proto-Atlantic Ocean during part of the Early Paleozoic; such a separation was indeed proposed by Wilson (1966) on the basis of the contrasting Pacific and Atlantic faunas from opposite sides of the island.

In contrast to pole 150 from eastern Newfoundland, poles 152, 153, and 154 from the central and western part of that island lie much closer to pole 149 from the slightly younger Trenton Limestone of New York. This suggests that at least the western part of Newfoundland may not have rotated grossly with respect to cratonic North America during the Paleozoic. However, a 30° anticlockwise rotation of Newfoundland with respect to Labrador had been proposed by Wegener (1929). Recently, Deutsch and Rao (1977a) presented paleomagnetic results contradicting an earlier test (Black, 1964) of Wegener's proposed rotation. By comparing paleomagnetic poles from late Precambrian to Ordovician rocks of western Newfoundland with coeval poles from the mainland, Deutsch and Rao (1977a) concluded that the data are inconsistent with the full 30° rotation proposed by Wegener, although a rotation of 5° to 10° might have occurred.

In view of the uncertainties in the tectonic history of Newfoundland, poles from that island have been excluded from interval pole determinations of this study. It is recognized, however, that poles from western Newfoundland probably lie within about 30° of coeval poles from cratonic North America. Thus, the proximity of poles 153 and 154 (St. George Formation) from western Newfoundland to pole 149 (Trenton Limestone) from New York suggests that no more than about 15° of apparent polar wandering occurred with respect to North America during the early to middle Ordovician.

This interpretation of the paleomagnetic data from interval 23 is consistent with preliminary paleomagnetic results obtained from the uppermost Goodwin Limestone, Nevada (Gillett and Van Alstine, unpublished data). About 50 samples from the Goodwin yield reversed-polarity characteristic magnetizations corresponding to a pole near that of the contemporaneous (uppermost Canadian) West Spring Creek Formation studied by Steiner (1973). This suggests that the West Spring Creek pole 151 may be close to the pole for the late Early Ordovician, although perhaps only coincidentally. Tentatively, a pole near 46°N, 129°E is thought to be the best estimate of the pole for interval 23 with respect to North America.

Intervals 24 and 25

In Table 4.1 are listed 13 paleomagnetic investigations of rocks with ages between 504 and 548 m.y. (ca. earliest Ordovician and late Cambrian). These studies are of igneous and sedimentary rocks from western and central U.S., Quebec, and Newfoundland. From these studies, 26 poles have been derived (Table 4.1, Figure 4.10), only one of which (pole 160) is not taken directly from the original references. Pole 160 is a pole recalculated by Van der Voo et al. (1976) based on incorporation of four normal-polarity site poles into the pole calculated by Ku et al. (1967) for the Wichita granites, Oklahoma.

Determination of poles representative of intervals 24 and 25 is complicated by (1) the presence of multiple components of magnetization in many of the rocks from these intervals, and (2) a remarkable, nearly great-circle distribution of poles over approximately 125° of arc. (Figure 4.10)

Twelve of the 26 poles from intervals 24 and 25 are derived from intrusive igneous rocks from cratonic North America. Seven of these poles (160 through 166 of Table 4.1) are from extensive paleomagnetic investigations of the Wichita granites, Oklahoma (Ku et al., 1967; Spall, 1968a, 1970; Vincenz et al., 1975). The 7 poles from the Wichita granites are distributed roughly along a great circle between pole 161 (34°N, 118°E) and pole 164 (4°N, 164°E) (Figure 4.10). One explanation for this linear distribution was provided by Vincenz et al. (1975), who cited paleomagnetic and rock magnetic evidence for a pervasive remagnetization of these rocks. They suggested that this remagnetization resulted from extreme hydrothermal alteration of the Wichita granites, probably during regional uplift in Pennsylvanian time. They admit, however, that (1) even some rocks with an appreciable amount of unaltered, primary magnetite have directions of magnetization near that of the late Paleozoic geomagnetic field, and (2) the hydrothermal alteration of these rocks did not necessarily occur during the late Paleozoic tectonism.

The other five poles from igneous rocks with possibly late Cambrian ages are derived from studies of alkalic intrusive complexes in central Colorado (Larson and Mutschler, 1971; French et al., 1977). Although these rocks are less altered than the Wichita granites, and although the primary carrier of the characteristic magnetizations in the Colorado intrusives is magnetite (French et al., 1977), the igneous rocks from Colorado also yield multimodal distributions of magnetization directions. Each of the three poles (157, 158, and 159) derived from the Colorado intrusives studied by French et al. (1977) represents a two-polarity magnetization with roughly antiparallel directions. There appears to be no clear-cut correlation between rock type (gabbro, syenite, and lamprophyre dikes) and pole position. The 5 poles from the two studies of the Colorado intrusives (poles 155 through 159) are distributed along nearly the same great-circle path defined by poles from the Wichita granites, although the total arc-length represented by the Colorado intrusives is somewhat greater, between pole 158 (5°N, 174°E) and pole 156 (44°N, 100°E). French et al. (1977) suggested that the multivectorial nature of the magnetization of the Colorado intrusives might reflect their emplacement over the full time-span represented by the range in isotopic ages reported for these rocks (704 to 485 m.y.B.P.).

Curiously, the multiple components of magnetization from the Colorado and Oklahoma intrusives yield poles that are remarkably similar to some of those obtained from Late Cambrian sedimentary rocks in Texas and Missouri. Reeve (1975) reported reversed-polarity magnetizations from hematitic sedimentary rocks of the Riley and Wilberns Formations, Texas, spanning most of the time between the latest Middle Cambrian through the middle Late Cambrian. He found that poles from 9 different units of these formations fall roughly on a great circle between pole 169 (41°N, 127°E), from the Lion Mountain Sandstone member of the Riley Formation, and pole 168 (10°N, 163°E), from the Point Peak Shale member of the Wilberns Formation (from which Van der Voo et al. (1976) obtained the very similar pole 167). Reeve noted that the position of a given pole along this arc is not a simple function of the stratigraphic position of the rocks from which it is derived. Reeve suggested that the directions of magnetization observed in these rocks generally represent the vector sum of two components of magnetization that cannot be cleanly separated by either AF or thermal demagnetization techniques. Pole 169 from the Lion Mountain Sandstone was thought to contain the greatest proportion of a component roughly aligned with the Late Paleozoic field direction, and pole 168 from the Point Peak Member was thought to be dominated by a component acquired in the

Late Cambrian. Reeve assumed that pole 168 is more representative of the Late Cambrian geomagnetic field, largely because of its divergence from Late Paleozoic poles, and also because of its proximity to some of the poles obtained from Late Cambrian rocks of Missouri (discussed below). Reeve considered the possibility that the magnetization represented by pole 169 was acquired during a period of tectonic disturbance during the late Pennsylvanian. He concluded that this magnetization is probably a chemical remanent magnetization (CRM), especially since the maximum temperature to which these rocks have been subjected was estimated to be less than 100°C since the late Devonian. (It should be noted that the arc defined by poles from the Wilberns and Riley Formations is concordant with that defined by the poles from the Wichita granites.)

Other Late Cambrian multivectorial magnetizations have been reported from flat-lying sandstones of the Lamotte Formation in Missouri (Al-Kafaji and Vincenz, 1971). Two poles were derived from reddish, coarser-grained rocks in the lower part of the formation: Pole 173, representing normal-polarity "Group I" directions, and pole 174, representing reversed-polarity "locality E" directions. On the basis of rock magnetic studies, Al-Kafaji and Vincenz concluded that these magnetizations are largely CRM's of secondary origin, the locality E directions reflecting remagnetization in the Carboniferous to Triassic, and the Group I directions perhaps reflecting superposition of this component with a Cambrian component. In contrast, poles 171 and 172 (both representing reversed-polarity magnetizations) are derived from fine-grained, white sandstone, which is the dominant rock type of the Lamotte Formation. These two significantly different poles come from different stratigraphic horizons; rocks containing "Group IIa" magnetizations (represented by pole 171) underlie rocks containing "Group IIb" directions (represented by pole 172). Partly on the basis of rock magnetic studies, Al-Kafaji and Vincenz (1971) concluded that the magnetizations represented by poles 171 and 172 are primary depositional remanent magnetizations (DRM) which have recorded two somewhat different positions of the Late Cambrian geomagnetic field.

Late Cambrian acquisition of the component represented by Lamotte pole 171 is supported by its proximity to pole 170, from the stratigraphically overlying (Dresbachian) Bonneterre dolostone, Missouri (Beales et al., 1974). Pole 170, representing reversed-polarity directions from galena ore and host rock of the Bonneterre, is only about 4° from pole 171 from the Lamotte (Figure 4.10). This suggests that the geomagnetic pole occupied a position near 36°S, 191°E during at least part of Late Cambrian time.

Poles markedly different from that of the Bonneterre have been reported from Cambrian sedimentary rocks of the Grand Canyon and southeast Arizona (Elston and Bressler, 1977). Pole 175 is derived from a 1 m thick hematitic oolite bed in the Abrigo Formation, for which faunal evidence indicates a late Middle or early Late Cambrian age. The pole from the Abrigo is similar to pole 176, from the hematitic Muav Limestone of Middle Cambrian age. Field evidence supports a Cambrian time of acquisition of the hematite in these strata. Both poles 175 and 176 represent magnetization directions that have unimodal, nearly axially-symmetric distributions, with means that move less than 10° upon thermal demagnetization to 500°C. Thus, unlike the Wichita granites, Colorado intrusives, and Wilberns, Riley, and Lamotte Formations, the Muav and Abrigo Formations contain only a single component of magnetization.

Four poles have been obtained from rocks of Newfoundland and Quebec that may have ages falling within intervals 24 and 25: pole 152 (from the Moreton's Harbour basalt, central Newfoundland; Deutsch and Rao, 1977a); poles 153 and 154 (from the St. George Limestone of western Newfoundland; Deutsch and Rao, 1977a; Beales et al., 1974); and pole 177 (from an ophiolite complex in southern Quebec; Seguin, 1976). These poles all lie roughly on the arc defined by possibly contemporaneous poles from the craton. However, the ages of these poles are not precisely known, and some could be younger (poles 152, 153, and 154) or older (pole 177) than the time represented by intervals 24 and 25. Moreover, these poles are derived from rocks either in structurally complex settings or from regions that might have been in motion relative to cratonic North America during the Paleozoic (cf. interval 23 discussion). Thus, the significance of these four poles from Newfoundland and Quebec is difficult to assess, especially in view of the scatter of poles from the craton.

Partly to help resolve the ambiguity of the currently available poles from the Late Cambrian of North America, the author has undertaken paleomagnetic investigations of Early Paleozoic sedimentary rocks in New York and in the Canadian Rocky Mountains of Alberta.

Although this work is still in progress, a preliminary pole calculated from reversed-polarity directions of the earliest Ordovician Tribes Hill Formation (New York) is at 35°N, 128°E. A similar provisional pole is obtained from reversed-polarity directions found in the Lyell and Bison Creek Formations (Alberta) of Late Cambrian age (late Dresbachian through Franconian). The characteristic magnetizations of the Tribes Hill, Lyell, and Bison Creek Formations reside in components with coercivities typical of magnetite, and are derived predominantly from gray limestone and dolomitic sandstone.

Poles from intervals 24 and 25 lie roughly along a great-circle between pole 175 (Abrigo) and poles 170-171 (Bonneterre-Lamotte) (Figure 4.10). If rocks represented by these poles acquired their characteristic magnetizations penecontemporaneously with deposition, then there is no simple, unidirectional progression of poles as a function of age. In fact, based on the relative ages of the Wilberns, Lamotte, Bonneterre, Abrigo, Muav, Bison Creek, Lyell, and Tribes Hill Formations, the paleogeographic pole could not possibly have occupied all the positions indicated by the paleomagnetic data, unless APW rates were more than an order of magnitude faster than the 0.3°/m.y. typical of most continents for the Phanerozoic (McElhinny, 1973). Nor can one invoke any simple "remagnetization hypothesis," like that proposed by Creer (1967) (summarized in McElhinny and Opdyke, 1973) to explain some anomalous paleomagnetic results from Devonian rocks of Britain. The remagnetization hypothesis purportedly can account for secondary CRM's in rocks which are presently hematitic and which experienced near-surface oxidation in the Late Paleozoic. Yet, even those formations from intervals 24 and 25 in which the characteristic magnetization is likely to be a DRM (e.g., Lamotte, Bonneterre, Lyell, Bison Creek, Tribes Hill) have grossly discordant pole positions.

Since the distribution of poles for intervals 24 and 25 essentially reflects differences in declination, such a distribution could, in principle, result from large vertical-axis rotations. Indeed, such distributions of poles have been observed in paleomagnetic studies in the Wyoming thrust belt (Grubbs and Van der Voo, 1976) and in the western Cordillera (Beck, 1976). However, large vertical-axis rotations cannot account for the elongate distribution of poles from intervals 24 and 25, because many of these poles are derived from tectonically stable regions on the craton.

One might speculate that the scatter of poles from these two intervals reflects complex behavior of the late Middle Cambrian and Late Cambrian geomagnetic field. Specifically, this great-circle distribution could have been caused by long-period ($\sim 10^6$ years), large ($\pm 45°$) declination swings about a pole that was itself migrating from the position occupied in interval 26 to that of interval 23. These declination swings might have been similar to those reported by Helsley and Steiner (1974) in the Moenkopi Formation, of Early Triassic age (cf. interval 11 discussion). However, the amplitudes of the hypothetical Late Cambrian oscillations must have been about three times larger than those recorded in the Moenkopi. If this explanation were correct, then these oscillations should be observable in a single stratigraphic section in which the characteristic magnetization is a DRM or early-formed CRM. Perhaps Lamotte Formation poles 171 and 172, which differ by about 50° from each other, and which were thought by Al-Khafaji and Vincenz (1971) to represent DRM's, have recorded different parts of a Late Cambrian oscillation.

From the data presently available, it is not yet possible to reconstruct unambiguously the North American APW path from the late Middle Cambrian into the Early Ordovician. However, a pole near 35°N, 128°E (preliminary Tribes Hill) is tentatively accepted as being the best estimate of the pole for interval 24. The pole for interval 25 has been interpolated between those of intervals 24 and 26. Clearly, many more paleomagnetic investigations of Late Cambrian rocks are needed to resolve the ambiguities in the data summarized above.

Interval 26

In Table 4.1 are listed 6 paleomagnetic investigations of rocks with ages between 548 and 570 m.y. (ca. middle to early Cambrian). These studies are of igneous and sedimentary rocks from western and eastern U.S. and from southern Quebec. From these studies, 10 poles have been derived (Table 4.1, Figure 4.10), all of which are taken directly from the original references.

Four of these poles are from Early and Middle Cambrian sedimentary rocks of Arizona, Nevada, and Tennessee. Probably the pole most representative of interval 26 with respect to cratonic North America is pole 181, derived from the Tapeats Sandstone in the Grand Canyon, Arizona (Elston and Bressler, 1977). The Tapeats pole, which has an α_{95} of 3°, is based on 129 samples exhibiting antiparallel normal and reversed-polarity directions. The Tapeats pole is nearly identical to poles from the slightly younger Carrara and Bonanza King Formations (poles 179 and 178), after correction for 36° of clockwise rotation of the sample locality in the Desert Range, Nevada (Chapter 2). (Independent geological evidence for clockwise vertical-axis rotations of this magnitude in the southern Great Basin has been summarized in Chapter 2.) Pole 180 is derived from sampling of the Rome Formation (Early and/or Middle Cambrian) in four thrust sheets in the southern Appalachian Mountains of Tennessee (French, 1976). As noted by French, if the pole from the Rome Formation were corrected for about 30° of clockwise vertical-axis rotation (to line up the structural trends of the southern and central Appalachian Mountains), this pole would become concordant with the pole from the Tapeats. The Tapeats pole 181 is also consistent with poles derived from limestones representing most of Middle Cambrian time from the Canadian Rocky Mountains, Alberta (Van Alstine, unpublished data).

The other 6 poles with ages possibly falling in interval 26 are from igneous rocks of Colorado and Quebec. Five of these poles (155 through 159) are from alkalic igneous complexes in central Colorado (Larson and Mutschler, 1971; French _et al._, 1977), which might have been emplaced over a considerable time-span (704 to 485 m.y.B.P.). However, the discordance between poles derived from these rocks and the pole from the Tapeats suggests that the intrusives were not emplaced during Tapeats time. (The significance of these poles from the Colorado intrusives has been discussed in greater detail in intervals 23 and 24-25.) The age of pole 177, from an ophiolite complex in southern Quebec (Seguin, 1976), is also imprecisely known, and this pole is derived from rocks in a structurally complex setting; hence, its discordance with the pole from the Tapeats is not surprising.

The pole for interval 26 is taken to be the mode of pole 181 (Tapeats) and poles 178 and 179 (corrected poles from the Carrara and Bonanza King Formations). This mode (6°N, 158°E) differs by only 1° from the pole for the Tapeats (5°N, 158°E). Apparently, the late Early and early Middle Cambrian pole as seen from North America occupied a position nearly 60° from the pole for the Middle Ordovician (interval 22). Determining the precise timing of this shift is one goal of current paleomagnetic research. From the data presently available to this author, the shift seems to have occurred between middle Middle Cambrian and earliest Ordovician time.

Interval 27

In Table 4.1 are listed 7 paleomagnetic investigations of rocks with ages possibly between 570 and 592 m.y. (ca. early Cambrian and perhaps very latest Precambrian). Five of these studies are of sedimentary rocks (predominantly red beds) from western and eastern U.S. and from eastern Canada; the other two are of intrusive igneous rocks from Colorado. From these studies, 10 poles have been derived (Table 4.1, Figure 4.11), all of which are taken directly from the original references.

The position of the pole in the early Cambrian (interval 27) evidently was similar to that for interval 26. The Tapeats pole 181 may well represent a significant part of early Cambrian time. Moreover, this pole differs by only 6° from pole 182, from the middle member of

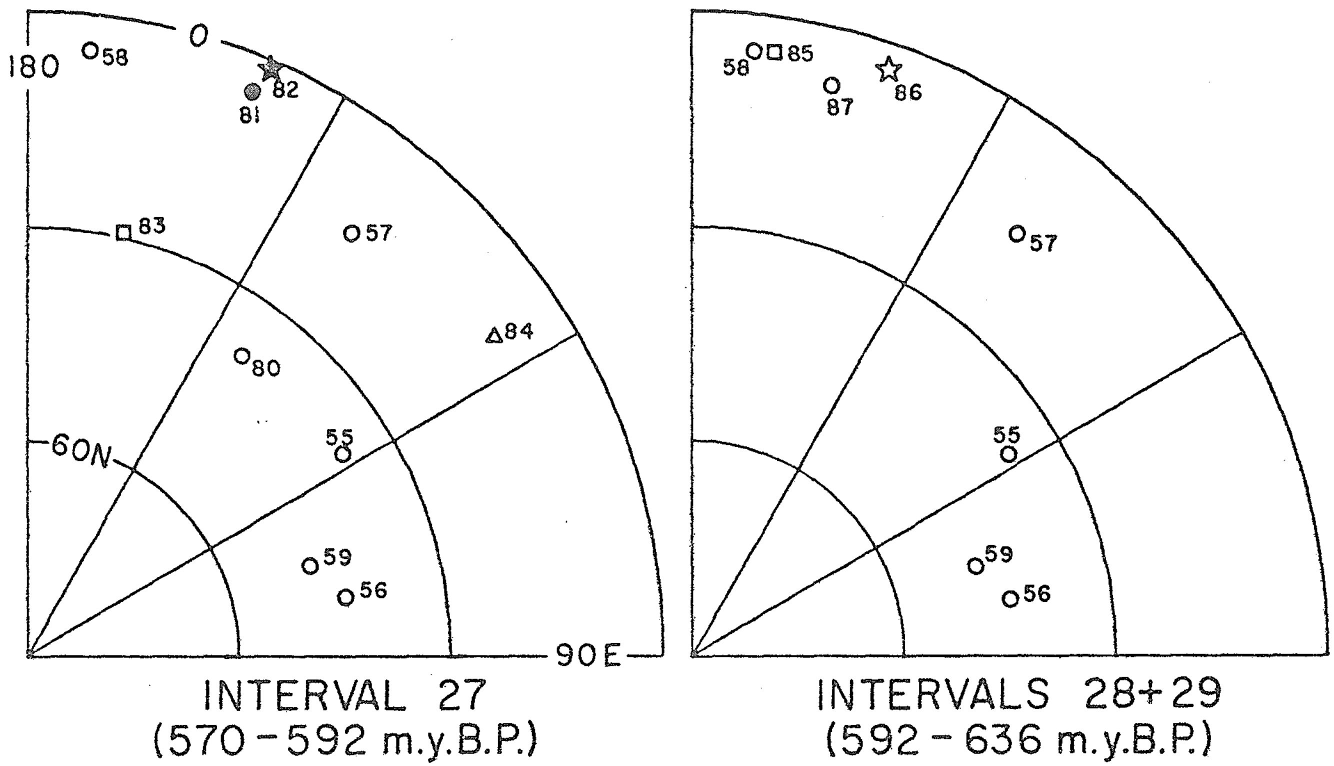


Figure 4.11 North American paleomagnetic poles for intervals 27, 28, and 29 (570 to 636 m.y.B.P.; approx. early Cambrian and latest Precambrian). Symbols and notation are explained in the captions for Figs. 4.2 and 4.7. Stars represent poles in the southern hemisphere.

the Wood Canyon Formation, Nevada. As discussed in Chapter 2, the magnetization of the middle Wood Canyon is probably somewhat older than that represented by pole 181 from the Tapeats.

Other poles from interval 27 might not represent the position of the early Cambrian pole with respect to North America. Pole 183, from the Bradore Formation, western Newfoundland (Rao and Deutsch, 1976), is associated with an ambiguous structural correction and may reflect rotation of western Newfoundland with respect to cratonic North America (cf. interval 23 discussion). Pole 184, from the Ratcliffe Brook Formation, New Brunswick (Black, 1964), diverges by about 50° from pole 181 from the Tapeats. This discordance is consistent with New Brunswick's having been part of a "Maritime block" (Scott, 1975) which drifted independently of cratonic North America during at least part of the Paleozoic. Pole 180, from the Rome Formation, Tennessee (French, 1976), may reflect a vertical-axis rotation of the southern Appalachian Mountains (cf. interval 26 discussion). The ages of the Colorado intrusives studied by Larson and Mutschler (1971) and French et al. (1977) are imprecisely known and may not even be early Cambrian, as suggested by the discordance between poles 155 through 159 from these intrusives and pole 181 from the Tapeats.

The best estimate of the interval 27 pole with respect to North America is taken to be the mean of poles 181 (Tapeats) and 182 (Wood Canyon). The pole for interval 27 (2°N, 157°E) is only about 5° from the mode determined for interval 26. This suggests that the paleogeographic pole was quasi-static with respect to North America for most of Early and part of Middle Cambrian time.

Intervals 28 and 29

In Table 4.1 are listed 7 paleomagnetic investigations of rocks with ages possibly between 592 and 636 m.y. (latest Precambrian). These studies are of igneous rocks from Colorado, Newfoundland, and the Canadian Arctic. From these studies, 8 poles have been derived (Table 4.1, Figure 4.11), all of which are taken from the original references.

Determination of a pole representative of intervals 28 and 29 is hindered by the imprecise ages of poles from the craton. Five of the eight poles listed in Table 4.1 are from Colorado intrusives which may have been emplaced at any time between 704 and 485 m.y.B.P. (French et al., 1977). Two other poles are from flows, dikes, and sills of the Canadian Arctic (Palmer and Hayatsu, 1975; Fahrig and Schwarz, 1973; Robertson and Baragar, 1972; Fahrig et al., 1971). Pole 186 ("Group A," reversed-polarity directions) and pole 187 ("Group B," two-polarity directions) were computed by Palmer and Hayatsu (1975), incorporating results from the earlier studies. As discussed in Chapter 3, an age of 625 m.y.B.P. is suggested for these rocks, based on a whole-rock K-Ar isochron and on the mean of post-1963 K-Ar age determinations.

The age of the Cloud Mountain plateau basalt, represented by pole 185, may be more precisely known, since these lavas are thought to be contemporaneous with petrologically similar dikes dated as 605$\pm$10 m.y. (Deutsch and Rao, 1977a). However, the Cloud Mountain pole 185 is derived from only 3 flows, and hence may not have averaged out secular variation. Moreover, this pole may reflect some rotation of western Newfoundland with respect to cratonic North America (cf. interval 23 discussion).

Tentatively, the pole from the Cloud Mountain basalt (5°N, 172°E) is thought to be the best estimate of interval 28 on the North American APW path. The mean (2°N, 164°E) of poles 186 and 187 from igneous rocks of the Canadian Arctic is provisionally accepted as being the best estimate of the pole for interval 29. The general similarity of poles for intervals 29 through 26 suggests relatively low rates of apparent polar wander with respect to North America from the latest Precambrian into the middle Cambrian. Evidence for a 45° late Precambrian apparent polar shift, probably just prior to interval 29, has been presented in Chapter 3.

Discussion

Poles representing each of the 29 time intervals considered in this study are listed in Table 4.2 and plotted on the APW path of Figure 4.12. For intervals containing at least three unit vectors, both the mode and the mean were determined. Of the 15 intervals for which probability contours were derived (including the three interval 1 subintervals), the difference between the mode and mean was typically less than 3° and in no case greater than 5°. Most of the differences between the mode and mean are attributable to the effects of a single outlier in a small sample of vectors, as in intervals 1A, 4, and 7. In constructing the APW path of Figure 4.12, preference has been given to the mode rather than the mean for intervals in which both were determined. In a few instances, interval poles of Table 4.2 differ slightly from those reported by Van Alstine and de Boer (1978). Generally, this difference is due to the incorporation in the present study of additional data published in 1977 and early 1978; in some cases, a slightly different weighting scheme has been used. The poles reported in Table 4.2 of this study are considered to supersede those listed in Table 1 of Van Alstine and de Boer (1978).

It is difficult to estimate the confidence limits of the interval poles of Table 4.2. Traditionally, points on APW paths are represented by the means of poles for geologic periods. Determinations of these mean poles are usually accompanied by calculations of α_{95}, the half-angle of the cone of 95% confidence for the mean of a Fisherian distribution. In this study, computed α_{95} values would probably be unreliable measures of the precisions of the interval poles for three reasons:

Table 4.2

POINTS ON THE REVISED PHANEROZOIC APW PATH FOR NORTH AMERICA

Interval	Time-span (m.y.B.P.)	Period*	Mid-point (m.y.B.P.)	N	Mode	Mean
1A	0- 3	Q+ℓTP	1.5	11	87N, 50E	89N, 298E
1B	3- 7	mTP+eTP	5	6	85N, 103E	87N, 177E
1C	7- 21	ℓTM-eTM	14	4	85N, 200E	87N, 190E
1	0- 21	Q-eTM	11	20	88N, 74E	89N, 134E
2	21- 42	eTM-ℓTE	32	7	85N, 138E	83N, 150E
3	42- 63	ℓTE-eTPA	53	7	86N, 195E	86N, 181E
4	63- 82	eTPA-ℓK	72	4	66N, 191E	69N, 198E
5	82-103	ℓK-mK	92	3	66N, 187E	67N, 189E
6	103-126	mK-eK	115	2	--	70N, 184E
7	126-147	eK-ℓJ	137	4	66N, 145E	65N, 158E
8	147-169	ℓJ-mJ	158	1	--	68N, 111E
9	169-191	mJ-eJ	180	2	--	66N, 103E
10	191-215	eJ-ℓTr	203	5	60N, 83E	60N, 81E
11	215-239	ℓTr-eTr	227	3	55N, 103E	55N, 103E
12	239-260	eTr-mP	250	2	--	47N, 121E
13	260-281	mP-ℓPN	271	3	50N, 114E	51N, 117E
14	281-302	ℓPN-mPN	292	3	46N, 129E	46N, 129E
15	302-325	mPN-ePN	314	2	--	40N, 132E
16	325-348	ePN-mM	337	1	--	43N, 127E
17	348-371	mM-ℓD	360	2	--	39N, 134E
18	371-394	ℓD-mD	383	5	48N, 121E	49N, 121E
19	394-416	mD-eD	405	1	--	46N, 120E
20	416-438	ℓS-eS	427	2	--	35N, 105E

Table 4.2 - continued

Interval	Time-span (m.y.B.P.)	Period*	Mid-point (m.y.B.P.)	N	Mode	Mean
21	438-460	eS-ℓO	449	1	--	32N, 114E
22	460-482	ℓO-mO	471	1	--	36N, 114E
23	482-504	mO-eO	493		--	(46N, 129E)
24	504-526	eO-ℓ€	515		--	(35N, 128E)
25	526-548	ℓ€-m€	537		--	(21N, 144E)
26	548-570	m€-e€	559	3	6N, 158E	7N, 159E
27	570-592	e€-ℓP€(?)	581	2	--	2N, 157E
28	592-614	ℓP€	603	1	--	(5N, 172E)
29	614-636	ℓP€	625	1	--	(2N, 164E)

Notes:

N is the number of poles given unit weight in determining the mode and/or mean.

Poles in parentheses are based on data that are ambiguous, preliminary, or have poor age-control.

*Based on the revised Geological Society of London Phanerozoic Time-Scale recommended by Lambert (1971). Symbols: Q (Quaternary); TP (Pliocene); TM (Miocene); TO (Oligocene); TE (Eocene); TPA (Paleocene); K (Cretaceous); J (Jurassic); TR (Triassic); P (Permian); PN (Pennsylvanian); M (Mississippian); D (Devonian); S (Silurian); O (Ordovician); € (Cambrian); P€ (Precambrian). Subdivisions of periods are denoted by: e (early); m (middle); ℓ (late).

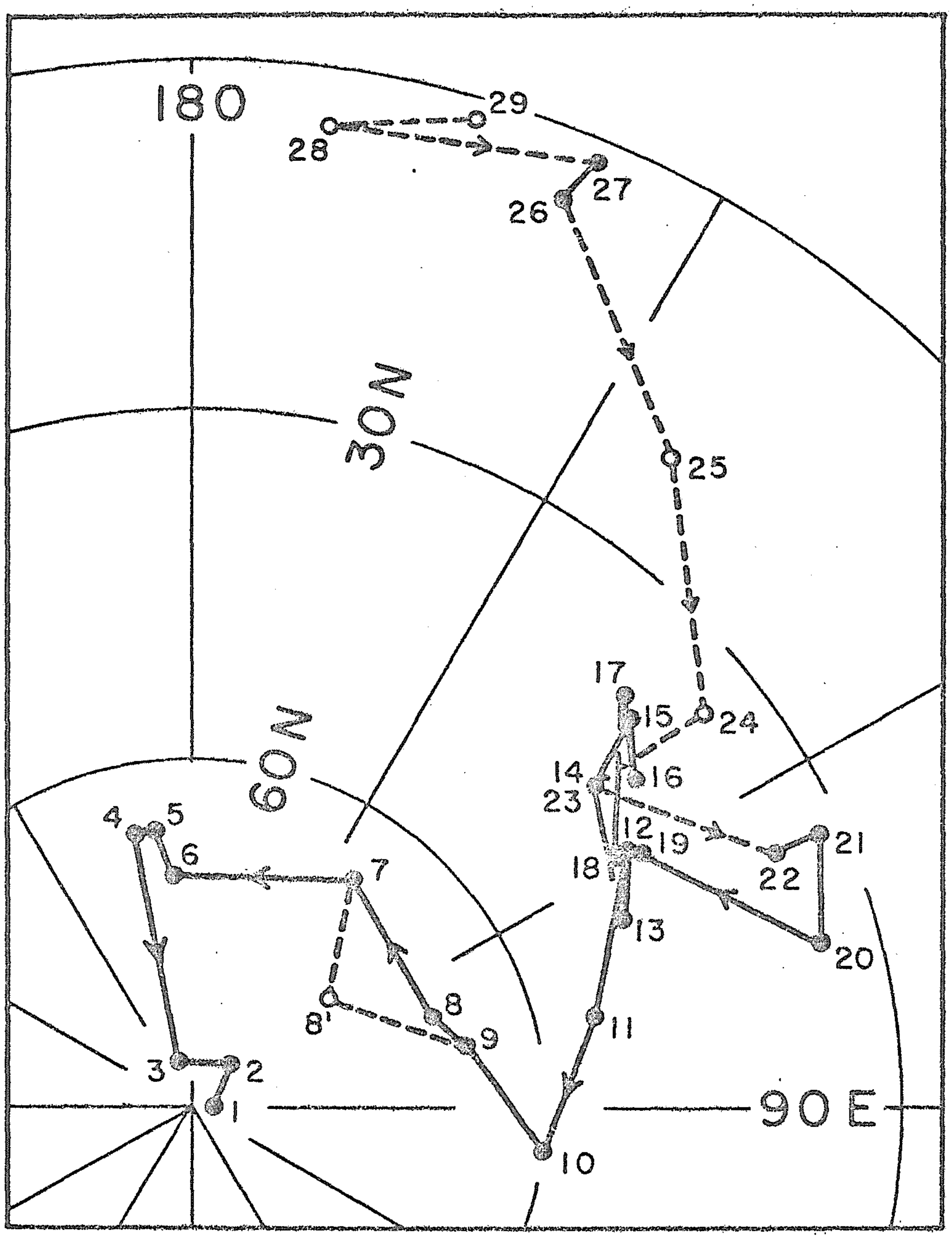


Figure 4.12 The revised Phanerozoic APW path for North America. The points on the path are the interval poles listed in Table 4.2. The mode rather than the mean was plotted for intervals in which both were determined. Segments of the path represented by dashed lines connecting open circles are based on data that are ambiguous, preliminary, or have poor age-control. Azimuthal equidistant projection.

(1) Unless a given interval pole represents a "quasi-static interval," the vector sample from which it is derived will generally not have been drawn from a Fisherian distribution; (2) Even a single deviating pole in these small samples (typically less than 5 unit vectors per interval) can greatly increase the α_{95}, even though the remaining poles may be well grouped; and (3) The distribution in time of published paleomagnetic poles for many intervals is biased toward a small part of the total 22-m.y. time-spans (e.g., intervals 1 and 19). The absence of large jumps in the revised APW path, which has been generated by connecting interval poles based on non-overlapping data sets, suggests that most interval poles are devoid of gross errors. Poles for intervals 8, 23, 24, 25, 28, and 29 are regarded as being especially uncertain, because they are based on data that are ambiguous, preliminary, or have poor age control; hence, they may well change by 10° or more as further data are accumulated.

By constructing an APW path with points representing nearly uniform time intervals, it becomes possible to estimate rates of APW by the spacing between interval poles. Inspection of the path shown in Figure 4.12 suggests that the rate of apparent polar wandering with respect to North America has varied considerably over the past 600 m.y. Low rates of APW (less than 0.3°/m.y.) are indicated for intervals 27 and 26 (early and middle Cambrian), 21 (late Ordovician and early Silurian), 16 and 15 (late Mississippian into early Pennsylvanian), 9 to 8 (middle Jurassic), 5 (middle to late Cretaceous), and 2 and 1 (middle Tertiary to present). In contrast, high APW rates (0.7° to 2.0°/m.y.) are indicated for interval 25 (late Cambrian), parts of intervals 19 and 18 (Devonian), interval 7 (latest Jurassic and earliest Cretaceous), and parts of intervals 4 and 3 (latest Cretaceous and earliest Tertiary).

The revised Phanerozoic APW path for North America (Figure 4.12), which incorporates poles obtained over two decades of paleomagnetic research, exhibits significant changes in both rate and direction of apparent polar wandering. As more data have accumulated, the smooth APW path of Creer et al. (1957) has been superseded by paths that display increasingly more structure (Figure 4.13). Irving and Park (1972) and Irving and McGlynn (1976) have called attention to the "hairpin" bends and "loops" that occur in the North American APW path; they attribute changes in rate and direction of apparent polar wandering to global reorganization of plate motions. Speculation on the significance of the timing and shape of the North American APW path is the subject of Chapter 5.

Some Caveats Concerning APW Path Construction

In revising the North American Phanerozoic APW path, two problems have emerged that may complicate construction of APW paths for all continents: (1) Continental plates may not be as structurally coherent as commonly supposed; and (2) Occasionally, the time-averaged geomagnetic field may deviate markedly from being a geocentric axial dipole.

Most previous APW paths for North America have incorporated paleomagnetic poles from nearly the entire present continent. This procedure can lead to errors of tens of degrees in the APW path if parts of the present North American plate have, at some time in the

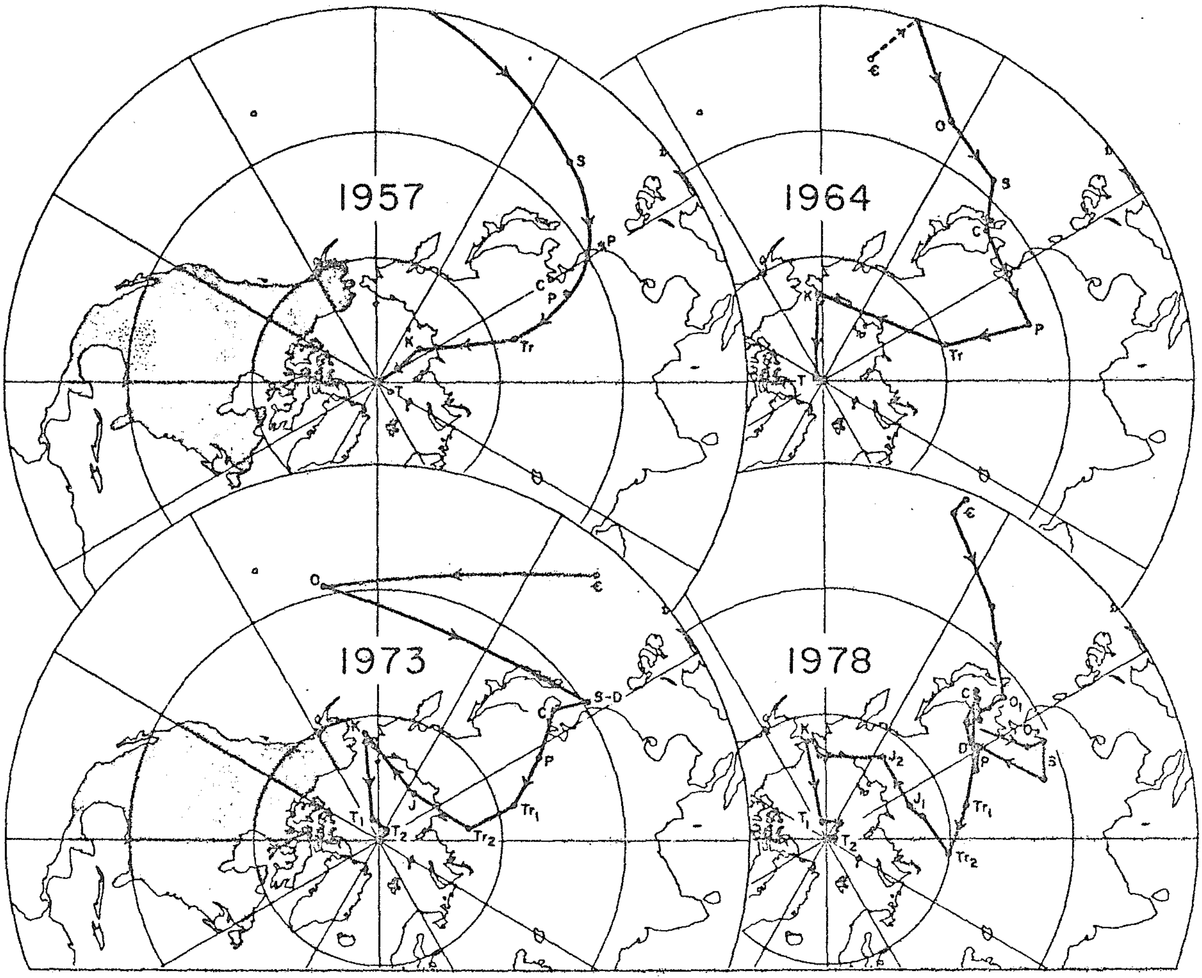


Figure 4.13 Evolution of the Phanerozoic APW path for North America at approximately 7-year intervals. 1957 (Creer <u>et al.</u>, 1957); 1964 (Irving, 1964); 1973 (McElhinny, 1973); 1978 (this study). Symbols: T (Tertiary); K (Cretaceous); J (Jurassic); Tr (Triassic); P (Permian); C (Carboniferous = Miss. + Penn.); D (Devonian); S (Silurian); O (Ordovician); Ꞓ (Cambrian). Subscripts 1 and 2 refer to Early and Late, respectively.

past, drifted or rotated with respect to the cratonic interior. The most spectacular example of this problem occurs in interval 18 (late Devonian) (Figure 4.8), for which poles from the Canadian Maritime provinces and New England deviate by more than 20° from coeval poles from the craton. The consistently lower latitudes of pre-Permian poles from this "Maritime block" almost certainly indicate that this region moved independently of cratonic North America during at least part of the Paleozoic.

Similarly, the question as to whether Newfoundland can be regarded as part of cratonic North America in the Paleozoic has yet to be fully answered. At least the eastern part of Newfoundland was probably not attached to the craton, as evidenced both by the contrasting faunas between the eastern and western parts of the island (Wilson, 1966) and by the discordance of the pole from the early to middle Ordovician Wabana Group. Indeed, it was the incorporation of the Wabana pole in the North American APW path of McElhinny (1973) that caused that path to differ radically from the revised path of this study (Figure 4.13). Moreover, although Deutsch and Rao (1977a) claim that present data do not support the 30° anticlockwise rotation of Newfoundland proposed by Wegener (1929), a possible rotation of up to 15° cannot be dismissed; Paleozoic poles from even western Newfoundland are not particularly concordant with coeval poles from the craton (cf. intervals 16, 17, 19, 23, 24, 25, 27, and 28).

Complex tectonic histories may also be reflected in discordant paleomagnetic poles from much of western North America. Specifically, the Alaska Peninsula (Stone and Packer, 1977), Pacific Northwest

(Watkins and Baksi, 1974; Simpson and Cox, 1977; Basham and Larson, 1978), southern Great Basin (Gillett and Van Alstine, Chapter 2), and in fact the entire western Cordillera (Beck, 1976) seem to have had a peculiar tectonic history involving significant clockwise rotations and, in some cases, northward translations of entire blocks or "microplates." Even the comparatively stable Colorado Plateau may have experienced about 5° of clockwise rotation, since poles from the Plateau for intervals 10 (late Triassic) and 18 (late Devonian) are generally west of those from eastern North America.

As the art of constructing APW paths becomes more sophisticated, the number of poles given unit weight may temporarily decrease, with the increased recognition of local and regional tectonic rotations. Paradoxically, this implies that increasingly more accurate APW paths of the future may be constructed by connecting interval poles that have increasingly large statistical errors; the α_{95} values tabulated for many current APW paths probably give misleading estimates of the accuracy of individual points on the paths. For example, the α_{95} circles associated with the "Carboniferous" and "Silurian-Devonian" points on the North American APW path of McElhinny (1973) do not even include the interval poles for corresponding times on the revised path of this study (largely because the path of McElhinny (1973) incorporates Paleozoic poles from the "Maritime block").

A second problem that may complicate construction of APW paths for all continents involves possible first-order failures of the geocentric-axial-dipole hypothesis for periods of time that may be as long as 10^6 years. That the time-averaged geomagnetic field can deviate at least slightly from that of a geocentric-axial-dipole has been demonstrated most recently by Merrill and McElhinny (1977). They showed that during the past 5 m.y., paleomagnetic poles differ from the geographic pole by an average of about 3°, if calculated by assuming a geocentric axial dipole model of the geomagnetic field. Merrill and McElhinny found that most of the deviation from the geocentric axial dipole results from a consistently negative inclination anomaly. This negative inclination anomaly causes paleomagnetic poles to lie on the "far side" of the geographic pole as viewed from the sampling site. A smaller declination anomaly that yields "right-handed" poles (Wilson, 1972) was dismissed by them as a probable artifact resulting chiefly from an uneven distribution of sampling sites.

Curiously, perhaps the best documented case of suspected failure of the geocentric axial dipole hypothesis in the North American Phanerozoic paleomagnetic record involves large declination anomalies. The "undulatory" directional variations in the Lower Triassic Moenkopi Formation (Helsley and Steiner, 1974) represent predominantly declination swings, with peak-to-peak amplitudes of about 30° and estimated periodicities of about 10^5 to 10^6 years (Helsley, 1977). These long-term directional variations have been reproduced in three continuous stratigraphic sections separated by hundreds of kilometers (cf. interval 11 discussion). If these sedimentary rocks have recorded real directional variations in the geomagnetic field, then it is difficult to imagine that these variations occurred only once in the Phanerozoic, during the Early Triassic. As discussed earlier in this study, the distribution of poles for intervals 24 and 25 (Late Cambrian) does not seem to be explicable in terms of remagnetization, tectonic rotations, or any simple apparent polar wander model. Perhaps the Late Cambrian, like the Early Triassic, was a time of large-amplitude, long-period declination swings.

Since the directional variations in the Moenkopi Formation are primarily declination anomalies, they mimic effects of vertical-axis rotations. Hence, the discordance of some anomalous paleomagnetic poles that have been attributed to tectonic rotations may instead reflect long-term directional variations in the geomagnetic field. It is somewhat disturbing that for some time intervals (e.g., 6 and 8), there are far more "anomalous" poles, for which tectonic rotations have been proposed, than there are reliable poles from the craton.

Future paleomagnetic studies may show that times of first-order departure from the geocentric-axial-dipole hypothesis can be recognized and correlated globally. A detailed discussion of anomalous poles from continents other than North America is beyond the scope of this paper. However, Shaw (1975, 1977) has suggested that the persistence of nearly equatorial pole positions during two Plio-Pleistocene polarity transitions in Iceland and Nevada might be indicative of a third metastable state of the geomagnetic field. In addition, Thomas and Briden (1976) have reported that some Late Ordovician intrusions from Britain yield anomalous paleomagnetic poles that are suspiciously nearly 90° away from other published Ordovician poles from that region; they speculated that these anomalous poles may reflect an anomalous configuration of the geomagnetic field that persisted for several million years during the Early Paleozoic. Similarly, Sallomy and Piper (1973) and Spall (1968) have suggested a possibly anomalous configuration of the geomagnetic field to explain results from the Early Devonian of Britain and the Middle to Late Jurassic of Spitsbergen.

Clearly, postulating a failure of the geocentric-axial-dipole hypothesis to account for an anomalous paleomagnetic pole must be made with considerable caution and only after alternative explanations (e.g., tectonic rotations or remagnetization) have been thoroughly explored. On the other hand, if the time-averaged geomagnetic field deviates grossly from being that of a geocentric axial dipole during say 5% of Phanerozoic time, then failure to recognize this phenomenon will generate segments of APW paths that do not reflect the apparent motion of the geographic pole with respect to the continent.

Conclusions

Representations of the Phanerozoic APW path for North America have evolved from the smooth curve of the late 1950's to one exhibiting significant changes in both direction and rate of apparent polar wandering. Directional changes, or turning points in the path, occur between the middle Ordovician and middle Devonian, between the late Devonian and late Pennsylvanian, between the middle Triassic and early Jurassic, and between the late Jurassic and early Tertiary. Low rates of APW (less than 0.3°/m.y.) seem to be typical of parts of the early and middle Cambrian, late Ordovician and early Silurian, late Mississippian and early Pennsylvanian, early and middle Jurassic, middle to late Cretaceous, and middle Tertiary to present. High rates of APW (0.7° to 2.0°/m.y.) are indicated for the late Cambrian, parts of the

Devonian, latest Jurassic and earliest Cretaceous, and latest Cretaceous and earliest Tertiary.

Phanerozoic paleomagnetic observations suggest that the present continent of North America is a mosaic of structural blocks, many of which have rotated through large angles with respect to the cratonic interior. The distribution of discordant poles from the western Cordillera (from southern California to the Alaska Peninsula) suggests that during the Cenozoic, parts of this region have experienced clockwise rotation and northward translation with respect to the craton. Moreover, pre-Permian poles from the Canadian Maritime provinces and New England states consistently have lower latitudes than poles from time-equivalent formations on the craton; this systematic discrepancy suggests that this region drifted independently of the North American craton during at least part of the Paleozoic. Thus, previous syntheses of the North American APW path incorporating poles from these regions have probably misrepresented apparent polar wandering with respect to the craton.

Some anomalous but apparently reliable poles that lie off the revised path may reflect long-term directional variations in the geomagnetic field. These directional variations, which have been observed in several stratigraphic sections from the craton, may include large declination swings that mimic the effects of vertical-axis rotations. Some paleomagnetic poles that have not averaged at least 10^6 years may deviate by tens of degrees from the paleogeographic pole.

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Chapter 5

QUASI-PERIODIC POLAR WANDERING ON EARTH AND MARS?

In this chapter, I will propose that both Earth and Mars have experienced a highly ordered form of true polar wandering. The evidence for Earth is based on a consideration of paleomagnetic pole positions, viewed chiefly with respect to North America, over the past 1,300 m.y. The evidence for Mars is based on Mariner 9 and Viking 2 images of the Martian polar regions.

I will first show that the revised North American Phanerozoic apparent polar wander (APW) path can be modeled by the superposition of a quasi-periodic (230 ± 35 m.y.) component and a secular component of apparent polar wandering. I will then consider whether this cyclic apparent polar wander model may be applicable to other continents and to Precambrian time. Possible explanations for quasi-periodic apparent polar wandering will be examined and their relative probabilities assessed. From the presently available paleomagnetic evidence, however, one cannot confidently isolate either plate tectonics, true polar wandering, or long-term variations in the geomagnetic field as the dominant cause of apparent polar wander cycles.

I will then show that <u>true</u> polar wandering on Mars, which has been proposed to explain the distribution of peculiar, quasi-circular features near the poles of that planet, seems to exhibit a regularity reminiscent of <u>apparent</u> polar wandering on Earth. This unexpected similarity suggests that quasi-periodic, <u>true</u> polar wandering may occur on Earth and Mars as a consequence of a physical process common to both planets. An oscillatory form of mantle convection is tentatively proposed to be the excitation function of quasi-periodic polar wandering on Earth and Mars.

Sinusoidal Shape and Quasi-Periodicity of the Phanerozoic APW Path for North America

The revised Phanerozoic APW path derived in Chapter 4 and the geologic time-scale employed in its construction are reproduced here in Figures 5.1 and 5.2. Because the path has been determined by connecting points representing nearly uniform time intervals, rates of apparent polar wandering can be estimated by the spacing between successive interval poles. The proximity of paleomagnetic poles from intervals 3 to 1 (63 m.y.B.P. to present) to the present position of the geographic pole suggests that the paleogeographic pole has been relatively stationary, or "quasi-static" (Briden, 1967) with respect to North America since the middle Tertiary. During the latest Cretaceous or earliest Tertiary (between about 75 and 50 m.y.B.P.), an apparent polar shift of 20° seems to have occurred at a rate of at least 0.8°/m.y. Another quasi-static interval persisted throughout most of Cretaceous (intervals 6 to 4) time. The Cretaceous quasi-static interval was preceded by an earlier episode of rapid apparent polar wandering (at least 1.0°/m.y.), occurring primarily during the late Jurassic (interval 7). This late Jurassic apparent polar shift was in a direction about 135° counter to the direction of latest Cretaceous-earliest Tertiary apparent polar motion. Poles for intervals 11 to 8 (239-147 m.y.B.P., or early Triassic to middle Jurassic) define another turning point in the APW path, characterized

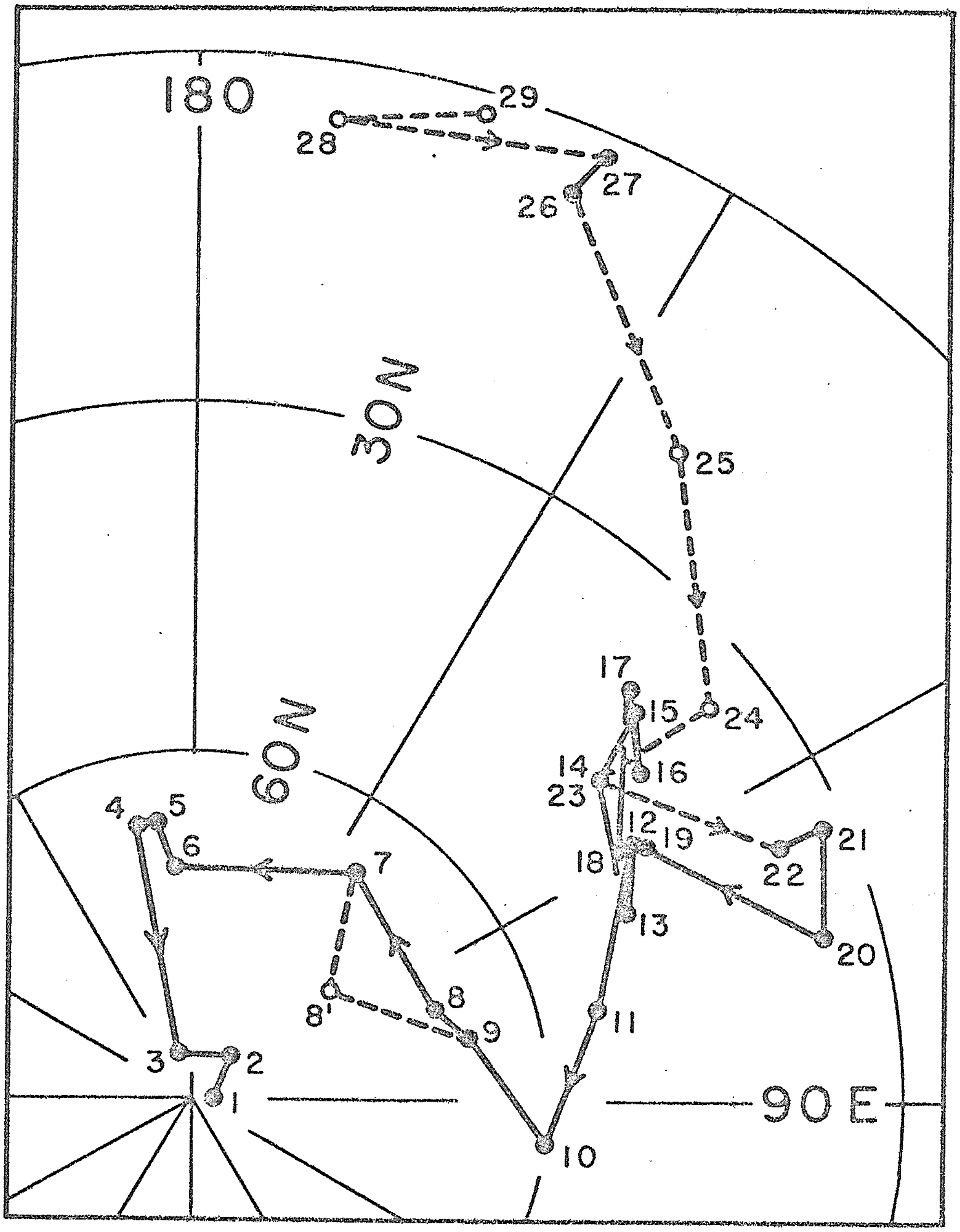
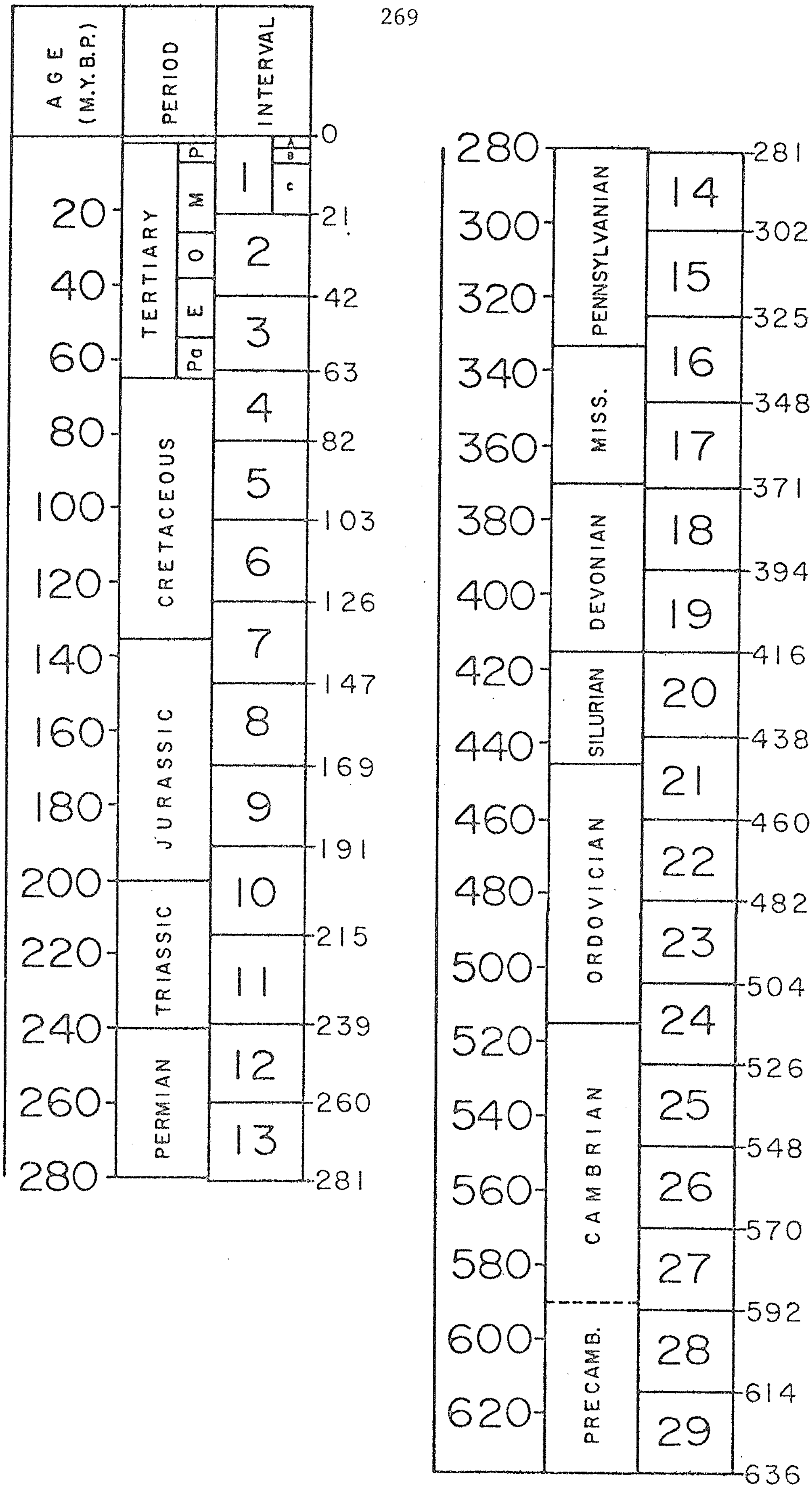


Figure 5.1 The Phanerozoic apparent polar wander path for North America determined in Chapter 4. The points on the path are thought to represent the position of the paleogeographic pole during the approximately 22-m.y. time intervals labeled beside each point. The boundaries of these time intervals are shown in Figure 5.2. Segments of the path represented by dashed lines connecting open circles are based on data that are ambiguous, preliminary, or have poor age control. Azimuthal equidistant projection.

Figure 5.2 Diagram showing the relationships between the 29 time intervals referred to in this study, the geological periods, and corresponding absolute ages according to the revised Geological Society of London Phanerozoic Time-Scale (Lambert, 1971).


AGE (M.Y.B.P.)
PERIOD
INTERVAL
20
40
60
80
100
120
140
160
180
200
220
240
260
280
TERTIARY
P
M
O
E
Pa
CRETACEOUS
JURASSIC
TRIASSIC
PERMIAN
1
A
B
C
2
3
4
5
6
7
8
9
10
11
12
13
0
21
42
63
82
103
126
147
169
191
215
239
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300
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480
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520
540
560
580
600
620
PENNSYLVANIAN
MISS.
DEVONIAN
SILURIAN
ORDOVICIAN
CAMBRIAN
PRECAMB.
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526
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570
592
614
636

by relatively slow apparent polar wandering and a change in direction through an angle of about 130°. Poles for intervals 14 to 10 (325-191 m.y.B.P.) lie on an arc with a length of about 35°, indicating an average apparent polar wandering rate of nearly 0.4°/m.y. between late Pennsylvanian and late Triassic time. During Mississippian and Pennsylvanian intervals 17 to 14 (371-281 m.y.B.P.) the paleogeographic pole was essentially stationary with respect to North America; apparent polar wandering rates averaged less than 0.2°/m.y. The Devonian (intervals 19 and 18; 416-371 m.y.B.P.) was a time of comparatively rapid apparent polar wandering (about 0.7°/m.y.) in a direction about 105° different from that of the late Pennsylvanian to late Triassic shift. Another quasi-static interval occurred in the middle Ordovician to late Silurian (intervals 22 to 20; 482-416 m.y.B.P.), preceded by a rapid late Cambrian polar shift of about 60° at a rate probably in excess of 1.0°/m.y. The rapid shift in the late Cambrian was in a direction about 115° to the Devonian shift. From about 625(?) m.y.B.P. to the middle Cambrian (intervals 29 to 26), the pole does not seem to have moved significantly with respect to North America.

The North American Phanerozoic APW path, therefore, is characterized by 5 episodes of comparatively rapid (0.4-1.0°/m.y.) apparent polar wandering, separated by 6 quasi-static intervals, or times of minimal apparent polar wandering (<0.2°/m.y.). Because each rapid polar shift occurred at an average angle of about 120° to its predecessor, the APW path has a nearly sinusoidal shape. By determining the mid-points of the quasi-static intervals, it will now be shown that the path is quasi-periodic, as well.

The most recent quasi-static interval seems to have begun between 40 and 50 m.y.B.P. Determination of the secular drift in the center of the Chandler wobble (by the International Latitude Service) suggests that the most recent quasi-static interval may be at an end. The astronomical observations seem to indicate that since 1900, true polar wandering has been occurring at the rate of about 1°/m.y. (Rochester, 1973; Dickman, 1977a; Soler and Mueller, 1978). However, the uncertainties in the astronomical determination of this polar motion and the existence of polar-wander excitation processes with a short time-constant (e.g., melting of the Greenland ice-cap (Dickman, 1977b) and sea-level fluctuations (Rochester, 1973)) call for caution in associating the astronomically-determined true polar wandering over the past century with the paleomagnetically-determined apparent polar wandering over tens and hundreds of millions of years. Although it is impossible, therefore, to determine the mid-point of the most recent quasi-static interval, this mid-point must be younger than about 25 m.y.B.P.

Proceeding back in time, estimated mid-points of the other quasi-static intervals with respect to North America are at 93 m.y.B.P. (the middle of interval 5, or approximately middle Cretaceous); 191 m.y.B.P. (the interval 9/10 boundary, or approximately earliest Jurassic); 325 m.y.B.P. (the interval 15/16 boundary, about earliest Pennsylvanian); and 449 m.y.B.P. (the middle of interval 21, or latest Ordovician). The limits of an earlier, latest Precambrian and early Cambrian quasi-static interval are less well defined.

Its beginning is probably at least 605 m.y.B.P. (Cloud Mountain basalt; Deutsch and Rao, 1977) and possibly as old as 625 m.y.B.P. (Franklin K/Ar isochron; Palmer and Hayatsu, 1975). (The evidence for a rapid apparent polar shift just prior to the Precambrian/Cambrian boundary will be discussed in a later section of this chapter.) The end of this quasi-static interval seems to be near the end of the middle Cambrian interval 26. Estimating limits of this quasi-static interval at 615 m.y.B.P. and 548 m.y.B.P. yields a mid-point at 582 m.y.B.P.

In summary, quasi-static intervals followed by abrupt changes in direction of apparent polar wandering seem to occur on the North American Phanerozoic APW path about every 100 to 135 m.y. Unless Paleozoic ages on the revised Geological Society of London Phanerozoic Time-scale are systematically too great by about 15% (rather than the more probable error of ±5%), then the recurrence rate of major changes in Phanerozoic apparent polar wandering with respect to North America is quasi-periodic rather than strictly periodic.

Modeling the North American Phanerozoic APW Path

The nearly sinusoidal shape and quasi-periodicity of the North American Phanerozoic APW path suggest that it can be modeled by superposing secular and oscillatory contributors to apparent polar wandering. Model APW paths were generated in a Cartesian coordinate system by superposing an oscillatory component of specified period, amplitude, and eccentricity on a secular component of specified speed and direction. The Cartesian coordinates of points on these model APW paths were determined by solving the pair of parametric equations:

$$x(t) = R_{min}\cos(\omega t) + V_{sec}t \cos \delta \qquad (5.1)$$

$$y(t) = R_{maj}\sin(\omega t) + V_{sec}t \sin \delta \qquad (5.2)$$

where t represents time, R_{min} and R_{maj} are the lengths of the semi-minor and semi-major axes of the oscillatory component, ω is the angular velocity of the oscillatory component, V_{sec} is the speed of the secular component, and δ is the angle between the direction of propagation of the secular component and the minor axis of the oscillatory component. Eventually, it may become desirable to solve a similar set of equations in a spherical rather than Cartesian coordinate system. However, comparison between a model APW path derived from equations (5.1) and (5.2) versus an actual APW path plotted on a stereographic projection is valid if the actual APW path does not deviate by large angles from the pole of the projection on which it is displayed.

Figures 5.3 to 5.6 illustrate the variety of shapes of model APW paths that can be produced by superposing oscillatory and secular components of apparent polar wandering. Inspection of these figures suggests that, as a first approximation, the North American Phanerozoic APW path can be modeled by superposing an oscillatory component of high eccentricity on a secular component propagating in a direction nearly perpendicular to the major axis of the oscillation (Figure 5.5). If the secular component propagated at any other angle, then the APW path could become quite asymmetric, as shown in Figure 5.6.

Because the 2.5 oscillations on the North American Phanerozoic

Figure 5.3 Model apparent polar wandering (APW) paths for superposed periodic and secular components, generated by solutions to equations (5.1) and (5.2) of the text. The dimensions of the periodic component and the length (per cycle) and orientation of the secular component are shown to scale on the right of the figure. The corresponding APW paths are shown on the left, plotted in a Cartesian coordinate system. Points on the paths are spaced at equal time increments, with 10 points per cycle. The periodic component is circular, and APW paths are shown for a superposed secular component traveling at 0.5, 1.0, 1.5 and 2.0 circle diameters per cycle.

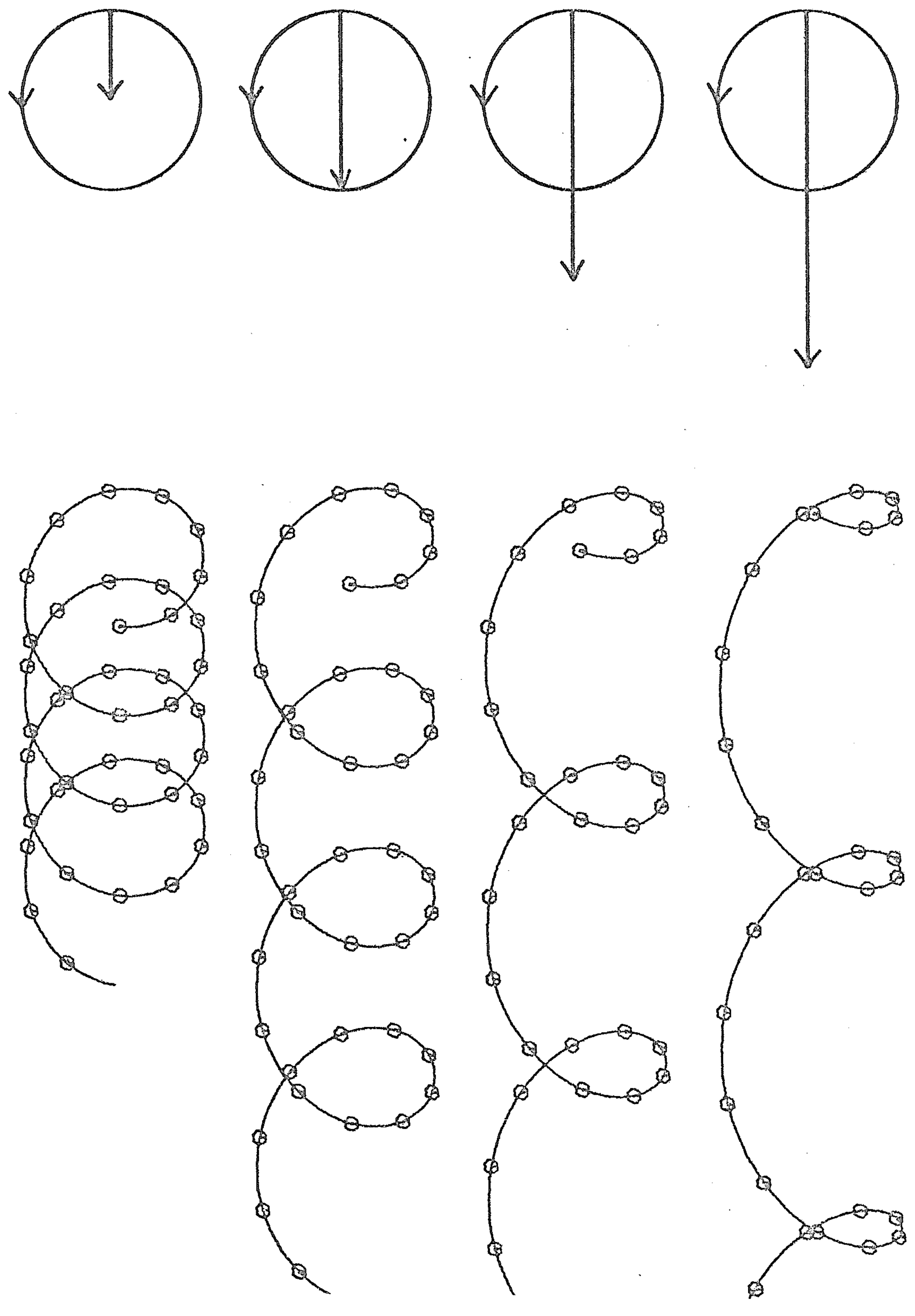


Figure 5.4 Model APW paths as for Figure 5.3, except that here the periodic component is elliptical, with a major axis twice the length of its minor axis. APW paths are shown for a superposed secular component traveling at 0.5, 1.0, 1.5 and 2.0 times the length of the major axis per cycle. The direction of propagation of the secular component is perpendicular to the major axis of the periodic component.

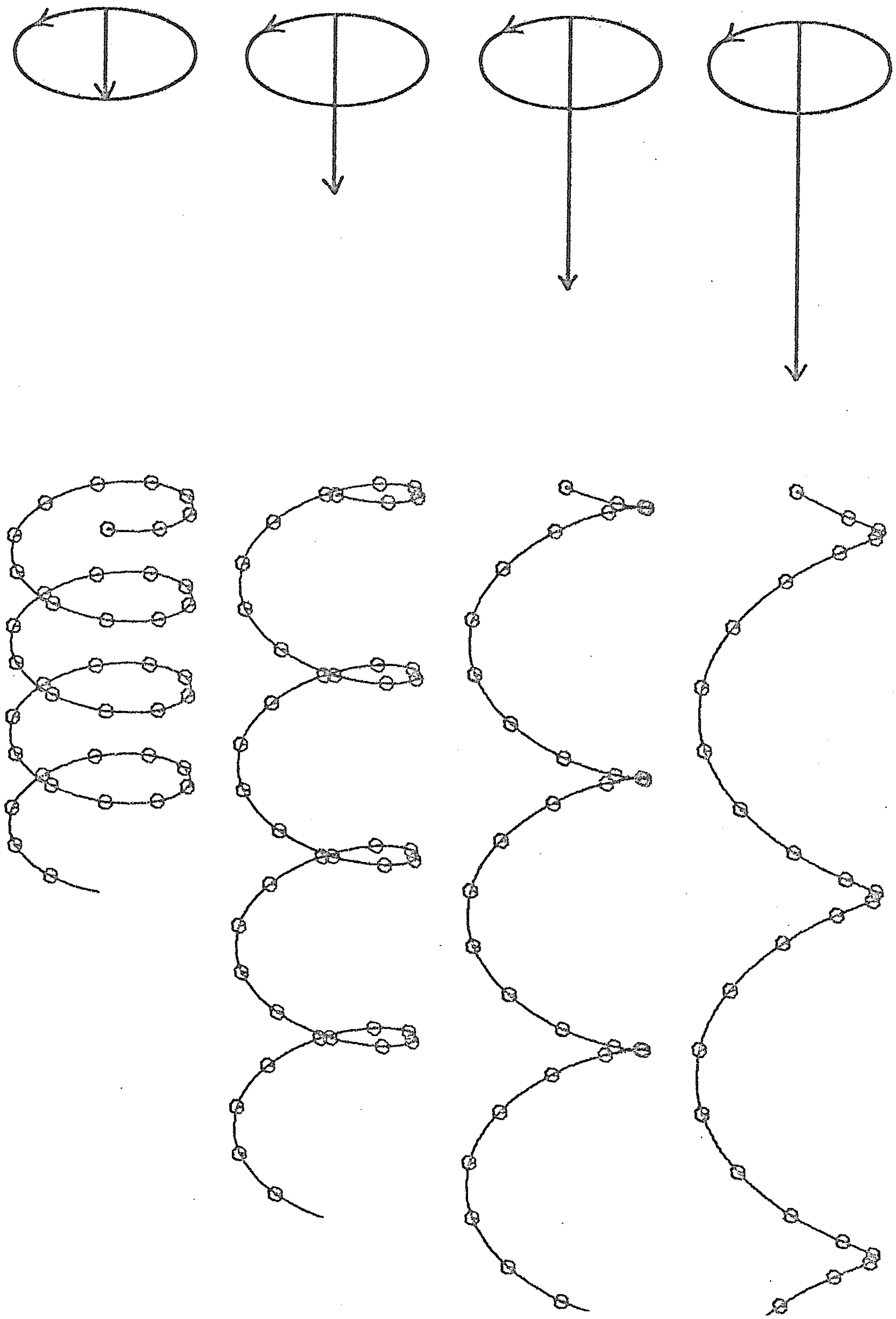


Figure 5.5 Model APW paths as for Figure 5.3, except that here the minor axis of the periodic component is 0, so that the oscillation is only in the ±y direction. APW paths are shown for a superposed secular component traveling at 0.5, 1.0, 1.5 and 2.0 times the length of the major axis per cycle. The direction of propagation of the secular component is perpendicular to the major axis of the periodic component.

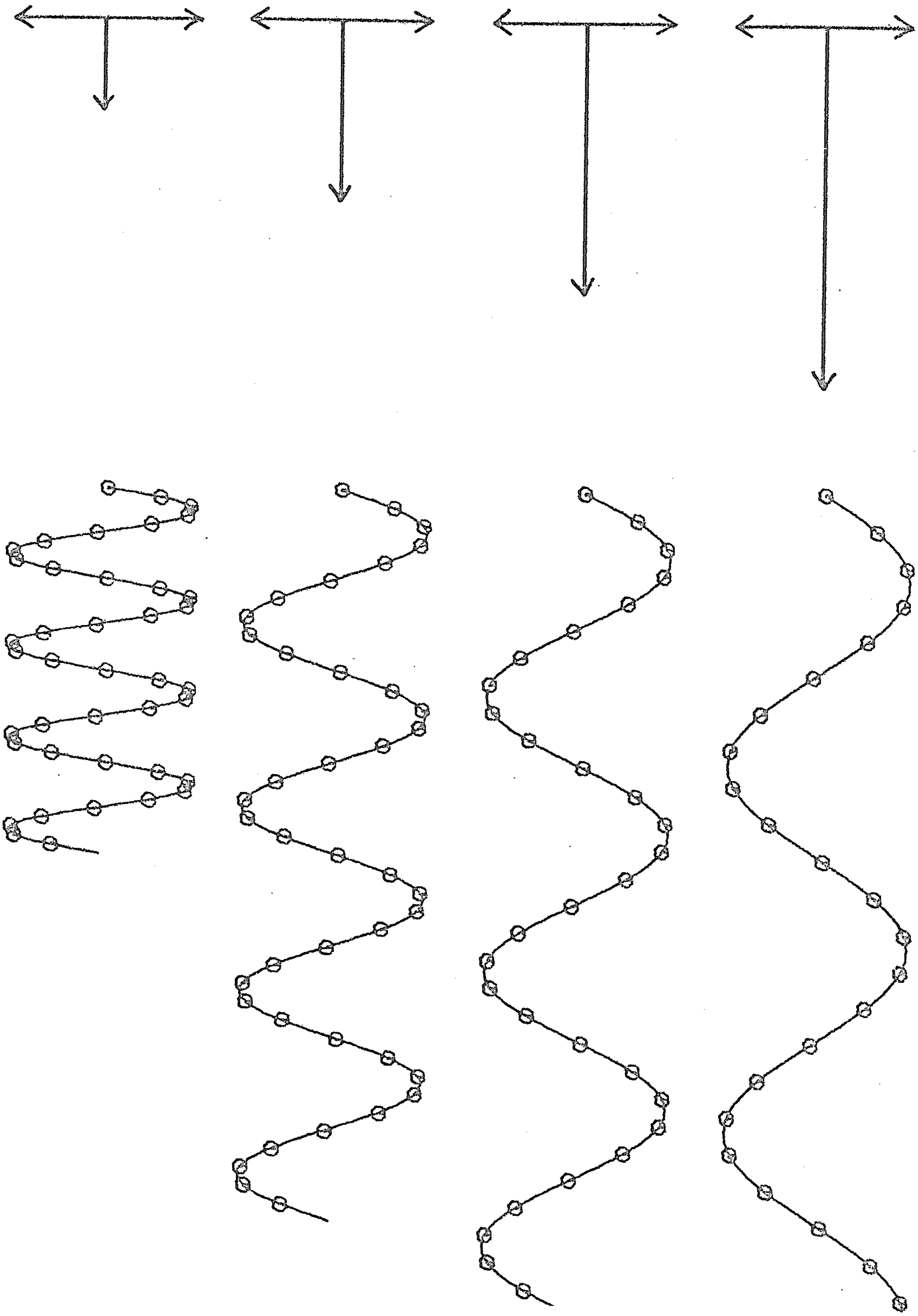
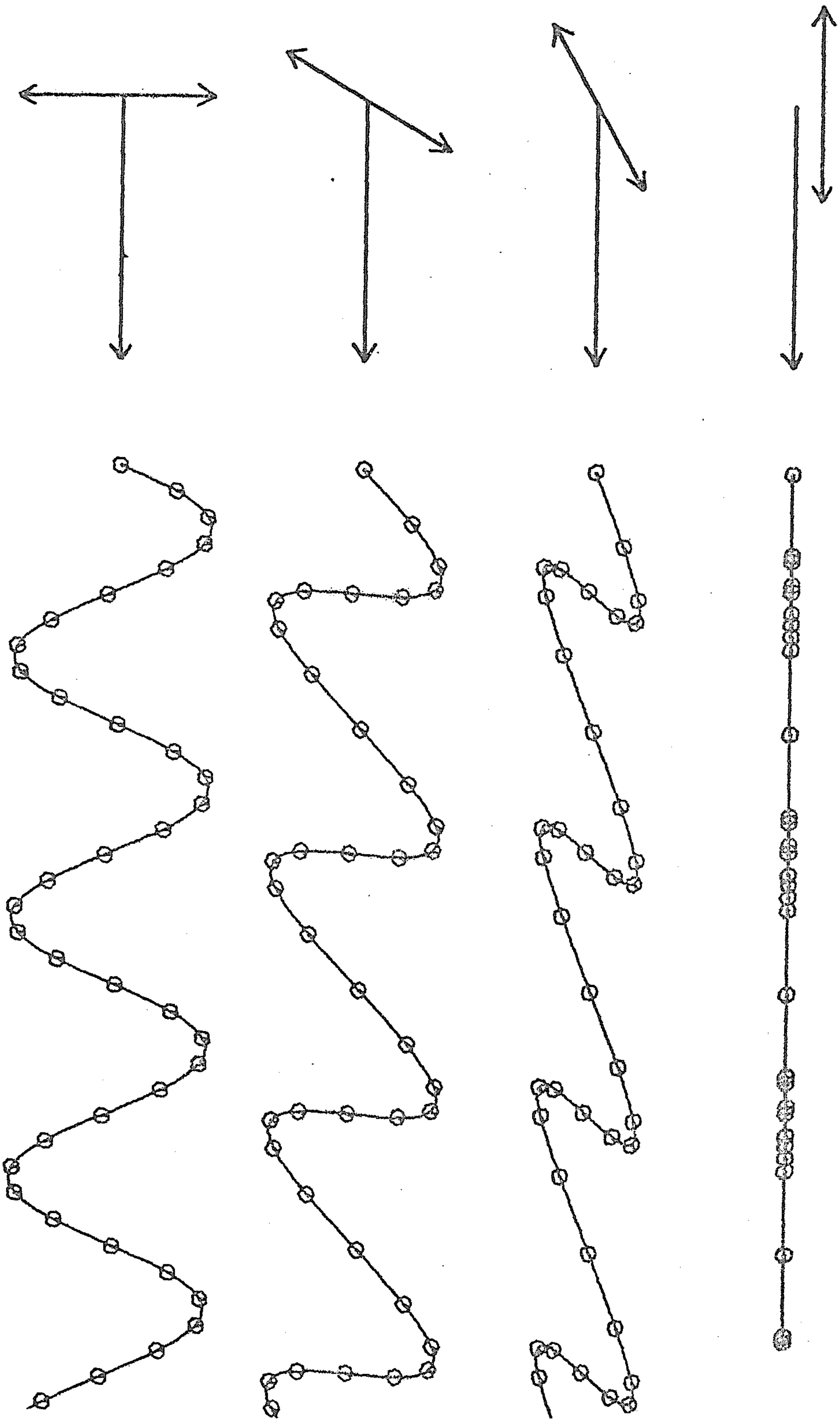


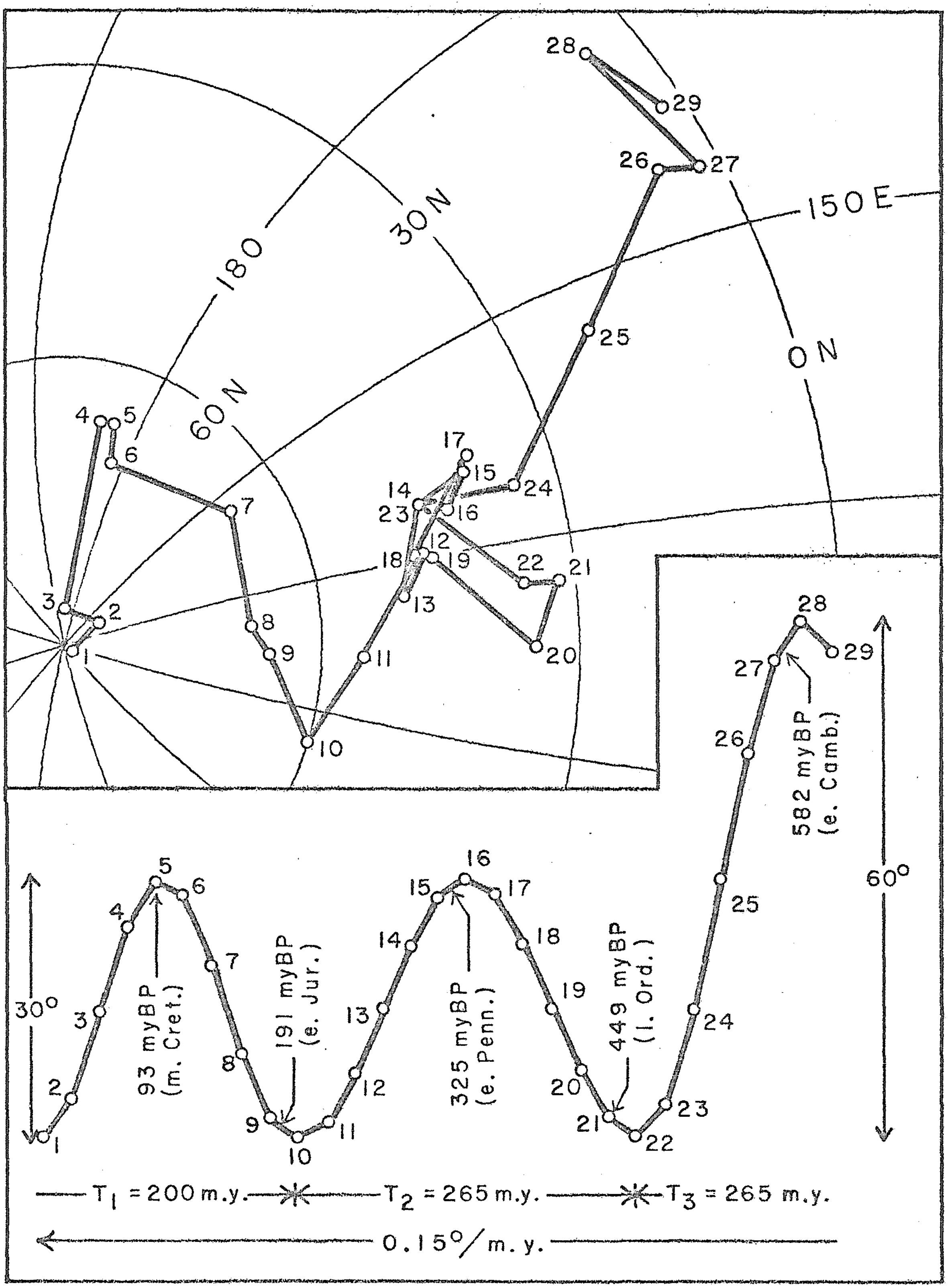
Figure 5.6 Model APW paths as for Figures 5.3 through 5.5, except that here the direction of propagation of the secular component is (from top to bottom) at 90°, 60°, 30° and 0° to the major axis of the periodic component. In all four cases, the secular component travels at a fixed rate of 1.36 times the length of the major axis of the periodic component per cycle.



APW path display neither precise periodicity nor constant amplitude, the path can be modeled more accurately by allowing the period and amplitude to vary from one cycle to the next. The estimated mid-points of quasi-static intervals at 93, 191, 325, 449, and 582 m.y.B.P. imply that a more accurate representation of the North American APW path could be made by modeling with an oscillatory component of period = 265 m.y. for the time 625 to 190 m.y.B.P. and of period = 200 m.y. for the time 190 m.y.B.P. to the present. Moreover, the length of the major axis of the oscillatory component appears to be about 60° from 625 to 450 m.y.B.P. and about 30° thereafter.

Figure 5.7 compares the more accurate computer model with the revised North American Phanerozoic APW path. Differences between the model and the actual APW path can be attributed to a variety of factors. First, points on the actual path may be in error by in some cases up to 10°; APW paths of all continents have been in a state of flux ever since the earliest paths were drawn in the late 1950's (c.f., Figure 4.13). Second, the model path assumes a constant rate of secular drift in a direction perpendicular to the major axis of an extremely eccentric oscillatory component; any variation in these model parameters will cause the actual and model APW paths to differ. Perhaps most important, actual APW paths probably include, in addition to these fairly regular oscillatory and secular components, other components that operate in random directions but with similar amplitudes and rates as the quasi-periodic component. Indeed, in recognition of the probable variety of processes contributing to apparent polar wandering, it is remarkable that

Figure 5.7 Comparison of actual and model Phanerozoic APW paths with respect to North America. The actual path, shown at the top, is plotted on an azimuthal equidistant projection centered at 45°N, 105°E. The model path, shown at the bottom, was generated from solutions to equations (5.1) and (5.2) of the text and is plotted in a Cartesian coordinate system. The period of the oscillatory component is assumed to be 265 m.y. for the time 625 to 190 m.y.B.P. and 200 m.y. for the time 190 m.y.B.P. to the present. These values for the periods are based on estimates of the mid-points of the "quasi-static intervals" (Briden, 1967), which are marked by arrows on the model APW path. The major axis of the oscillatory component is 60° from 625-450 m.y.B.P. and 30° thereafter. The secular component propagates at a constant rate of 0.15°/m.y. in a direction perpendicular to the major axis of the oscillatory component. Points plotted on both the actual and model paths represent nearly uniform, 22-m.y. time intervals, as defined in Figure 5.2.


180
30N
60N
150E
0N
93 myBP (m. Cret.)
191 myBP (e. Jur.)
325 myBP (e. Penn.)
449 myBP (l. Ord.)
582 myBP (e. Camb.)
30°
60°
T1 = 200 m.y.
T2 = 265 m.y.
T3 = 265 m.y.
0.15°/m.y.

any APW path could display the degree of regularity suggested by Figure 5.7.

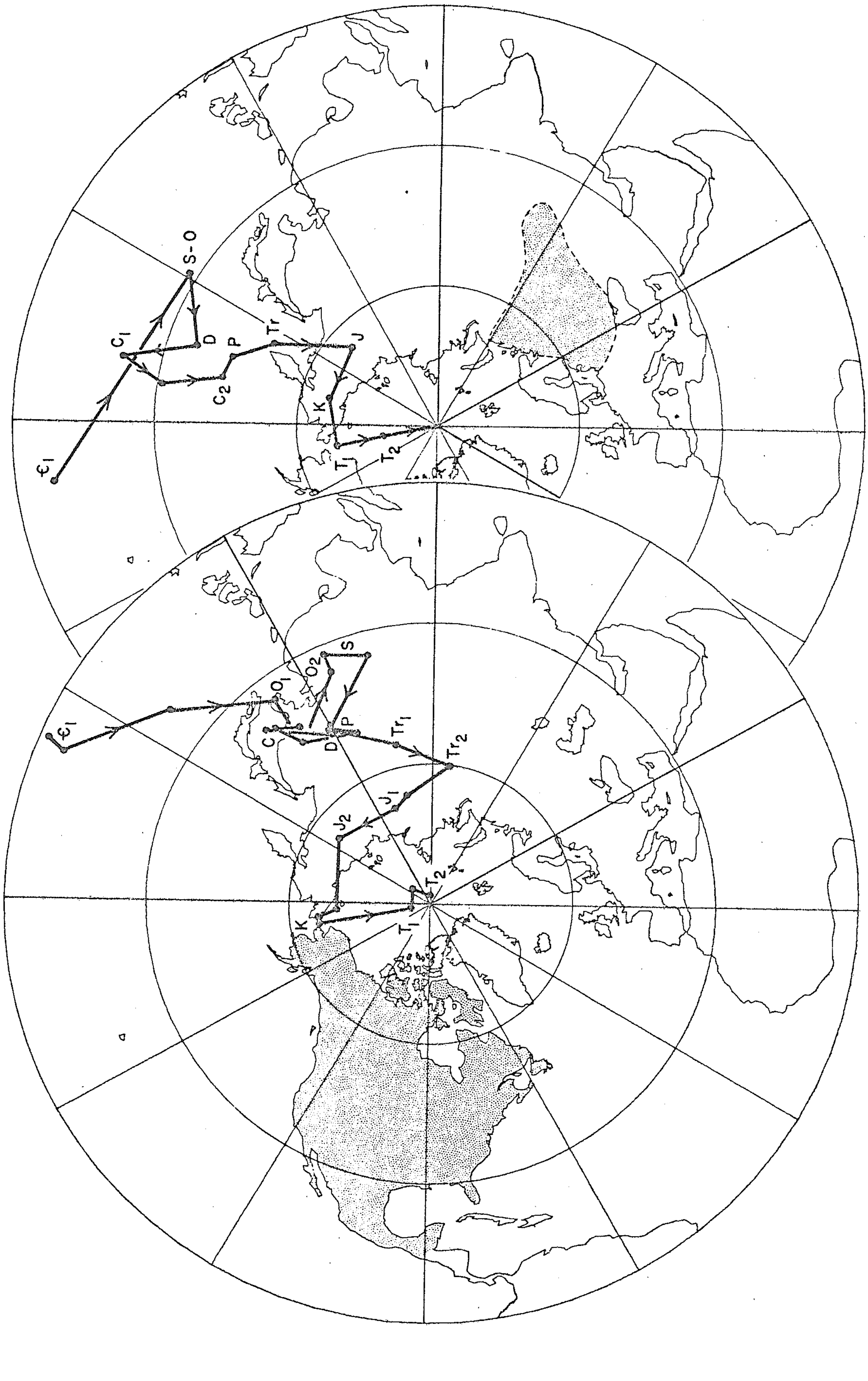
Comparison with Phanerozoic APW Paths for Other Continents

Ideally, in order to determine whether quasi-periodic apparent polar wandering is a global phenomenon, one would like to compare the revised Phanerozoic APW path for North America with APW paths of comparable accuracy from every continent. At present, this is not possible because paleomagnetic poles are not abundant enough for most regions, with adequate time-coverage throughout the Phanerozoic. Even recently determined APW paths have commonly been constructed employing techniques that have smoothed the apparent polar motion and decreased the time resolution. (This problem has been discussed in greater detail in Chapter 4.) Moreover, in revising the North American APW path, Van Alstine (Chapter 4) has noted the following two complications that have probably led to unrecognized distortions in many presently-available paths: (1) lithospheric plates may not be as rigid as commonly supposed; and (2) occasionally, the time-averaged geomagnetic field may deviate significantly from being a geocentric axial dipole. Until these problems are recognized and dealt with on a plate-by-plate basis, all APW paths will, to some degree, misrepresent the true record of apparent polar wandering. Even if presently-available APW paths were extremely accurate, however, the known geophysical process of plate tectonics, which requires differential motions between plates, will act to conceal any possible components of apparent polar wandering that may be common to all the plates.

Despite the complexities that will tend to make the APW path for each plate unique, the Phanerozoic path for the Russian platform (McElhinny, 1973a, p.213) strikingly resembles the revised North American path (Figure 5.8). Abrupt changes in direction of apparent polar wandering with respect to the Russian platform occur in the Cretaceous/early Tertiary, Jurassic, early Carboniferous, and Ordovician/Silurian, corresponding to turning points in the North American APW path. An oscillatory component of apparent polar wandering is also discernible, although less obvious, in the post-Devonian APW path for all of northern Eurasia recently compiled by Irving (1977). However, the Russian platform APW path determined by McElhinny (1973) seems preferable for comparison with the revised North American path because the post-Devonian tectonic integrity of all of northern Eurasia is questionable.

Details of Phanerozoic apparent polar wandering with respect to each of the continents that once formed Gondwanaland are less well known than for North America. Paleomagnetic results from the southern continents are frequently combined into a pre-middle Jurassic APW path thought to be common to all of Gondwanaland (e.g., Irving, 1977; Embleton and Valencio, 1977). Most of the reconstructions of the APW path with respect to Gondwanaland are of comparable length but show broader undulations than the paths from North America and Eurasia. It is certainly possible that this comparative smoothness is a real feature of apparent polar wandering with respect to the southern continents. On the other hand, it may be that as more paleomagnetic data are accumulated, these APW paths may begin to exhibit the

Figure 5.8 Comparison between Phanerozoic APW paths for North America on the left (from Van Alstine, Chapter 4) and for the Russian platform on the right (from McElhinny, 1973a). Symbols: T (Tertiary); K (Cretaceous); J (Jurassic); Tr (Triassic); P (Permian); C (Carboniferous = Penn. + Miss.); D (Devonian); S (Silurian); O (Ordovician); Є (Cambrian). Subscripts 1 and 2 refer to Early and Late, respectively. Azimuthal equidistant projections.


Є$_1$
C_1
S-O
D
P
Tr
J
C_2
K
T_1
T_2
Є$_1$
O_1
O_2
S
C
D
P
Tr_1
Tr_2
J_1
J_2
K
T_1
T_2

apparent regularity of the paths from North America and the Russian platform.

Consideration of Late Precambrian APW Paths

If the approximately 60° polar shift in the North American APW path between the middle Cambrian and early Ordovician represents the younger half of a large-amplitude, high-eccentricity oscillation, then the APW model of Figure 5.7 predicts a similarly large shift in the latest Precambrian (between about 715 and 580 m.y.B.P.). Until recently, however, the North American paleomagnetic data suggested that a quasi-static interval persisted for most of late Precambrian and Cambrian time (e.g., Van der Voo et al., 1976). Partly to test the hypothesis of quasi-periodic apparent polar wandering, a paleomagnetic investigation was undertaken (by the author, in collaboration with S. L. Gillett) of a structurally intact, conformable sequence of uppermost Precambrian to Middle Cambrian miogeoclinal strata in the Desert Range, Nevada. Results of that study, described in Chapters 2 and 3, confirmed at least 45° of apparent polar wandering between about 675(?) m.y.B.P. and the Early Cambrian. The azimuth of this apparent polar shift differs by more than 90° from that predicted by the model APW path of Figure 5.7, indicating either that the model or some of its parameters are in error or that some additional components of apparent polar wandering distorted the path from its hypothetical shape. In support of the model, large-amplitude, latest Precambrian apparent polar shifts also seem to have occurred with respect to the Gondwana continents (McElhinny _et al._, 1974), Siberia (McElhinny, 1973a,

p. 217), and trans-Caledonide Britain (Lomax and Briden, 1977). Rapid polar wandering in the late Precambrian has been suggested by McElhinny et al. (1974) as an explanation for the occurrence of glaciogenic deposits in late Precambrian strata from nearly every continent.

However, all Precambrian APW paths must be viewed with considerable caution because of a variety of complicating factors, including: (1) the difficulty of making stratigraphic correlations without precise isotopic or paleontological age control; (2) the ambiguity of geomagnetic polarity when apparent polar wandering is through angles approaching 90°; (3) the uncertainty of Precambrian plate boundaries; and (4) the increased probability of acquiring secondary components of magnetization in rocks of older age. These problems are compounded by the more limited geographic distribution and stratigraphic coverage of undeformed Precambrian formations. Indeed, Briden (1976) has estimated that while there have been about 25 paleomagnetic studies (worldwide) per 10 m.y. time interval in the Phanerozoic, there have been only about 1 per 10 m.y. interval of the Precambrian between 2,600 and 600 m.y.B.P.

Despite these difficulties, the paleomagnetism of late Precambrian rocks with ages between about 1,000 and 1,300 m.y.B.P. has been particularly well studied and seems to define another large-amplitude APW loop with a period between 200 and 300 m.y. The existence of this loop, which was first suggested by paleomagnetic investigations of igneous rocks from the Canadian Shield near Lake Superior (Dubois, 1962; Spall, 1971; Robertson and Fahrig, 1971), has recently been corroborated by a paleomagnetic study of the Grand Canyon Supergroup

(Elston and Grommé, 1974). The amplitude of the loop is 50° to 75°, depending on whether certain high-latitude poles from the Canadian Shield and Grand Canyon Supergroup represent primary magnetizations (c.f., Irving and McGlynn, 1976, Figure 7). In addition, Piper (1975) has noted that the African APW path for the interval 1,000-1,300 m.y.B.P. also exhibits a loop of comparable size (about 60°), although this interpretation of the African path has been challenged by McElhinny and McWilliams (1977). Thus, the process(es) contributing to quasi-periodic apparent polar wandering in the Phanerozoic may have been operative as long ago as 1,300 m.y.B.P. In addition, there is a suggestion that the amplitudes of the late Precambrian and Early Paleozoic oscillations were about twice as large as the amplitudes of the two most recent oscillations.

In summary, the Precambrian paleomagnetic record from most continents is consistent with a large polar shift between the Precambrian/Cambrian boundary (about 600 m.y.B.P.) and about 750 m.y.B.P. This shift may represent the older portion of a third APW cycle (proceeding back from the present). A fourth cycle, between about 750 and 1,000 m.y.B.P. is ill-defined with respect to North America (c.f., Stewart and Irving, 1974; Irving and McGlynn, 1976), but may account for large-amplitude, late Precambrian apparent polar shifts with respect to Africa (McElhinny and McWilliams, 1977). The APW loop between about 1,000 and 1,300 m.y.B.P., recognized in rocks from the Canadian Shield, Grand Canyon Supergroup, and possibly Africa, may represent a fifth APW cycle. Additional 200-300 m.y. APW cycles may have occurred earlier in the Precambrian, but they do not seem to be resolvable with the currently available paleomagnetic data. However, one must remember that "the state of the art in Precambrian paleomagnetism is not unlike that of 20 years ago with respect to Phanerozoic studies" (McElhinny and McWilliams, 1977) and that the oscillatory component of apparent polar wandering was certainly not discernible in the Phanerozoic APW paths of the late 1950's (Chapter 4, Figure 4.13).

Possible Causes of Quasi-Periodic Apparent Polar Wandering

Thus far I have attempted to present the empirical data base for quasi-periodic apparent polar wandering with a minimum of speculation concerning possible driving mechanisms. I will now consider four processes that may have acted, alone or in concert, to cause quasi-periodic apparent polar wandering: (1) plate tectonics; (2) true polar wandering; (3) coherent movement of a lithospheric shell over the lower mantle; and (4) long-term directional variations in the geomagnetic field.

(1) Plate Tectonics

In assessing the degree to which differential motions of lithospheric plates contribute to APW paths, one must recognize that rates of apparent polar wandering caused by plate tectonics depend critically on the following two geometric relationships: (1) the latitudes of the Eulerian poles of relative rotation between pairs of plates; and (2) the distances between the plates and their corresponding Eulerian poles of rotation. The higher the latitude of the pole of relative rotation, and the greater the distance between this pole and the moving plates, the shorter the arc-length of apparent polar wandering produced by a given rotation about that pole. Suppose, for example, that a supercontinent which was centered on the equator began to split in two along a ridge axis aligned north-south. Then the pole of relative rotation between the two continental fragments would coincide with the earth's spin axis. In this case, an ocean basin of enormous width could be created without producing <u>any</u> apparent polar wandering as viewed from either plate. On the other hand, if the pole of relative rotation between the two continental fragments were on the earth's equator and if this pole were near the center of one of the plates, then even a modest horizontal displacement of that plate would produce a large apparent polar shift with respect to that plate. (The importance of these geometric relationships is illustrated nicely in Figure 126 of McElhinny, 1973a.)

By <u>assuming</u> that plate motions are the sole cause of apparent polar wandering, one can envision scenarios in which plate tectonics could in principle cause quasi-periodic loops in APW paths. Irving and Park (1972), for example, have also noted the tendency of the North American APW path to exhibit frequent "hairpin" bends, although they attributed hairpins to abrupt, global reorganizations of plate motions:

> After a time of free-plate motion a momentary 'plate-jam' occurs, brought about either through a locking of a major convergent juncture by continental lithosphere, or by locking of major transcurrent junctures through the development of marginal irregularities. The forces at depth causing plate motions are still operative and the

> plates soon free themselves, but not without reorganization of their margins and changes in directions of motion.

Irving and McGlynn (1976) illustrated (in their Figure 14) how hairpins may be produced in an APW path by a plate's rotating first clockwise and then counterclockwise about a Eulerian pole situated near the plate and at intermediate or low latitude. In addition, they showed (in their Figure 17) how long-period loops in APW paths might result from operation of the "Wilson cycle" of opening and closing ocean basins:

> . . .the a.p.w. signature of the Wilson cycle may be a pair of complementary loops joined at one end. . .which is the signature of a collision orogeny. Hairpins signify large changes in continental motions, and divide the a.p.w. path into a series of tracks, which spans intervals comparable to the time scale of plate tectonics, namely 10^8 years.

It would seem, however, that the frequency of hairpin bends, especially in the Phanerozoic APW path for North America, is significantly higher than that predicted by this Wilson-cycle explanation. Indeed, nearly a complete APW loop has occurred with respect to North America (and apparently the Russian platform as well) since 180 m.y.B.P., which is approximately when the Atlantic Ocean began opening (Dalrymple et al., 1975). Because the Atlantic Ocean is still opening, the "period" of a complete Wilson cycle (if indeed ocean basin formation and destruction is a cyclic process) would seem to be at least twice as long as the observed period of APW cycles.

A purely plate tectonics explanation for quasi-periodic apparent polar wandering can also be criticized upon consideration of the actual geometries and rates of Cenozoic plate motions. For example, the mechanism proposed by Irving and McGlynn (1976) for producing an APW loop solely by plate tectonics requires both that the Eulerian rotation pole be at relatively low latitude and that it be close to the moving plate. Yet, poles of relative rotation for most of the major plates are presently at high or intermediate latitudes (Minster et al., 1974), and the major spreading ridges are oriented predominantly north-south (Moore, 1973). This has led several authors to suggest that perhaps the directions of plate motions are in some way affected by the earth's rotation (e.g., Moore, 1973; Solomon et al., 1975, 1977; Lutz and Foland, 1978), possibly as a result of tidal bulge stresses (Bostrom, 1976, 1977). In addition, the absolute plate motion model of Minster et al. (1974) indicates that, with respect to a reference frame defined by "hotspots" in the lower mantle (Wilson, 1973, 1965; Morgan, 1971, 1972), predominantly continental plates have moved about 5 times slower than predominantly oceanic plates during the past 10 m.y. The slow rates of absolute motion of continental plates, together with the generally high latitudes of poles of relative rotation, imply that the dominant component of continental APW paths may not be plate tectonics. This might be especially true for a large continent, such as North America, which has been centered at relatively low latitudes (and hence far from the poles of relative rotation) throughout most of the Phanerozoic.

If apparent polar wandering were produced solely by plate tectonics (and if poles of relative rotation remained fairly stationary and did not coincide with the earth's spin axis), then rates of apparent polar wandering would be directly proportional to plate velocities. In particular, quasi-static intervals on APW paths should generally correspond to times of minimum plate velocities. However, just the opposite relationship seems to have existed with respect to North America during the most recent APW cycle. For example, Larson and Pitman (1972) proposed that the inordinate width of the Middle to Late Cretaceous magnetic quiet zone in the Pacific and central and south Atlantic Ocean basins indicated an episode of more rapid sea-floor spreading at that time. (Although Baldwin et al. (1974) suggested that constant plate velocity is still permissible considering the uncertainties in the absolute time-scale for internal subdivisions of Cretaceous time.) An independent argument for rapid plate motions in the Late Cretaceous has been made on the basis of the observed, world-wide marine transgression. As determined quantitatively by Hays and Pitman (1973), increased plate velocities would be accompanied by an elevation of the oceanic ridges, a reduction in volume of the ocean basins, and a displacement of great volumes of water onto the land. Thus, both the width of the magnetic quiet zone and the partly contemporaneous marine transgression suggest a pulse of rapid sea-floor spreading during the Middle and Late Cretaceous (about 110-85 m.y.B.P.). Yet, this time of probably rapid plate motions occurred during a quasi-static interval on the North American APW path (intervals 6, 5 and 4). In addition, the Late Cretaceous transgression may well have marked the maximum elevation of sea-level of the entire Mesozoic and Cenozoic (Bond, 1978). This suggests that the rapid apparent polar shifts in the late Jurassic and the latest Cretaceous/early Tertiary were not associated with particularly vigorous plate motions.

In short, a comparison of the sea-floor spreading, paleomagnetic, and geologic records suggests that quasi-periodic apparent polar wandering may be caused by some process other than plate tectonics.

(2) True Polar Wandering

Another possibly major component of apparent polar wandering is true polar wandering, or a shift of the solid body of the earth with respect to the spin axis. The feasibility of true polar wandering has received much attention in recent years (Gold, 1955; Munk and MacDonald, 1960; MacDonald, 1965; McKenzie, 1966; Kaula, 1967; Goldreich and Toomre, 1969; Fisher, 1974). Much of the debate concerns the extent to which the earth may be stabilized against polar wandering by an equatorial bulge. Goldreich and Toomre (1969) showed that the non-hydrostatic part of the earth's figure is distinctly triaxial. They argued that this triaxial body is "quasi-rigid", because its shape is unaffected by its rotation, but evolves in time due to internal processes. Polar wandering on such a body is then defined as follows:

> The adiabatic invariant in the present case turns out to be the solid angle swept out by the angular momentum vector as viewed from the instantaneous principal axes of inertia. This implies that if the axis of rotation of the quasi-rigid body did once coincide with its major axis of inertia., then. . .it will always continue to do so to high accuracy, regardless of where that principal axis may have shifted relative to the 'geography'.

Goldreich and Toomre emphasized (as did Munk and MacDonald) that the density inhomogeneities contributing to the non-hydrostatic figure of the earth exceeded by more than an order of magnitude the ocean-continent distribution as an excitation function for polar wandering. They regarded the non-hydrostatic earth, then, "as a collection of more or less random density inhomogeneities" (presumably convection cells in the mantle) which "'steer' the rotation axis to maximize the resultant polar moment of inertia".

Based on computer simulations of polar wandering on a quasi-rigid body, Goldreich and Toomre concluded that on the average the pole should wander through 90° in a time slightly longer than the duration of a mantle convection cell (which they estimated to have lifetimes of about 200 m.y.). However, the rate at which polar wandering may proceed depends in part on the viscosity of the lower mantle and can be highly variable. Estimates of the viscosity of the lower mantle range from 10^{22} to 10^{26} poises. As Goldreich and Toomre point out, if this number exceeds about 10^{24} poises, polar wandering would be at best a very sluggish response to the changing axis of maximum inertia. Even if the mantle viscosity were in the lower range and the density inhomogeneities were evolving at a uniform rate, polar wandering could be quite erratic. For example, the simulated polar wandering path shown in Goldreich and Toomre's figure 3 exhibits times of featureless secular drift, episodes of rapid polar shifts through approximately 90° (representing exchanges of the principal and intermediate axes of inertia), and even abrupt changes in direction (hairpin bends). If true polar wandering were a significant component of apparent polar wandering, therefore, at least some hairpins in the observed APW paths may have an origin not directly related to surficial tectonic events (e.g., orogenies or changes in direction of plate motions).

It is a formidable problem, however, to separate the plate-tectonic contribution to APW paths in order to determine whether any true polar wandering has occurred. In fact, a test for true polar wandering is in principle impossible unless (1) the complete history of relative motions of the lithospheric plates is known (by determining poles and rates of relative rotation based on dated magnetic anomalies and on the orientation of transform faults in the ocean basins), or (2) accurate APW paths are available from every lithospheric plate.

Early tests for the occurrence of true polar wandering were made by Irving and Robertson (1969) and McElhinny and Wellman (1969), both with positive results. Their techniques, however, required the special assumption that a particular plate or plate boundary was fixed with respect to the earth's spin axis. McKenzie (1972) proposed a test for true polar wandering requiring no such special assumptions. His test, which was applied by McElhinny (1973b) with a negative result for the early Tertiary (50 m.y.B.P.) to present, requires that reliable APW paths be available for each lithospheric plate. More recently, Jurdy (1974) and Jurdy and Van der Voo (1975) described still another test, requiring a reliable APW path from only one plate, together with the complete relative rotation history of all the plates. These authors concluded that the data are consistent with no more than about 5° of true polar wandering during the past 115 m.y.

Yet, since the middle Cretaceous, the North American APW path exhibits about 25° of apparent polar motion, most of which seems to have occurred prior to about 50 m.y.B.P. Indeed, it is this rapid late Cretaceous to early Tertiary shift that forms the younger half of the most recent loop on the North American APW path. Thus, if the negative result of Jurdy's test is taken literally, then true polar wandering can probably be discounted as the chief cause of quasi-periodic apparent polar wandering.

Despite the mathematical validity of the McKenzie and Jurdy tests for true polar wandering, the geological significance of these tests depends on the completeness and accuracy of the sea-floor spreading and paleomagnetic records. For example, as noted by Jurdy (1978), even seemingly minor changes in our understanding of former plate margins can significantly alter conclusions concerning plate velocities. Moreover, as emphasized in Chapter 4, APW paths are evolving constructs; hence, it is quite likely that major changes (>10°) will occur in some presently-available APW paths, even for post-Jurassic times. Conceivably, when the McKenzie and Jurdy tests are applied using the more accurate APW paths of the future, a statistically significant contribution from true polar wandering may become discernible.

In several respects, it would seem rather surprising if significant true polar wandering had not occurred since the early Cretaceous. The alignment of the "most principal" axis of inertia of the non-hydrostatic figure of the earth with the earth's spin axis suggests that the viscosity of the lower mantle is not high enough to preclude polar wandering (although this argument has been challenged by McKenzie et al. (1974)). Furthermore, determinations of post-glacial rebound of the Canadian shield are consistent with a uniform viscosity of the lower mantle of 10^{22} poises (Cathles, 1975); this viscosity would permit quite vigorous polar wandering. In fact, the computer simulations by Goldreich and Toomre (1969) imply that polar wandering of several tens of degrees could be expected in a time interval as long as 115 m.y. Thus, the apparent lack of polar wandering since the early Cretaceous seems to require some explanation other than prohibitively high viscosity of the lower mantle. Perhaps the past 115 m.y. represents one of the quasi-static intervals on Goldreich and Toomre's simulated polar wandering path for a quasi-rigid body. Alternatively, the life-span of the non-hydrostatic density anomalies might be considerably longer than the roughly 200 m.y. suggested by Goldreich and Toomre.

It appears premature, therefore, to discount true polar wandering as the chief cause of APW cycles. However, the empirical case against the occurrence of polar wandering (at least in the recent past) is sufficiently strong to warrant consideration of other processes that could cause regularities in APW paths.

(3) Coherent Movement of a Lithospheric Shell over the Lower Mantle

Although plate tectonics and true polar wandering are generally thought to be the major contributors to APW paths, one can certainly envision even more exotic components. For example, because the lithosphere may be completely separated from the lower mantle by a low-viscosity layer (the asthenosphere), it might be possible for the lithosphere to slip as a coherent shell over the lower mantle. If density heterogeneities in the lithosphere (e.g., the gravity anomalies associated with subduction zones) were not symmetric with respect to the earth's equator, then the entire lithosphere might be capable of slipping over the lower mantle so that the axis of maximum moment of inertia of the lithospheric shell coincided with the earth's spin axis (H. J. Melosh, personal communication, 1978). The possibility of coherent slippage of a lithospheric shell over the lower mantle has also been considered by Wegener (1929), Munk and MacDonald (1960), Jardetsky (1962), and Soler and Mueller (1978). Even if coherent movement of a lithospheric shell were possible, however, it is difficult to imagine how this process could account for the regularity implied by 250-m.y. APW cycles.

Alternatively, coherent rotation of the lithosphere with respect to the lower mantle could be produced by a mantle "mainstream" (Nelson and Temple, 1972), caused by rotation of the base of the upper mantle at a lightly higher angular velocity than that of the lithosphere. Nelson and Temple proposed that the existence of a mantle mainstream can be inferred from the evolution of island arcs and the asymmetry of fracture zones offsetting segments of spreading ridges. Their proposed mainstream would be viewed at the earth's surface as a westward rotation of the lithosphere with respect to the lower mantle about a nearly polar axis at the rate of 36 cm/year near the equator, or one complete rotation in about 10^8 years. If the axis around which the mantle mainstream flowed coincided exactly with the earth's spin axis, then no apparent polar wandering would be produced. However, if the mainstream flowed around an axis inclined to the spin axis, then the mantle mainstream hypothesis would nicely explain 10^8-year APW cycles. The pole would appear to wander in a circle with a radius equal to the angle at which the rotation pole of the mainstream was inclined to the earth's spin axis.

The mantle mainstream hypothesis, however, is unlikely to explain the oscillatory component of apparent polar wandering as viewed from North America for several reasons. First, the oscillatory component of apparent polar wandering predicted by the mainstream hypothesis would be perfectly circular (as long as the axis of the mantle mainstream remained fixed with respect to the earth's spin axis); however, the observed oscillatory component is highly eccentric. Second, the oscillatory component would be precisely periodic, with a period of about 1 x 10^8 years; however, the observed oscillatory component is only quasi-periodic, with a period in excess of 2 x 10^8 years. Third, the amplitude of the APW cycles would be vanishingly small, since the postulated mainstream rotates essentially about a polar axis (especially if it is influenced by the Moon (Bostrom, 1976)). The axis of the mantle mainstream would have to be inclined to the earth's spin axis at angles between 15-30° to explain the observed 30-60° APW cycles.

Coherent motion of a lithospheric shell could be distinguished from true polar wandering if "hotspots" or "plumes" rising from the lower mantle defined some mesospheric reference frame (Wilson, 1963, 1965; Morgan, 1971, 1972). In the case of no true polar wandering, APW paths and "plume traces" (i.e., the time sequence of volcanic extrusion at the earth's surface) could include two components: one resulting from plate tectonics, and another from coherent motion of the lithospheric shell over the mesospheric reference frame. But if true polar wandering also occurred, then the plume trace would still include only these two components, but the APW path would include a third component. Thus, the differences between plume traces and APW paths could be used as a measure of true polar wandering.

With this motivation, Hargraves and Duncan (1973) and Jurdy (1974) compared paleomagnetic, sea-floor spreading, and plume-trace data for the past 50 m.y. Both studies suggested that during the past 50 m.y.

the lithosphere as a whole has indeed rotated by about 10° with respect to a reference frame defined by the hotspots. Hargraves, Duncan and Jurdy concluded that it was the hotspot reference frame which had actually moved, becasue the paleomagnetic and sea-floor spreading data seemed to indicate that the lithospheric shell had been fixed with respect to the spin axis. In both studies, the axis of rotation of the "mantle roll" was found to be inclined at about 45° to the earth's spin axis (contrary to the prediction of Nelson and Temple's mainstream hypothesis), although the axes of net rotation defined in the two studies differ considerably. However, the accuracy of these determinations of "mantle roll" depends upon the assumption that the mantle hotspots do not move with respect to one another. This assumption may be incorrect, for inter-plume movements at up to 2 cm/year seem to have occurred during the past 40 to 50 m.y. (Molnar and Atwater, 1973; Jurdy, 1974).

Coherent movement of a lithospheric shell over the lower mantle would be indistinguishable paleomagnetically from true polar wandering of a quasi-rigid earth. Therefore, the negative outcomes of the tests for polar wandering by McElhinny (1973b) and by Jurdy and Van der Voo (1975) apply to any mechanism involving rotations of the lithosphere with respect to the lower mantle about a non-polar axis. This would seem to preclude the possibility that coherent rotations of the lithospheric shell may be responsible for cyclic apparent polar wandering. However, the negative outcomes of the tests for polar wandering must be viewed with discretion in light of the uncertainties in the paleomagnetic and sea-floor spreading records.

(4) Long-Term Directional Variations in the Geomagnetic Field

Another hypothetical source of the oscillatory component of apparent polar wandering is a quasi-periodic directional variation in the geomagnetic field. This would require that the time-averaged geomagnetic dipole be grossly non-axially geocentric for periods of nearly 10^8 years. However, the hypothesis that the time-averaged magnetic field can be represented as a geocentric axial dipole is one of the most fundamental postulates of paleomagnetism. Only if this hypothesis is correct can a paleomagnetic pole accurately represent the paleogeographic pole. Evans (1976) has recently presented an elegant test for discriminating between a dipolar geomagnetic field and higher order multipoles over geologic time-scales, based on the world-wide frequency distribution of observed paleoinclinations. He concluded that the geomagnetic field has been essentially dipolar throughout the Phanerozoic.

But has this dipole always been axially geocentric? Certainly, strong geophysical arguments can be made that the geomagnetic field <u>should</u> be symmetrical about the earth's spin axis (e.g., Runcorn, 1954, 1959; Busse, 1975). Comparisons of paleomagnetic pole positions with paleoclimatic indicators (e.g., the distribution of reefs, evaporites, carbonates, or glacial deposits), though not precise, suggest that the time-averaged geomagnetic dipole has indeed been parallel to the earth's spin axis for at least the past several hundred million years (see McElhinny, 1973a for a review). Thus, the geocentric axial dipole model does seem to be a good <u>first-order approximation</u> over geologically long time-scales.

During the past decade, however, detailed analyses (summarized by Merrill and McElhinny, 1977) have suggested that the geocentric-axial-dipole model may suffer some second-order departures. The chief departure, the "far-sided" effect first noted by Wilson and Ade-Hall (1970), can be thought of as resulting from an axial dipole which is offset north of the geocenter. The far-sided effect can also be modeled by the superposition of a geocentric axial dipole and a quadrapole field (Merrill and McElhinny, 1977). For the past 25 m.y., the axial dipole to quadrapole ratio seems to have been large, since the far-sided effect has biased paleomagnetic poles by no more than about 3° over the past 5 m.y. (Merrill and McElhinny, 1977) and perhaps by 5° between 5-25 m.y.B.P. (Wilson and McElhinny, 1974). Hailwood (1977) has presented evidence suggesting that the far-sided effect may have persisted for at least the past 53 m.y. However, this conclusion cannot be made as confidently as for the late Tertiary and Quaternary because of difficulties in correcting for plate motions and possible true polar wandering over this longer time-scale. In addition, Van Alstine (Chapter 4) has noted a few brief intervals, representing perhaps 5% of Phanerozoic time, during which the time-averaged geomagnetic dipole may have deviated from the paleogeographic pole by up to 45°; however, these large deviations, if real, do not seem to have persisted for much more than about 10^6 years on any one occasion. Thus, there is no paleomagnetic evidence for <u>first-order</u> departures from the geocentric axial dipole assumption over 10^8-year time-scales.

Of course, this observation concerning paleomagnetic <u>directional</u>

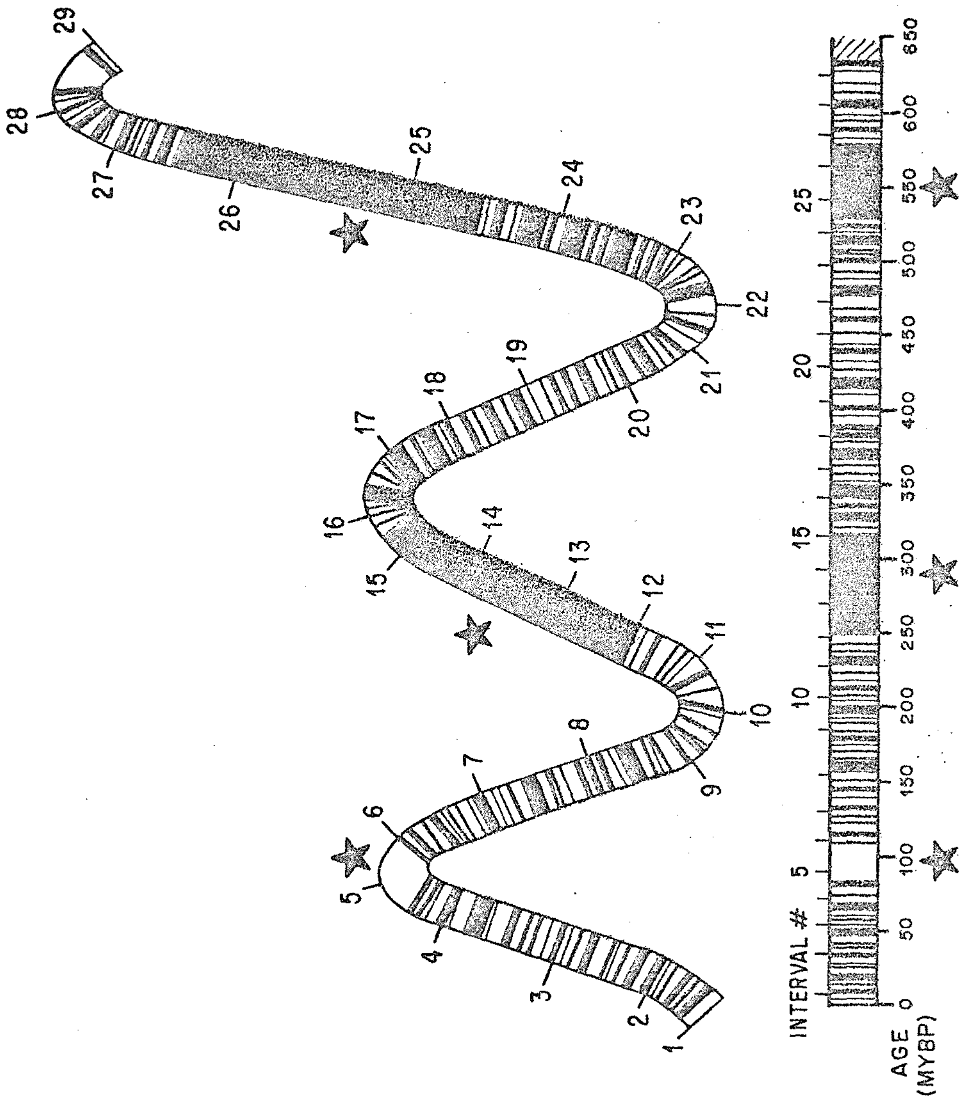
behavior does not preclude the possibility that the polarity history of the geomagnetic field may exhibit long-period variations. Indeed, the Phanerozoic paleomagnetic record does seem to indicate a quasi-periodic variation in geomagnetic reversal frequency. Irving and Pullaiah (1976) have recently performed a maximum entropy spectral analysis on a compilation of world-wide polarity observations throughout the Phanerozoic. They found a dominant period near 300 m.y., although this "period" essentially reflects the length of time between the Cretaceous long normal interval and the Pennsylvanian-Permian long reversed interval.

In testing the hypothesis of quasi-periodic apparent polar wandering, this author and colleagues have obtained paleomagnetic directions from over 3,000 oriented samples ranging in age between late Precambrian (about 700 m.y.B.P.) and late Devonian (about 380 m.y. B.P.). Results of these studies (currently in preparation for publication), as well as recent paleomagnetic studies by other investigators of Early Paleozoic rocks, permit a refinement and tentative extension of the polarity record analyzed by Irving and Pullaiah (1976). The more complete record suggests frequent polarity reversals during the late Precambrian and early Cambrian; a strong reversed polarity bias during perhaps all of middle Cambrian and most of late Cambrian time; frequent reversals during the middle Ordovician through early Devonian; and perhaps an increasingly strong reversed bias prior to the beginning of the Pennsylvanian-Permian long reversed (Kiaman) interval.

A sketch based on current knowledge of the polarity history since the latest Precambrian has been plotted on the model North American APW path in Figure 5.9. (The time-scale used in this figure is the revised Geological Society of London Phanerozoic Time-scale recommended by Lambert (1971); this differs slightly in the Paleozoic from that used by Irving and Pullaiah (1976).) Inspection of Figure 5.9 suggests that there may be only one strong polarity bias interval per APW cycle. Moreover, strong polarity bias intervals do not seem to occur at the same point on successive APW cycles. Thus, there is no reason to suppose that long-term variations in the geomagnetic field directly cause oscillatory apparent polar wandering (e.g., by periodic changes in the axial dipole to quadrapole ratio). However, the similarity of the "periods" (a few hundred million years) of strong polarity bias and oscillatory apparent polar wandering does hint at a common origin.

To recapitulate, it has been demonstrated that the Phanerozoic APW path of at least one major continent can be approximated by the superposition of an oscillatory and a secular component of apparent polar wandering. Consideration of the geometry and rates of recent plate motions suggests that this path may not be dominated by plate-tectonic effects. The major competing mechanism may be true polar wandering, excited by evolving density inhomogeneities in the mantle. A test designed to determine the extent of true polar wandering gives a negative result for the time during which one-half of the most recent APW cycle occurred; however, the geophysical significance of this

Figure 5.9 Generalized sketch of current knowledge of the geomagnetic polarity zonation for the Phanerozoic plotted on the model North American APW path (above) and on the absolute time-scale (Lambert, 1971) (below). Black represents reversed and white represents normal polarity. The stars mark the approximate mid-points of times of strong polarity bias. References for post-Early Paleozoic polarity determinations are given in Irving and Pullaiah (1976), supplemented in the Jurassic by Steiner (1978). Zonation of the Early Paleozoic is based on references in Irving and Pullaiah (1976), supplemented with post-1976 investigations, including unpublished determinations by Van Alstine and colleagues.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test depends on the accuracy of APW paths, which the history of paleomagnetism has shown to be evolving constructs. To complicate matters even more, the geomagnetic field, the same phenomenon that permits construction of APW paths, seems to exhibit a long-term variation with roughly the same frequency as the oscillatory component of apparent polar wandering. Clearly, with only this empirical data base, any attempt to attribute the regularity of apparent polar wandering to a specific geophysical process must be highly controversial.

I will now demonstrate that a polar wandering path for Mars, inferred from the distribution of sedimentary deposits near the north pole of that planet, shares a common trait with the North American APW path: <u>both</u> paths can be modeled by superposed periodic and secular components. This observation may be highly significant in the search for a mechanism for terrestrial APW cycles, because Mars has neither plate tectonics, a strong magnetic field, nor a large moon. Yet, both Earth and Mars seem to exhibit mantle convection, which would be associated with evolving density inhomogeneities that could excite true polar wandering.

Quasi-Periodic Polar Wandering on Mars?

At latitudes on Mars higher than about 75°, the youngest stratigraphic unit (other than the annual and perennial ice) is the distinctive "layered terrain". The nature and origin of the layered terrain have been described and interpreted by Soderblom *et al.* (1973), Cutts (1973a, b); Dzurisin and Blasius (1975), and Cutts *et al.* (1976). The stacked layers, which may have a cumulative thickness of 4-6 km at the north pole (Dzurisin and Blasius, 1975), are thought to be eolian deposits of volatile-cemented dust. During the Martian global storms, dust is transported to the poles, where it settles and is evidently trapped as a result of cementation by volatiles (H_2O and/or CO_2 ice). These deposits form layers, consistently several tens of meters thick, with a regularity thought to be controlled by astronomically induced climatic change over tens of thousands of years (Murray et al., 1972). An even longer-period astronomical effect was suggested by Murray et al. (1972) to explain the apparent grouping of several tens of these layers into overlapping "plates".

The discovery of long-period variations in the eccentricity and obliquity of Mars' orbit by Ward (1973), Murray et al. (1973), Ward (1974), and Ward et al. (1974) provides a mechanism for the periodic climatic fluctuations seemingly implied by the segregation of polar sedimentary deposits into layers and plates. As emphasized by Ward et al. (1974), the obliquity variations probably cause the greatest fluctuations in the Martian climate, since the total annual insolation at the poles is proportional to the first power of the obliquity, but to second order on the eccentricity. By assuming that the atmosphere is in equilibrium with a perennial CO_2-ice reservoir at the north pole, Ward et al. (1974) showed that the obliquity variations would cause cyclic atmospheric pressure variations over nearly two orders of magnitude. These variations could strongly influence the extent of the polar caps as well as the rates of eolian erosion.

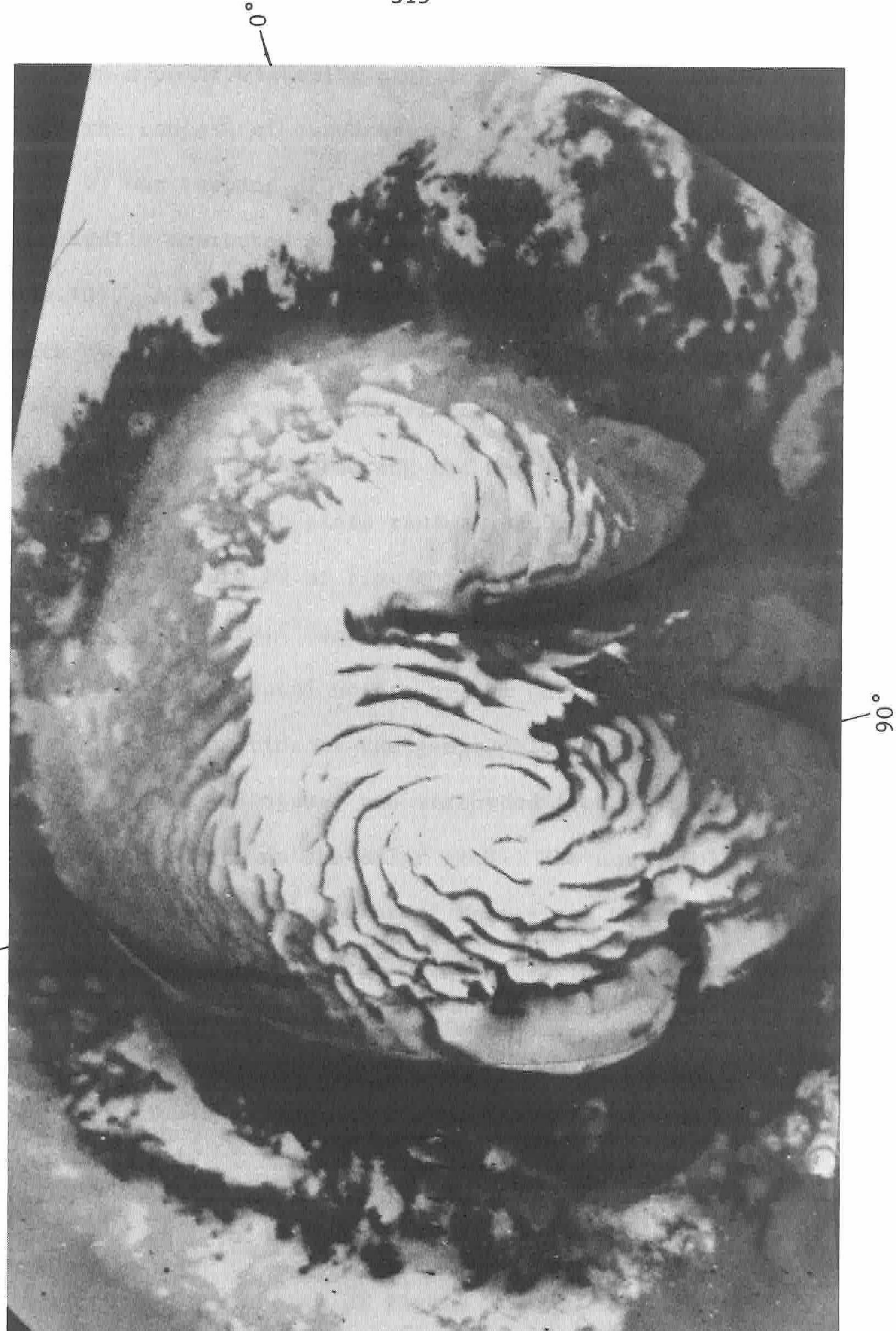
The obliquity variations have two periods:

1.2×10^5 years and 1.2×10^6 years. If one assumes that the shorter-period obliquity variations are associated with formation of the thinnest layers of the layered terrain, then each plate of 20 to 40 layers (Murray <u>et al</u>., 1972) would represent a few million years of deposition. There seem to be about 30 to 40 plates in the north polar region, implying deposition over a time of about 10^8 years (Murray and Malin, 1973a). Obviously, this estimate of the time represented by the layered terrain is only an order-of-magnitude approximation, since it depends upon assumed mechanisms of formation of the layers. However, Cutts (1973a) has independently estimated that the layers were deposited over a 500-m.y. time-span, on the basis of the inferred deposition rate of dust at the pole. Thus, a conservative estimate of the time represented by the polar layered terrain would be 50-500 m.y.

In plan view, the most remarkable features of the polar layered terrain are dark, quasi-circular bands (Figure 5.10) which are thought to be defrosted slopes facing the equator (Murray <u>et al</u>., 1972). It is these dark bands which delineate the boundaries of individual "plates" of polar layered deposits. Murray and Malin (1973a) proposed that these dark bands might represent edges of nearly circular caps that had formed symmetrically about the pole; the dark bands would then in effect be "fossil latitudinal circles", and the offsets in their centers of curvature from the present geographic pole might provide evidence for polar wandering on Mars during the past 10^8 years.

If the quasi-circular bands do represent fossil latitudinal

Figure 5.10 Mariner 9 stereographic photomosaic (from Dzurisin and Blasius, 1975) of the residual frost cap near the north pole of Mars. The offset in the centers of curvature of the dark, quasi-circular features from the present position of the pole suggests that polar wandering may have occurred on Mars during the past 10^8 years.


0°
90°
180°
270°

circles, then a polar wandering path for Mars can be obtained by connecting the centers of curvature of successive plate edges. Figure 5.11 shows my own mapping of the most prominent dark bands from an orthographically corrected Mariner 9 photomosaic of the north pole (Figure 5.10). A polar wandering path was derived by fitting (by eye) these dark bands to portions of circles of nearly constant radius and then connecting the centers of curvature in sequence. Figure 5.12 shows one attempt at reconstructing the Martian north polar wandering path, in which the average plate radius was 3.4°, with a range between 3.1° and 3.6°. Feature 29 of Figure 5.12 is the youngest band that displays radial symmetry. Features 30 to 32, although somewhat distorted, define additional points on the polar path between point 29 and the present position of the geographic pole. Plate edges younger than feature 32 appear too distorted from radial symmetry to provide accurate points on the polar path. The north polar wandering path of Figure 5.12 resembles that of Murray and Malin (1973a), except that this rendition of the path is continuous, whereas Murray and Malin's contains several breaks.

An interesting characteristic of the Martian north polar wandering path is its tendency to exhibit two major changes in direction. These two bends are reminiscent of the "hairpins" in the North American APW path. Moreover, the similar shape of the polar wandering path of Figure 5.12 to the model path of Figure 5.4 suggests that the Martian north polar wandering path can be modeled by a superposed periodic and secular component. Figure 5.13 shows that the Martian polar wandering path can indeed be modeled quite well by a periodic,

Figure 5.11 Map of the most prominent quasi-circular bands of the north polar layered terrain, traced from the photomosaic of Figure 5.10. The cross marks the present position of the geographic pole.

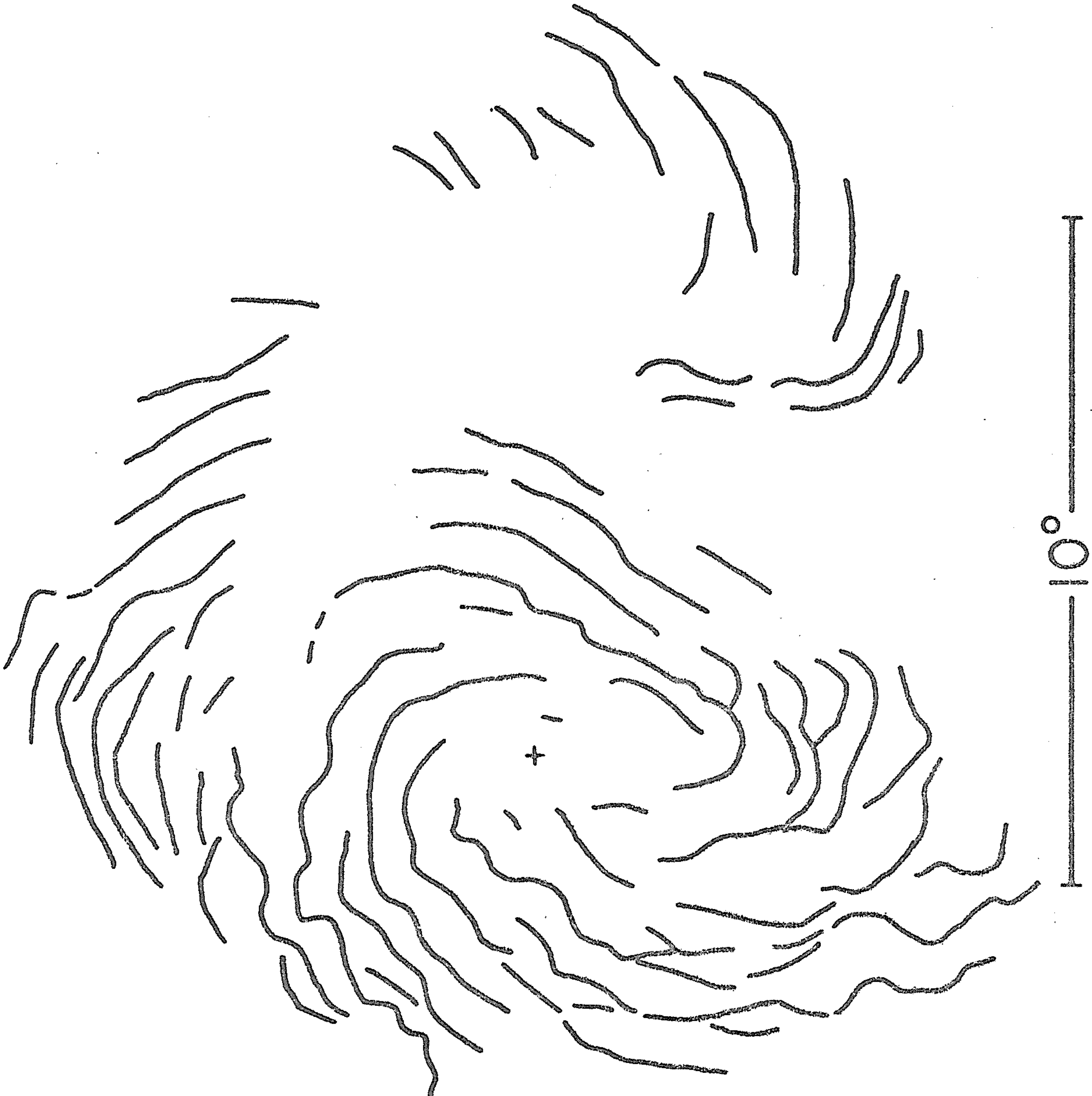


Figure 5.12 Determination of the north polar wandering path of Mars by fitting the most prominent quasi-circular features (shaded) to portions of circles of nearly constant radius (about 3.4°) and then connecting the successive centers of curvature. The numbers relate points on the polar path to the corresponding dark bands and show the inferred sequence of deposition of the polar "plates". Dark bands corresponding to points on the polar path between point 32 and the present north pole (cross) are greatly distorted from radial symmetry and hence are not shown.

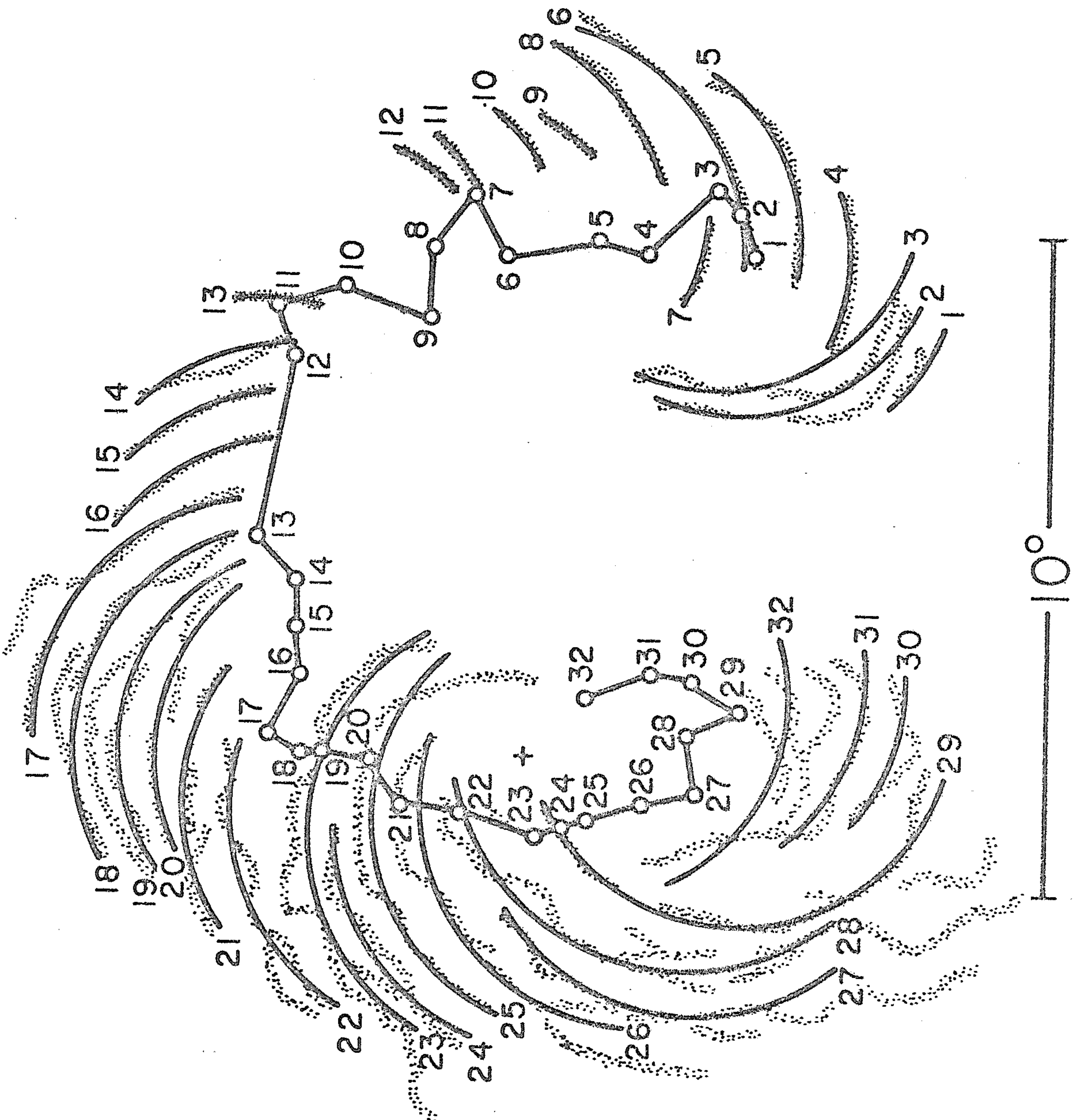

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Figure 5.13 Comparison between the observed (open circles and squares) and predicted (solid circles on curve) centers of plates of layered deposits. The open circles are points from the polar wander path derived in Figure 5.12. The open squares represent points from a duplicate determination of the polar path and demonstrate the sensitivity of the path to slightly different estimates of plate diameters and centers of curvature. The model curve was generated from solutions to equations (5.1) and (5.2) of the text for superposed periodic and secular components of polar wandering. The periodic component is elliptical, with a major axis of 8.0° and a minor axis of 5.3°. The secular component travels 8.0° per cycle in a direction perpendicular to the major axis of the elliptical component. The present position of the north pole is marked by the cross.

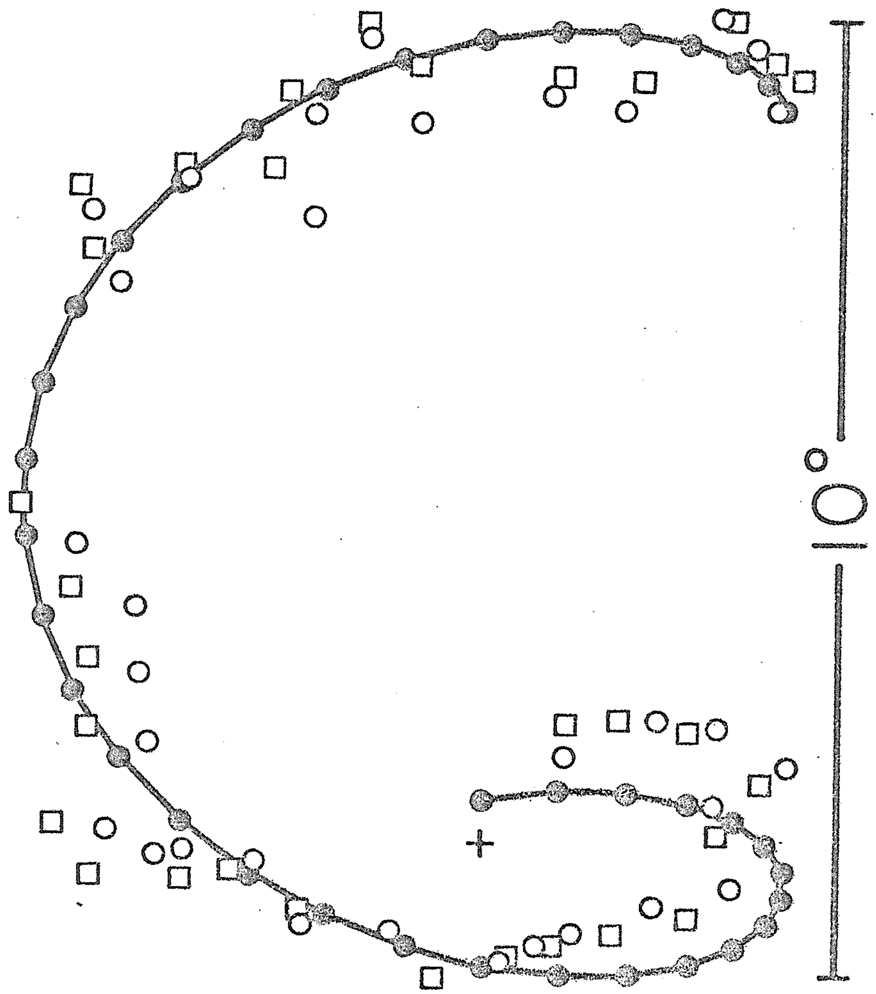
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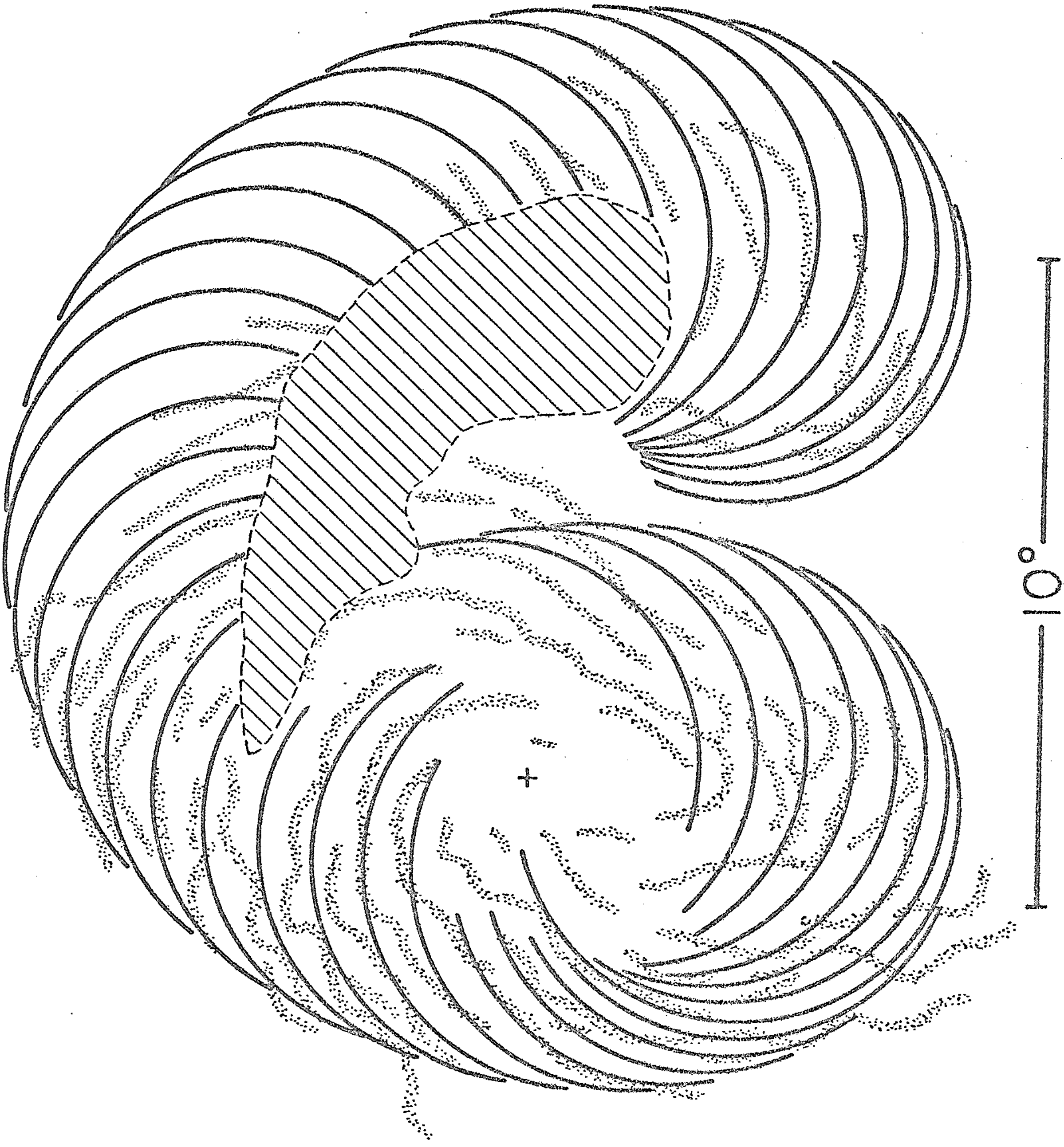
Figure 5.14 Comparison between the observed (shaded) and predicted (solid) distribution of plate edges, if the plates were perfectly circular, had a constant radius of 3.4°, and were centered at the locations (30 points/cycle, equally spaced in time) shown on the model path of Figure 5.13. The cross-hatched region represents a high-albedo deposit (thick enough to obscure the plate edges) which is thought to represent a CO_2-ice reservoir (Murray and Malin, 1973b).


10°

elliptical component (with a major axis of 8.0° and a minor axis of 5.3°) and a secular component (with a constant speed of 8.0°/cycle) propagating in a direction perpendicular to the major axis of the elliptical component. These model parameters are probably accurate to about 10%, which is sufficient to rule out the possibility that the periodic component could be circular (in which case the shape of the polar path would be the same regardless of the direction of propagation of the secular component).

Figure 5.14 compares the observed distribution of plate edges with the pattern predicted by the model parameters defined above and assuming a constant plate radius of 3.4°. The model north polar wandering path seems to explain not only the gross distribution of the layered terrain but also some of its finer details. For example, the model predicts two lobes of layered deposits separated by a trough (Chasma Boreale), as is indeed observed. In addition, the model accounts for one of the most spectacular features of the north polar region: the counter-clockwise spiral formed by the stratigraphically youngest plate edges near the present position of the geographic pole.

Deviations between the predicted and observed patterns of plate edges can be attributed to variations in plate diameter and circularity and to effects of topography on plate location and shape. The plate diameter will be sensitive to such parameters as the long-term temperature balance at the pole and the supply of dust and volatiles. Local topography and regional slopes can be expected to influence the location of the polar cap by shifting the point of minimum insolation away from the pole. Indeed, the south polar perennial cap is offset from the south geographic pole by about 3°. This offset has been explained by Dzurisin and Blasius (1975) as resulting from a regional slope at the south pole due to the proximity of a major impact basin.

The largest systematic deviation between the inferred and model north polar wandering paths occurs near the younger hairpin bend. The centers of curvature of the plates that define this bend appear to be shifted toward Chasma Boreale, and the youngest plates seem to be most greatly distorted from radial symmetry. This distortion may result in part from the younger plates' having to override a thick accumulation of older layered terrain, as the polar wandering path doubles back on itself. In addition, it is possible that some of the model parameters (i.e., the lengths of the major and minor axes of the oscillatory component or the speed and direction of propagation of the secular component) have not been optimized or may have changed slightly during the 1.2 oscillations. Alternatively, one could argue that this simple, hypothetical polar wandering model has no validity whatsoever.

Certainly, the notion that the polar layered deposits have recorded polar wandering on Mars can be challenged on a number of grounds. Perhaps the most serious problem is the failure of the south polar layered deposits to exhibit a well-defined polar wandering path, anti-symmetric with that observed in the north. However, the regional slope which seems to exist near the south pole, the higher mean elevation of the surface on which the south polar layered deposits rest, and the relative thinness (1-2 km) of the southern layered deposits compared to the 4-6 km thickness in the north (Dzurisin and

Blasius, 1975) are all factors that may well have distorted the pole path recorded in the south. In partial support of the model, the lobe of layered terrain extending farthest from the south pole is centered roughly along the 180° meridian, whereas the lobe farthest from the north pole is approximately along the 0° meridian (c.f., Soderblom *et al*. (1973), figure 8; Murray *et al* (1972), figure 4). Thus, the north and south polar layered terrain's asymmetric distribution with respect to the present position of the geographic pole is at least grossly consistent with the hypothesis of polar wandering on Mars.

Nash (1974) has challenged the polar wandering hypothesis of Murray and Malin (1973a) on the basis of a test designed to discriminate older from younger plate edges:

> It was assumed that if the features [dark, curvilinear bands] are constructional, continued exposure to erosional processes would change their planimetric form with time. Murray has postulated which features are oldest and which are youngest. It was found, however, that the supposed oldest features were statistically neither significantly more nor less irregular than the supposed youngest escarpments.
>
> This evidence has led to the conclusion that the features were probably carved nearly simultaneously into a previously deposited, featureless, stratified plateau rather than being constructed sequentially.

However, the validity of Nash's test of the polar wandering hypothesis is highly questionable for a number of reasons. First, Murray and Malin (1973a) did *not* postulate which plate edges are oldest and which are youngest, because Murray and Malin's north and south polar wandering paths are both *discontinuous*. Second, the time sequence of plate edges actually tested by Nash differs radically from that inferred in this present study. For example, the oldest and youngest plate edges of Nash (numbered 1 and 22 in his Figure 9) are inferred to have been deposited consecutively in this present study (Figure 5.12). Moreover, Nash has extended his plate edges 7 and 9 across a featureless, high-albedo deposit (Figure 5.14) which Murray and Malin (1973b) have suggested represents a large CO_2-ice reservoir "thick enough to bury the local topography." Nash's plate edges 7 and 9, therefore, are a composite of what I have inferred to be some of the oldest and youngest plate edges shown in his Figure 9. Finally, Nash has completely excluded from his test the plate edges farthest from the present position of the north pole (plates 1 to 16 of this study) and hence those that might be expected to be the most irregular. The outcome of Nash's test, therefore, is perhaps not so much a refutation of polar wandering on Mars as it is a predictable consequence of his inferred sequence of plate deposition.

Thus, although it is not without some difficulties, the hypothesis of polar wandering on Mars still seems to be a plausible explanation for the offset of centers of curvature of the dark, quasi-circular features. Clearly, however, processes must act in the polar regions to distort plate edges from being perfect "fossil latitudinal circles". Whether these processes modified the plate edges during or long after deposition is still an open question. All that the polar wandering hypothesis really requires is that the dark, curvilinear bands be primarily constructional, rather than erosional, features and that each plate be approximately radially symmetric about a former position of the spin axis. Support for a constructional origin of these curvilinear bands has been reported by Cutts _et al_. (1978) based on high-resolution images recently obtained by the Viking 2 orbiter.

Only about 1.2 polar wandering cycles appear to have been recorded by the Martian north polar layered deposits. Since the estimated total time represented by these deposits is 50 to 500 m.y., one polar wandering cycle would be completed in about 40 to 400 m.y. Earlier cycles could perhaps have occurred and the corresponding layered deposits (which would now be at latitudes lower than about 75°) subsequently eroded away. However, based on the distortions of plates from radial symmetry that seem to result when the polar wandering path loops back upon itself, it is difficult to imagine that the oldest plates (numbered 1 to 16 in Figure 5.12) could have formed on top of pre-existing layered terrain. Alternatively, earlier polar wandering cycles may have occurred on Mars at a time when the atmosphere was too thin to transport significant amounts of dust and volatiles to the poles.

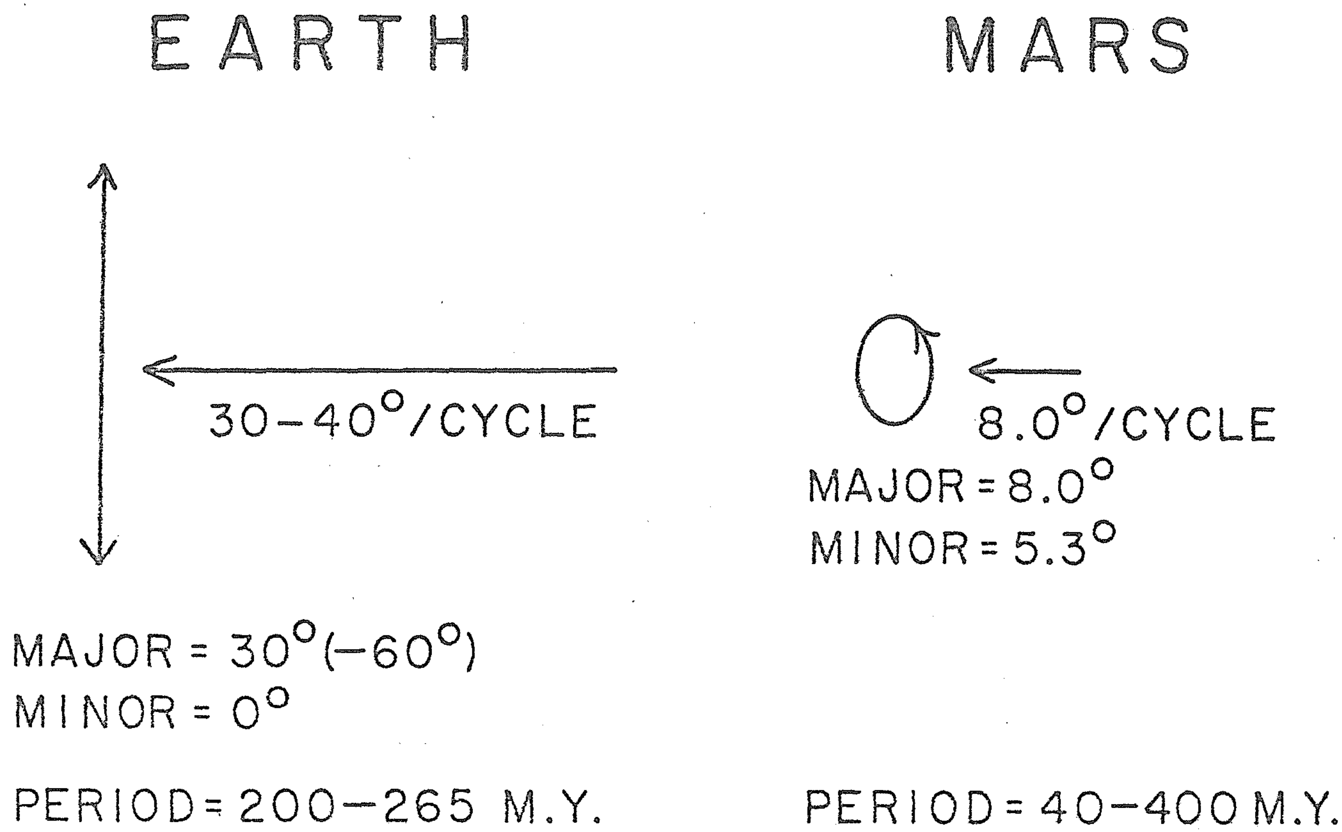
Synthesis of Terrestrial and Martian Observations

A comparison between apparent polar wandering with respect to North America and the inferred true polar wandering on Mars reveals a remarkable similarity: both processes can be modeled as the superposition of a (quasi-) periodic ($\sim 10^8$ years) component and a secular component. Admittedly, both the amplitude and eccentricity of the oscillatory component on Earth and Mars are considerably different. What is most curious, however, is that in addition to the oscillatory component, both planets exhibit a secular drift of the pole in a direction perpendicular to the major axis of the periodic component; and on both planets, the length of the secular shift per cycle is roughly the same as the length of the major axis of the oscillatory component. (These relationships are summarized and shown to scale in Figure 5.15.) It is tempting to suggest, therefore, that the APW path for North America and the inferred Martian north polar wandering path owe their regularity to a common origin: a highly ordered form of true polar wandering.

But what process might operate on both Earth and Mars to excite an oscillation with a period near 10^8 years? Probably the most widely recognized natural phenomenon with such a long period is the galactic year. The time it takes the sun to complete one revolution around the galactic center has been estimated as 2-3 x 10^8 years (see Steiner (1974) for a review of recent determinations). The period of galactic rotation has been associated not only with the quasi-periodicity of glaciations (Steiner and Grillmair, 1973; Williams, 1975) but also with long-term variations in the polarity of the geomagnetic field (Crain _et_ _al_., 1969; Niitsuma, 1973).

That the galactic year and polar wandering cycles on Earth and Mars may have similar periods is almost certainly fortuitous. Any galactic influence would constitute an exterior torque, which could conceivably influence a planet's obliquity and other orbital parameters. However, obliquity variations do not change the point of intersection of a planet's spin axis with its surface. Hence, changes in obliquity could neither be detected paleomagnetically nor be responsible for

Figure 5.15 Comparison between the proposed quasi-periodic polar wandering on Earth and Mars. The inferred polar wandering paths of both planets can be modeled by superposed periodic and secular components, shown here at the same scale. On both planets, the secular component propagates in a direction nearly perpendicular to the major axis of the periodic component and travels during one cycle a distance nearly equal to the length of the major axis of the periodic component.


EARTH
30–40°/CYCLE
MAJOR = 30°(–60°)
MINOR = 0°
PERIOD = 200–265 M.Y.
MARS
8.0°/CYCLE
MAJOR = 8.0°
MINOR = 5.3°
PERIOD = 40–400 M.Y.

the offset of centers of curvature of the Martian polar plates. Finally, the regularity of the galactic year is inconsistent with the _quasi_-periodicity of apparent polar wandering with respect to North America.

An internal process common to both Earth and Mars would seem to be a more likely mechanism for exciting quasi-periodic polar wandering on these planets. Interestingly, large-scale anomalies occur in the low-order harmonics of the gravity fields of both Earth (Gaposchkin, 1974) and Mars (Jordan and Lorell, 1975). Even more important, these low-order-harmonic gravity anomalies, which have been attributed to mantle convection cells, are symmetric about the equators of both planets (Goldreich and Toomre, 1969; Murray and Malin, 1973a). But this distribution of gravity anomalies is exactly what would be expected if polar wandering had occurred on both planets, for it is the evolution of these density inhomogeneities which "'steer' the rotation axis so as to maximize the resultant polar moment of inertia" (Goldreich and Toomre, 1969). If the evolution of these gravity anomalies drives quasi-periodic polar wandering, then some phenomenon seems to provide them with a long-term "memory". Perhaps the most plausible source of an internal pendulum in both Earth and Mars is oscillatory (time-dependent or intermittent) convection.

One form of time-dependent convection has been observed in externally-heated, Newtonian fluids with high Prandtl number and a Rayleigh number between 6×10^4 and 10^6 (Krishnamurti, 1970). The time-dependence is in the form of "knots" associated with temperature anomalies that are advected with the cellular flow. The period of the temperature fluctuations is given by the orbit time or twice the orbit time of the temperature anomalies. A mechanism that may help explain this form of time-dependent convection has been described by Keller (1966) and Welander (1967).

Another form of time-dependent convection occurs in Newtonian fluids that are heated from within. Two-dimensional numerical simulations by McKenzie et al. (1974) of convection in such a fluid exhibited a time-dependence at Rayleigh numbers above about 10^6 (for a fluid with infinite Prandtl number). The time-dependence consisted of the quasi-periodic instability of the upper thermal boundary layer, characterized by the formation and sinking of a cold blob of fluid. McKenzie et al. (1974) noted that the results of their numerical simulations closely resembled the time-dependent behavior observed by Kulacki and Goldstein (1972) in convection experiments on an internally-heated layer confined between rigid isothermal plates.

On the other hand, convection may become fully intermittent in externally-heated, Newtonian fluids with the very high Prandtl number typical of the earth's mantle and Rayleigh numbers near 10^7. Intermittent convection has been described by the following cyclic process:

> . . .the formation of a thermal boundary layer by diffusion, the instability of this layer when it becomes sufficiently thick, the destruction of the layer by convective flow, the dying down of the convection, and the reforming of the thermal boundary layer by diffusion. (Foster, 1971).

The phenomenology of intermittent convection has been explored numerically by Howard (1966) and Foster (1971) and experimentally by Sparrow et al. (1970). In addition, to explain the long-term variations in geomagnetic reversal frequency, Jones (1977) has proposed that the earth may experience intermittent, whole-mantle convection, which would be expected to cause periodic temperature fluctuations at the core-mantle boundary. Assuming a uniform lower mantle viscosity of 10^{22} poises (inferred by Cathles (1975) from post-glacial uplift of the Canadian shield), Jones computed a Rayleigh number of 3.2×10^7 and a period of intermittency of 309 m.y. Jones' calculated period of intermittency, however, is perhaps only an order-of-magnitude estimate, since the period will be influenced by possible non-uniform viscosity of the lower mantle, as well as by properties of the thermal boundary layer itself.

Several characteristics of the oscillatory forms of convection suggest that mantle convection may excite the proposed quasi-periodic polar wandering on Earth and Mars. First, the periods of the oscillatory forms of convection, when applied to the mantles of these planets, could well fall in the 10^8-year range, which is consistent with the time-scales of inferred true polar wandering cycles on Earth and Mars. Second, oscillatory forms of convection are not as strictly periodic as astronomically-forced oscillations. Indeed, the approximately 15% variation in the period of Phanerozoic APW cycles with respect to North America resembles the variations in periods in the numerical analyses of convection at high Rayleigh number by Welander (1967), McKenzie *et al*. (1974), and Foster (1971). Third, oscillatory forms of mantle convection would be associated with quasi-periodic temperature and hence density anomalies, which might be capable of driving quasi-periodic polar wandering, depending on the size and number of the anomalies. Fourth, some form of unsteady mantle convection is suggested by the geologic record, as has been discussed by Elder (1967), Rice (1972), and Walzer (1974). Finally, if the convective flow does involve the whole mantle, as seems likely from the analysis of Davies (1977), then periodic temperature fluctuations at the core-mantle boundary might explain the apparent long-term variation in geomagnetic reversal frequency.

Although oscillatory forms of mantle convection are attractive candidates for exciting quasi-periodic polar wandering on Earth and Mars, the dynamics of the process are obscure. How could any form of oscillatory convection (which is, after all, on the transition to turbulence) produce the apparently simple geometries of the observed polar wandering? (Unless the convective flow can be modeled by spherical harmonics of very low degree (c.f., Runcorn, 1962; Kanasewich, 1976; Gough, 1977; Kanasewich et al., 1978) so that perhaps only a single cell exhibits time-dependence.) Specifically, why should an oscillatory component of polar wandering be observed on both Earth and Mars in conjunction with a secular drift of the pole, comparable in size to the amplitude of the oscillations? And why should the direction of propagation of the secular drift be perpendicular to the major axis of the oscillatory component?

A skeptic could certainly argue that the entire idea of quasi-periodic polar wandering on Earth and Mars is sheer speculation or a figment of the imagination. Undoubtedly, the human mind is prone to infer the existence of lines, circles, and cycles in nature to establish order out of seeming chaos. Moreover, a small segment of even a random-walk polar wandering path could certainly exhibit quasi-periodicity. In addition, Earth and Mars may undergo processes, unique to each planet, that could cause the apparent regularities of the inferred polar wandering paths to have separate explanations. Finally, the unexpected nature of quasi-periodic polar wandering and the apparent lack of a simple dynamic mechanism also encourage doubt as to whether a highly ordered form of polar wandering has ever occurred on either planet.

For these reasons, I regard the notion of quasi-periodic polar wandering on Earth and Mars as an "outrageous hypothesis":

> The idea is set forth simply as an outrage, to do violence to certain generally established views about the earth's behavior that perhaps do not deserve to be regarded as established: and it is set forth chiefly as a means of encouraging the contemplation of other possible behaviors: not, however, merely a brief contemplation followed by an off-hand verdict of 'impossible' or 'absurd,' but a contemplation deliberate enough to seek out just what conditions would make the outrage seem permissible and reasonable.
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> --W. M. Davis, 1926
> "The Value of Outrageous Geological Hypotheses"

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Appendix A

COMPUTER PROGRAM FOR DETERMINING MODES OF DIRECTIONAL DATA

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C
C     PROGRAM  'MODE'
C
C      THIS PROGRAM MAY BE USED EITHER: 1) TO CONTOUR, PLOT, AND COMPUTE
C     THE MEAN, MODE, AND FISHERIAN STATISTICS OF SETS OF PALEOMAGNETIC
C     DIRECTIONS OR POLES; 2) MERELY TO PLOT THEM ON AN AZIMUTHAL EQUI-
C     DISTANT OR LAMBERT EQUAL-AREA PROJECTION; 3) MERELY TO COMPUTE THE
C     MEAN AND FISH. STATS.; OR 4) MERELY TO COMPUTE THE MODE, MEAN, AND
C     FISHERIAN STATISTICS.
C        IN COMPUTING THE MODE, EACH POINT FROM A GIVEN DATA SET IS
C     REPRESENTED BY A FISHERIAN PROB. DENSITY FUNCTION ('SMALL GRID'),
C     THE TIGHTNESS OF WHICH IS DETERMINED BY BOTH: 1) A COMPUTED ESTI-
C     MATE OF THE PRECISION PARAMETER (KAPPA) OF EITHER THE TOTAL SAMPLE
C     OR A SUBSET OF THE TOTAL SAMPLE (E.G. THE ENDPOINT OF A 'STREAKED'
C     DISTRIB); AND 2) AN ARBITRARILY-ASSIGNED 'SMOOTHING PARAMETER'.
C     THE INDIVIDUAL FISHERIAN PROB. DISTRIBS. ARE STAMPED ONTO A 'BIG
C     GRID', AND THE GRIDPOINT WITH THE HIGHEST PROB. VALUE IS IDENTI-
C     FIED AS THE MODE.  IF DESIRED, THE PROB. VALUES OF THE BIG GRID
C     ARE CONTOURED IN % OF THE MAXIMUM PROBABILITY AT THE MODE.
C
C     MANY DATA SETS MAY BE SUBMITTED IN A SINGLE RUN (UP TO 500 VECTORS
C     PER DATA SET), BUT EACH SET MUST BE PRECEDED BY 2 HEADER CARDS:
C
C     CARD 1: COLS.1 TO 8: LEAVE BLANK IF ONLY POINT PLOTS OR FISHERIAN
C                 STATS ARE DESIRED.
C                    TO LOCATE THE MODE ACCURATELY, IT IS IMPORTANT THAT
C                 ALL VECTORS BE ROTATED TO A PROJECTION CENTERED NEAR THE
C                 MODE.  IF THE DISTRIB. IS NEARLY FISHERIAN, LEAVE COLS.
C                 1 TC 8 BLANK, AND THE PROJECTION WILL AUTOMATICALLY BE
C                 CENTERED AT THE MEAN OF ALL THE POINTS.
C                    IF THE DISTRIB. IS SKEWED OR CONTAINS MANY OUTLIERS
C                 OR IS MULTIMODAL, THE PROJ. SHOULD BE CENTERED ON THE
C                 MEAN OF ONLY THOSE POINTS NEAR THE MODE OF INTEREST.
C                 ENTER A NUMBER OF THE SELECTED PTS. IN COLS. 1 TO 8
C                 (INTEGER, RIGHT-JUSTIFIED), AND PLACE THESE SELECTED
C                 PTS. AT THE BEGINNING OF THE DATA SET.
C                    POINTS GREATER THAN 90 DEGREES FROM THE CENTER OF THE
C                 PROJ. ARE EXCLUDED FROM DETERMINATION OF THE MODE.  THUS
C                 IF THE DISTRIB. IS BIMODAL (E.G. IF TWO POLARITIES ARE
C                 REPRESENTED), CENTER THE PROJ. NEAR THE MODE OF ONE
C                 POLARITY (BY ENTERING IN COL. 1 TO 8 A SELECTED NUMBER
C                 OF THE POLARITY OF INTEREST AND PLACING THESE SELECTED
C                 SAMPLES AT THE BEGINNING OF THE DATA SET).
C                    REMEMBER: THE SUBSET OF SELECTED SAMPLES SHOULD HAVE
C                 A FISHERIAN KAPPA REPRESENTATIVE OF THE TRUE SCATTER
C                 ABOUT THE MODE (I.E. INCLUDE ALL PTS. AROUND THE MODE,
C                 AND EXCLUDE ONLY THE OUTLIERS OR SKEWED TAILS).
C
C               COLS.11 TO 80: ALPHANUMERIC LABEL FOR DATA SET
C
C      CARD 2: ENTER ALL NUMBERS IN REAL FORMAT
C               COL. 1: ENTER 1. TO TREAT GEOGRAPHIC DIRECTIONS
C                       ENTER 2. TO TREAT STRATIGRAPHIC DIRECTIONS
C                       ENTER 3. TO TREAT VGP'S
C                COL. 11: ENTER 0. TO COMPUTE THE MODE, MEAN, FISH. STATS,
C                         AND PLOT AND CONTOUR POINTS.
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C                    ENTER 1. MERELY TO PLOT POINTS
C                    ENTER 2. MERELY TO COMPUTE MEAN & FISH. STATS.
C                    ENTER 3. MERELY TO COMPUTE THE MODE, MEAN, AND
C                    FISHERIAN STATISTICS.
C     ***THE REST OF THIS CARD CAN BE LEFT BLANK IF 1. OR 2. WAS ENTERED
C        IN COL. 11***
C              COL. 16: ENTER 0. FOR AZIMUTHAL EQUIDISTANT PROJECTION
C                    (THIS MUST BE USED FOR DETERMINING THE MODE)
C                    ENTER 1. FOR LAMBERT EQUAL-AREA PROJECTION
C              COL. 21: ENTER DIAM. IN INCHES (REAL) OF PLOTTED CIRCLE
C              COL. 26: ENTER 0. FOR NO POINT ANNOTATION
C                    ENTER 1. FOR POINT ANNOTATION (SM.#)
C              COL. 31: ENTER VALUE OF LOWEST CONTOUR IN % OF MAXIMUM
C                    PROBABILITY (REAL FORMAT). VALUES LESS THAN 35.
C                    MAY RESULT IN AESTHETICALLY DISPLEASING CONTOURS
C                    IF A SMALL # OF SCATTERED POINTS ARE TREATED.
C              COL. 41: ENTER VALUE OF HIGHEST CONTOUR IN % OF MAXIMUM
C                    PROBABILITY (REAL FORMAT)
C              COL. 51: ENTER CONTOUR INTERVAL IN % OF MAXIMUM
C                    PROBABILITY (REAL FORMAT)
C              COL. 61: ENTER SMOOTHING PARAMETER (REAL FORMAT)
C                    IN GENERAL, SET = 1.0.  AMOUNT OF SMOOTHING IS
C                    INVERSLEY PROPORTIONAL TO SMOOTHING PARAMETER.
C              COL. 71: ENTER THE DESIRED GRIDPOINT SPACING IN WHOLE
C                    DEGREES (REAL FORMAT).   MAX. ERROR IN LOCATING
C                    MODE IS 1.4 DEG FOR 2 DEG GRID AND 0.7 DEG FOR
C                    1 DEG GRID.  1 DEG GRID IS RECOMMENDED FOR PLOTS
C                    TO BE PUBLISHED OR WHEN FUNDS ARE NOT TIGHT.
C                    FOR 2. DEG. GRID, SGRID=61;BGRID=A=E=F=D=91
C                    FOR 1. DEG. GRID, SGRID=121,BGRID=A=E=F=D=181
C                 ***BE SURE DIMENSION STATEMENT (LINE 1) IS
C                    CONSISTENT WITH GRIDPOINT SPACING***
C              COL. 76: ENTER 0. IF PLOT IS TO BE CENTERED ON
C                    NLAT/INC=90. (THIS OPTION PERMITS CONTOUR PLOTS
C                    OF NORTHERN HEMIS. POLES OR POSITIVE-INC. DIREC-
C                    TIONS IN AN UNROTATED COORD. SYSTEM; HOWEVER, IT
C                    YIELDS LESS ACCURATE MODES.)
C                    ENTER 1. IF PLOT IS TO BE CENTERED ON THE MEAN
C                    OR SELECTED MEAN.
C                 ***1. MUST BE ENTERED FOR ACCURATE DETERMINATION OF
C                    THE MODE, EVEN IF NO PLOTTING IS REQUESTED***
C
C     END CARD: EACH DATA SET MUST BE TERMINATED BY A CARD WITH 'END' IN
C               COLS. 1 TO 3.
C
C     DATA CARD FORMAT:
C        THIS IS THE FORMAT USED AT THE U.S.G.S. PALEOMAGNETICS LAB
C        (FLAGSTAFF, AZ.).  SAMPLE INTERVAL, SAMPLE INTENSITY, AND
C        INDUCED/REMANENT RATIO ARE NOT USED IN THE PROGRAM 'MODE'.
C              COLS. 1 TO 16: ENTER SAMPLE DESIGNATION (ALPHANUMERIC)
C              (ONLY THE PORTION OF THE SAMPLE DESIGNATION BEGINNING IN
C               COL. 9 IS PLOTTED IF ANNOTATION IS REQUESTED).
C              COLS.17 TO 23: ENTER SAMPLE INTERVAL (REAL)
C              COLS.25 TO 29: ENTER GEOGRAPHIC DECLINATION (REAL)
C              COLS.31 TO 35: ENTER GEOGRAPHIC INCLINATION (REAL)
C              COLS.37 TO 41: ENTER STRATIGRAPHIC DECLINATION (REAL)
C              COLS.43 TO 47: ENTER STRATIGRAPHIC INCLINATION (REAL)
C              COLS.49 TO 53: ENTER VGP LATITUDE (REAL)
C              COLS.55 TO 60: ENTER VGP LONGITUDE (REAL) (POSITIVE=EAST)
C              COLS.62 TO 69: ENTER SAMPLE INTENSITY/ (E9.2 FORMAT)
C              COLS.71 TO 74: ENTER INDUCED/REMANENT RATIO (REAL)
C              COLS.76 TO 80: ENTER DEMAG. STEP (ALPHANUMERIC)
C
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      DIMENSION                AINTVL(500),ADECG(500),AINCG(500),
     FADECS(500),AINCS(500),ALAT(500),ALONG(500),AINTEN(500),AIR(500),
     FX(500),Y(500),DIFF(500),RCORD(18),DD(3),TITLE(3),XTTL(4),
     FSGRID(121,121),BGRID(181,181),A(181),E(181),F(181),D(181),
     FDIFFH(500),ZZH(10),YYH(10),XXH(10),ANH(10),AKH(10),ADH(10)
      DOUBLE PRECISION ADEMAG(500),ASAMP1(500),ASAMP2(500),END
      COMMON/ROTUV/XROT,YROT
      COMMON/PLTDM/XLNGTH
      DATA   DD/'STR','PLOT',1./
      DATA TITLE/'GEOG','STRT','VGP '/
      DATA XTTL/0.,' COO','RDIN','ATES'/
      DATA END/'END'/
      ITEST=1
      DEGRE=180./3.14159
      RADIA=1./DEGRE
C     READ TWO CONTROL CARDS
 3    READ(5,8,END=600) NPGOOD,RCORD
 8    FORMAT(4X,I4,18A4)
      READ (5,7) FLAG5,FLAG7,FLAG8,XLNGTH,FLAG9,AL,BL,CL,AFOC,DEG,CEN
 7    FORMAT(F10.0,4F5.0,4F10.0,2F5.0)
C     COUNT AND READ EACH DATA CARD IN FIRST DATA SET
      I=1
 6    READ(5,11)  ASAMP1(I),ASAMP2(I),AINTVL(I),ADECG(I),AINCG(I),
     FADECS(I),AINCS(I),ALAT(I),ALONG(I),AINTEN(I),AIR(I),ADEMAG(I)
 11   FORMAT(2A8,F7.0,5F6.C,F7.0,E9.0,F5.0,A6)
C     END OF DATA SET IS MARKED BY CARD WITH 'END' IN COLS. 1 TO 3
      IF(ASAMP1(I).EQ.END) GO TO 5
      I=I+1
      GO TO 6
C     NPOL IS THE NUMBER OF POINTS IN THE DATA SET
 5    NPOL=I-1
C     CONVERT ALL LONG.'S TO POSITIVE NOS. (DEGREES EAST).
      DO 9 I=1,NPOL
      IF(ALONG(I).LT.0.) ALONG(I)=360.+ALONG(I)
 9    CONTINUE
C     SET INC.'S=LAT.'S AND DEC.'S=LONG.'S
 15   IF(FLAG5.EQ.3.) GO TO 25
      IF(FLAG5.EQ.1.) GO TC 22
      DO 20 I=1,NPOL
      ALAT(I)=AINCS(I)
      ALONG(I)=ADECS(I)
 20   CONTINUE
      GO TO 25
 22   DO 24 I=1,NPOL
      ALAT(I)=AINCG(I)
      ALONG(I)=ADECG(I)
 24   CONTINUE
 25   IFLAG=FLAG5
      XTTL(1)=TITLE(IFLAG)
C     IDENTIFY DATA DECK AND TYPE OF COORDINATES
      WRITE(6,12) RCORD,NPOL,XTTL
 12   FORMAT(1H1,//7X,18A4/8X,I4,1X,4A4)
      IF(FLAG7.EQ.0.) GO TO 27
      WRITE(6,14) (ASAMP1(I),ASAMP2(I),ALAT(I),ALONG(I),ADEMAG(I),I,I=1,
     FNPOL)
 14   FORMAT(/8X,'SAMPLE #',5X,'INC/LAT',1X,'DEC/LONG',4X,'DEMAG',4X,
     F'POINT #'//(4X,2A8,1X,F6.1,3X,F6.1,5X,A6,5X,I4))
C     IF ONLY STATISTICS ARE DESIRED, GO TO 18
      IF(FLAG7.EQ.2.) GO TO 18
C     SET PLOT SCALING PARAMETERS
 27   YLNGTH=XLNGTH
      SX=XLNGTH/180.
      SY=SX
```

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      FLAG3=0.
C     IF ONLY POINT PLOT IS DESIRED, GO TO 477
      IF (FLAG7.EQ.1.) GO TO 477
C     SCALE SIZE OF BIG AND SMALL GRIDS
      NBGRID=180/DEG+1
      NSGRID=120/DEG+1
C     ZERO SMALL GRID
      DO 16 N=1,NSGRID
      DO 17 M=1,NSGRID
      SGRID(N,M)=0.0
 17   CONTINUE
 16   CONTINUE
C     CONVERT LAT. AND LONG. FROM DEGREES TO RADIANS
 18   DO 19 N=1,NPOL
      ALAT(N)=ALAT(N)*RADIA
      ALONG(N)=ALONG(N)*RADIA
 19   CONTINUE
C     COMPUTE AND WRITE MEAN AND FISHERIAN STATISTICS OF ALL DATA.
C     THEN, IF DESIRED, COMPUTE SELECTED MEAN AND ITS STATISTICS.
      FLAG1=0.
 26   ZZ=0.
      YY=0.
      XX=0.
      AN=0.
      IF(FLAG1.EQ.1.) GO TO 33
      DO 31 N=1,NPOL
      ZZ=ZZ+SIN(ALAT(N))
      YY=YY+COS(ALAT(N))*SIN(ALONG(N))
      XX=XX+COS(ALAT(N))*COS(ALONG(N))
      AN=AN+1.
 31   CONTINUE
      GO TO 41
 33   DO 40 N=1,NPGOOD
      ZZ=ZZ+SIN(ALAT(N))
      YY=YY+COS(ALAT(N))*SIN(ALONG(N))
      XX=XX+COS(ALAT(N))*COS(ALONG(N))
      AN=AN+1.
 40   CONTINUE
 41   IF(AN.EQ.1.) GO TO 47
      RR=SQRT(ZZ*ZZ+YY*YY+XX*XX)
      AK=(AN-1.)/(AN-RR)
      SIGMA=81./SQRT(AK)
      AD=ARCOS(1.+(AN-RR)*(1.-20.**(1./(AN-1.)))/RR)*DEGRE
      ALTWD=ARSIN(ZZ/RR)*DEGRE
      ALNWD=ATAN(YY/XX)*DEGRE
      IF(XX.LE.0.) GO TO 32
      IF(YY.LE.0.) ALNWD=360.+ALNWD
      GO TO 34
 32   ALNWD=180.+ALNWD
 34   IF(FLAG1.EQ.1.) GO TO 44
      WRITE(6,35) ALTWD,ALNWD,AN,RR,AK,AD,SIGMA
 35   FORMAT(//5X,'TOTAL MEAN: NLAT/INC=',F8.3,2X,'ELONG/DEC=',F8.3
     F/17X,'N=',F8.3,2X,'R=',F8.3,2X,'K=',F8.3,2X,'ALPHA95=',F8.3,2X,
     F'SIGMA=',F8.3)
      IF(NPGOOD.EQ.0.OR.NPGOOD.EQ.AN) GO TO 46
      FLAG1=1.
      GO TO 26
 44   WRITE(6,45) ALTWD,ALNWD,AN,RR,AK,AD,SIGMA
 45   FORMAT(//5X,'SELECTED MEAN: NLAT/INC=',F8.3,2X,'ELONG/DEC=',F8.3
     F/17X,'N=',F8.3,2X,'R=',F8.3,2X,'K=',F8.3,2X,'ALPHA95=',F8.3,2X,
     F'SIGMA=',F8.3)
C     IF ONLY STATISTICS ARE DESIRED, SKIP TO END OF PROGRAM
 46   IF(FLAG7.EQ.2.) GO TO 529
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      ZZ=ZZ/RR
      YY=YY/RR
      XX=XX/RR
C     BEGIN COMP. VALUES FOR SMALL GRID TO BE 'STAMPED' ONTO LARGE GRID.
C     FIRST, MULTIPLY KAPPA OF DISTRIB. BY SMOOTHING PARAMETER.
      AK=AFOC*AK
C     IF THERE IS ONLY 1 PT. IN A DATA SET, KAPPA IS THE NUMBER
C     APPEARING IN COL. 71 TO 74.
 47   IF(AN.EQ.1.) AK=AFOC*AIR(1)
C        NEXT, COMPUTE THE LIMITING ANGLE (BETA)
C     AT WHICH PROB. FOR THE GIVEN KAPPA IS 1% OF THE PROB. AT THE
C     CENTER.  I.E., THE SMALL GRID IS CLIPPED AT LOW PROB. VALUES TO
C     ELIM. UNNEC. CALCS.  IF THIS ANGLE EXCEEDS 60 DEGREES, THE SMALL
C     GRID IS CLIPPED AT 60 DEGREES ANYWAY.  THIS MAY CAUSE SHARP IRRE-
C     GULARITIES IN LOWER PROB. CONTOUR LINES IF A FEW SCATTERED POINTS
C     ARE CONTOURED.
      BETA=ARCOS(-4.605/AK+1.)*DEGRE
      IF(BETA.GT.60.) BETA=60.
      LIMANG=INT(BETA/DEG)
C     FINALLY, ASSIGN VALUES TO THE SMALL GRIDPTS. ACCORDING TO THE
C     EQUATION FOR A FISHERIAN DISTRIB.  THE PROB. AT THE CENTER IS SET
C     =1.  SINCE A FISHERIAN DISTRIB. IS CIRCULARLY SYMMETRIC, ONLY
C     VALUES FOR ONE QUADRANT ARE ACTUALLY COMPUTED.
      MID=NSGRID/2+1
      LEFT=MID-LIMANG
      DO 60 I=LEFT,MID
      DO 50 J=LEFT,MID
      APHI=(MID-I)*(MID-I)+(MID-J)*(MID-J)
      PHI=SQRT(APHI)
      PHIR=PHI*RADIA*DEG
      CCC=AK*(COS(PHIR)-1.)
      SGRID(I,J)=EXP(CCC)
      SGRID(NSGRID+1-I,J)=SGRID(I,J)
      SGRID(NSGRID+1-I,NSGRID+1-J)=SGRID(I,J)
      SGRID(I,NSGRID+1-J)=SGRID(I,J)
 50   CONTINUE
 60   CONTINUE
C     ZERO BIG GRID
 110  DO 130 N=1,NBGRID
      DO 120 M=1,NBGRID
      BGRID(N,M)=0.0
 120  CONTINUE
 130  CONTINUE
      WRITE(6,230)
 230  FORMAT(/8X,'SAMPLE #',5X,'INC/LAT',2X,'DEC/LONG',4X,'DEMAG',4X,
     F'POINT #')
C     NOW WE ARE READY TO BEGIN STAMPING THE SMALL GRIDS REPRESENTING
C     EACH PT. ONTO THE BIG GRID.  THE SECTION BETWEEN STMNTS. 350 AND
C     375 CONTAINS MANY BRANCHES, DEPENDING UPON WHETHER THE PT. IS A
C     DIRECTION OR A POLE AND UPON THE LOCATION OF THE CENTER OF THE
C     PROJECTION.  PTS. GREATER THAN 90 DEGREES FROM THE CENTER OF THE
C     PROJECTION ARE IGNORED.
      N2=1
      AN3=0.
 350  IF(CEN.EQ.1.) GO TO 352
      ALTD=ALAT(N2)*DEGRE
      ALND=ALONG(N2)*DEGRE
      IF(ALTD) 234,232,232
 352  ZZS=SIN(ALAT(N2))
      YYS=COS(ALAT(N2))*SIN(ALONG(N2))
      XXS=COS(ALAT(N2))*COS(ALONG(N2))
      DIFF(N2)=ARCOS(ZZ*ZZS+YY*YYS+XX*XXS)*DEGRE
C     CONVERT LAT. AND LONG. BACK INTO DEGREES.
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      ALTD=ALAT(N2)*DEGRE
      ALND=ALONG(N2)*DEGRE
C     PTS. GREATER THAN 90 DEGREES AWAY ARE NOT CONTOURED.
      IF(DIFF(N2)-90.) 232,232,234
 232  WRITE(6,233) ASAMP1(N2),ASAMP2(N2),ALTD,ALND,ADEMAG(N2),N2
 233  FORMAT(4X,2A8,1X,F6.1,3X,F6.1,5X,A6,5X,I4,3X,'CONTOURED')
      GO TO 250
 234  WRITE(6,235) ASAMP1(N2),ASAMP2(N2),ALTD,ALND,ADEMAG(N2),N2
 235  FORMAT(4X,2A8,1X,F6.1,3X,F6.1,5X,A6,5X,I4,3X,'NOT CONTOURED')
      GO TO 360
C     CONVERT D.'S AND I.'S INTO LONG.'S AND LAT.'S
 250  IF((CEN+FLAG5).EQ.4.) GO TO 253
      IF(FLAG5.EQ.3.) GO TO 251
      ALND=180.-ALND
      IF(ALND.LT.0.) ALND=360.+ALND
 251  IF(CEN.EQ.1.) GO TO 252
      XXXX=(90.-ALTD)*SIN(ALND*RADIA)+90.
      YYYY=-(90.-ALTD)*COS(ALND*RADIA)+90.
      GO TO 256
C     PREPARE VECTORS FOR INPUT INTO 'ROTATE' SUBROUTINE.  BECAUSE OF
C     ITS VAGARIES, ALL PTS. ARE FIRST TRANSPOSED TO THE SAME QUADRANT.
 252  ALNWD5=180.-ALNWD
      IF(ALNWD5.LT.0.) ALNWD5=360.+ALNWD5
 253  IF(FLAG5.EQ.3.) ALNWD5=ALNWD
      IF(ALNWD5.LT.90.) GO TO 300
      IF(ALNWD5.LE.180.) GO TO 304
      IF(ALNWD5.LE.270.) GO TO 302
      ALNWDR=ALNWD5-180.
      ALND=ALND-180.
      GO TO 254
 300  ALNWDR=ALNWD5+90.
      ALND=ALND+90.
      GO TO 254
 302  ALNWDR=ALNWD5-90.
      ALND=ALND-90.
      GO TO 254
 304  ALNWDR=ALNWD5
C     ROTATE ALL VECTORS TO BE CONTOURED TO THE PROJ. CENTERED ON MEAN.
 254  CALL ROTATE(ALTD,ALND,ALTWD,ALNWDR)
      XXXX=XROT
      YYYY=YROT
 256  XXX1=(180.-YYYY)*SX
      YYY1=XXXX*SY
C     IF CONTOUR PLOT IS NOT DESIRED, SKIP TO 260
      IF (FLAG7.EQ.3.) GO TO 260
C     PLOT THE ROTATED VECTORS TO BE CONTOURED.
      CALL CIRCLE(0.035,XXX1,YYY1,1)
      CALL SYSSYM(XXX1,YYY1,0.04,1,-1,0.)
C     ANNOTATE PTS., IF DESIRED.
      IF(FLAG9.EQ.1.) CALL SYSSYM(XXX1,YYY1+0.06,0.07,ASAMP2(N2),8,90.)
C     STAMP PREVIOUSLY COMPUTED SMALL GRID ONTO BIG GRID (I.E., SUM THE
C     INDIVIDUAL PROBABILITY DISTRIBUTIONS).  MAXIMUM ERROR IN ASSIGNING
C     CENTER OF SMALL GRID TO CORRECT BIG GRIDPT. IS 1.4 DEG FOR 2 DEG
C     GRID AND 0.7 DEG FOR 1 DEG GRID.
 260  IX=INT(XXXX/DEG+0.5)
      IY=INT(YYYY/DEG+0.5)
      NR=NBGRID-IY
      NS=IX+1
      NLEFT=NS-LIMANG
      NRIGHT=NS+LIMANG
      NTOP=NR-LIMANG
      NBOT=NR+LIMANG
      IF(NLEFT.LT.1) NLEFT=1
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      IF(NRIGHT.GT.NBGRID) NRIGHT=NBGRID
      IF(NTOP.LT.1) NTOP=1
      IF(NBOT.GT.NBGRID) NBOT=NBGRID
      DO 175 I=NLEFT,NRIGHT
      DO 170 J=NTOP,NBOT
      K=MID-NR+J
      L=MID-NS+I
      BGRID(J,I)=BGRID(J,I)+SGRID(K,L)
 170  CONTINUE
 175  CONTINUE
 355  AN3=AN3+1.
 360  N2=N2+1
      IF(N2-NPOL) 350,350,370
 370  WRITE(6,375) AN3
 375  FORMAT(/5X,'# POLES CONTOURED=',F8.3)
C     BETWEEN STMNTS. 380 AND 690, WE FIND THE POINT OF MAX. PROB. OF
C     THE BIG GRID (I.E., THE MODE) AND DETERMINE ITS LOCATION IN BOTH
C     ROTATED AND UNROTATED COORDINATE SYSTEMS.  FIRST, DETERMINE THE
C     PROB. VALUE AT THE MODE AND FIND ITS CARTESIAN COORDINATES.
 380  AMX=0.
      DO 410 I=1,NBGRID
      DO 400 J=1,NBGRID
      IF(BGRID(I,J).GT.AMX) GO TO 395
      GO TO 400
 395  AMX=BGRID(I,J)
      NNR=I
      NNS=J
 400  CONTINUE
 410  CONTINUE
      FX=(NNS-1)*DEG
      FY=(NBGRID-NNR)*DEG
 849  WRITE(6,850) AMX,FX,FY
 850  FORMAT(/5X,'THE MAXIMUM PROBABILITY IS',F8.3,2X,'AT X=',F5.1,
     F', Y=',F5.1)
      WRITE(6,851) AL,BL,CL,AFOL,DEG
 851  FORMAT(/5X,'CONTOUR PARAMETERS: LOWEST=',F5.2,'%  HIGHEST=',
     FF5.2,'%  INTERVAL=',F5.2,'%'/25X,'SMOOTHING PARAMETER=',F5.2,2X,
     F'GRIDPOINT SPACING=',F4.1,1X,'DEGREES')
C     CONVERT CARTESIAN COORDS. INTO LAT. AND LONG.
 852  XXX=FX-90.
      YYY=FY-90.
      AALAT=90.-SQRT(XXX*XXX+YYY*YYY)
      IF(YYY.EQ.0.) GO TO 435
      AALON=ATAN(XXX/YYY)*DEGRE
      IF(YYY.GT.0.) GO TO 430
      IF(XXX.LE.0.) AALON=360.-AALON
      IF(XXX.GT.0.) AALON=-AALON
      GO TO 440
 430  AALON=180.-AALON
      GO TO 440
 435  IF(XXX.GE.0.) AALON=90.
      IF(XXX.LT.0.) AALON=270.
C     FIRST TIME THROUGH, COMPUTE THE LAT AND LONG OF MODE IN ROTATED
C     COORD. SYSTEM, AND DETERMINE ITS DISTANCE FROM THE PROJ. CENTER.
 440  IF(CEN.EQ.1.) GO TO 439
      IF(FLAG5.EQ.3.) GO TO 464
      AALON=180.-AALON
      IF(AALON.LT.0.) AALON=360.+AALON
      GO TO 464
 439  IF(FLAG3.EQ.0.) GO TO 443
C     SECOND TIME THROUGH, DETERMINE LONG/DEC IN UNROT. COORD. SYSTEM.
      AALON=AALON+ALNWDR
      IF(ALNWD5.LT.90.) GO TO 460
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      IF(ALNWD5.LE.180.) GO TO 441
      IF(ALNWD5.LE.270.) GO TO 462
      AALON=AALON+180.
      GO TO 441
 460  AALON=AALON-90.
      GO TO 441
 462  AALON=AALON+90.
 441  IF(AALON.GT.360.) AALON=AALON-360.
      IF(FLAG5.EQ.3.) GO TO 463
      AALON=180.-AALON
      IF(AALON.LT.0.) AALON=360.+AALON
 463  ZMCDE=SIN(AALAT*RADIA)
      YMODE=COS(AALAT*RADIA)*SIN(AALON*RADIA)
      XMODE=COS(AALAT*RADIA)*COS(AALON*RADIA)
      DIFMM=ARCOS(ZZ*ZMODE+YY*YMODE+XX*XMODE)*DEGRE
      IF(DIFMM.LT.90.) GO TO 464
      AALAT=-AALAT
      AALON=180.+AALON
      IF(AALON.GT.360.) AALON=AALCN-360.
      ZMODE=-ZMODE
      YMCDE=-YMODE
      XMODE=-XMODE
 464  WRITE(6,442) AALAT,AALON
 442  FORMAT(//5X,'MODE: NLAT/INC=',F8.3,2X,'ELONG/DEC=',F8.3)
      GO TO 690
 443  DMM=90.-AALAT
      WRITE(6,444) DMM
 444  FORMAT(/5X,'THE MODE IS',1X,F4.1,1X,'DEGREES FROM THE CENTER OF TH
     FE PROJECTION')
      FLAG3=1.
C     ROTATE THIS LAT & LONG TO FIND LAT & LONG OF 'UNROTATED' MODE
      CALL ROTATE (AALAT,AALON,ALTWD,0.0001)
      FX=XROT
      FY=YROT
      GO TO 852
C     THIS SECTION (DO LOOPS 700 AND 750) COMPUTES THE FISHERIAN STATS.
C     OF PTS. INCREASINGLY CLOSE TO THE MODE. FIRST, THE DIFFERENCE
C     BETWEEN EACH DATA PT. AND THE MODE IS DETERMINED:
 690  DO 700 N=1,NPOL
      ZH=SIN(ALAT(N))
      YH=COS(ALAT(N))*SIN(ALONG(N))
      XH=COS(ALAT(N))*COS(ALONG(N))
      DIFFH(N)=ARCOS(ZMODE*ZH+YMODE*YH+XMODE*XH)*DEGRE
 700  CONTINUE
C     A FIRST CUT IS MADE AT 2*SIGMA OF THE SELECTED PTS.
      SIGMA2=2.*SIGMA
      SIGMAN=SIGMA2
      WRITE(6,710)
 710  FORMAT(///5X,'FISHERIAN STATISTICS OF POINTS INCREASINGLY CLOSE TO
     F MODE:'//8X,'CONF RADIUS',2X,'# CONTAINED',2X,'KAPPA',2X,'ALPHA95'
     F,2X,'NLAT/INC',2X,'ELONG/DEC',2X,'DIFF. MEAN-MODE')
      PRECIS=0.7071*DEG
      DO 750 N=1,20
      CN=N
      ANH(N)=0.
      ZZH(N)=0.
      YYH(N)=0.
      XXH(N)=0.
      DO 760 M=1,NPOL
      IF(DIFFH(M).GE.SIGMAN) GO TO 760
      ZZH(N)=ZZH(N)+SIN(ALAT(M))
      YYH(N)=YYH(N)+COS(ALAT(M))*SIN(ALONG(M))
      XXH(N)=XXH(N)+COS(ALAT(M))*COS(ALONG(M))
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      ANH(N)=ANH(N)+1.
  760 CONTINUE
      IF(ANH(N).LE.1.) GO TO 450
      RRH=SQRT(ZZH(N)*ZZH(N)+YYH(N)*YYH(N)+XXH(N)*XXH(N))
      AKH(N)=(ANH(N)-1.)/(ANH(N)-RRH)
      ADH(N)=ARCOS(1.+(ANH(N)-RRH)*(1.-20.**(1./(ANH(N)-1.)))/RRH)*DEGRE
      ALTWDH=ARSIN(ZZH(N)/RRH)*DEGRE
      ALNWDH=ATAN(YYH(N)/XXH(N))*DEGRE
      IF(XXH(N).LE.0.) GO TO 770
      IF(YYH(N).LE.0.) ALNWDH=360.+ALNWDH
      GO TO 780
  770 ALNWDH=180.+ALNWDH
  780 NPOINT=ANH(N)
      ZZH(N)=ZZH(N)/RRH
      YYH(N)=YYH(N)/RRH
      XXH(N)=XXH(N)/RRH
      DMMH=ARCOS(ZMODE*ZZH(N)+YMODE*YYH(N)+XMODE*XXH(N))*DEGRE
      WRITE(6,790) SIGMAN,NPOINT,AKH(N),ADH(N),ALTWDH,ALNWDH,DMMH
  790 FORMAT(/11X,F5.1,10X,I3,3X,F7.1,3X,F5.2,4X,F5.1,5X,F6.1,14X,F4.1)
      IF(DMMH.LE.PRECIS) WRITE(6,795) ADH(N)
  795 FORMAT(15X,'CANDIDATE FOR BETA95 =',F5.2,1X,'DEGREES')
C     THEN 19 MORE CUTS ARE MADE AT VALUES OF 0.95 TO 0.05*(2.*SIGMA)
C     IF THE NUMBER OF PTS. REMAINING DROPS TO 3, THE PROCEDURE STOPS.
      IF(ANH(N).LE.3.) GO TO 450
      SIGMAN=SIGMA2-CN*SIGMA2/20.
  750 CONTINUE
C     IF CONTOUR PLOT IS NOT DESIRED, SKIP TO END FOR NEW DATA SET
  450 IF (FLAG7.EQ.3.) GO TO 529
C     ASSIGN VALUES OF CONTOUR LINES
      AL=AMX*AL/100.
      BL=AMX*BL/100.
      CL=AMX*CL/100.
      ABGRID=NBGRID
C     AND CONTOUR VALUES OF BIG GRID.
      CALL TOPOG(BGRID,NBGRID,NBGRID,1.,ABGRID,1.,ABGRID,AL,BL,CL,BITE,
     FO,A,E,F,D)
      GO TO 526
  477 IF(FLAG5.EQ.3.) GO TO 500
C     CONVERT D.'S AND I.'S INTO LONG.'S AND LAT.'S
      DO 492 I=1,NPCL
      ALONG(I)=180.-ALONG(I)
      IF(ALONG(I).LT.0.) ALONG(I)=360.+ALONG(I)
  492 CONTINUE
C     PLOT VECTORS ON EITHER AZ. EQDST. OR LMBT. EQ. A. PROJECTION
  500 DO 525 I=1,NPCL
      ALONG(I)=ALONG(I)*RADIA
      ALAT7=ALAT(I)
      IF(ALAT(I).LT.0.) ALAT7=-ALAT(I)
      IF(FLAG8.EQ.0.) GO TO 506
      ALTEGA=(90.-ALAT7)/2.
      REQA=SIN(ALTEGA*RADIA)*SQRT(2.)*90.
      X(I)=REQA*SIN(ALONG(I))+90.
      Y(I)=-REQA*COS(ALONG(I))+90.
      GO TO 512
  506 X(I)=(90.-ALAT7)*SIN(ALONG(I))+90.
      Y(I)=-(90.-ALAT7)*COS(ALONG(I))+90.
  512 XX=(180.-Y(I))*SX
      YY=(X(I))*SY
C     SOLID SYMBOLS HAVE POSITIVE LAT/INC
      CALL CIRCLE(0.047,XX,YY,1)
      IF(ALAT(I).GE.0.) CALL SYSSYM(XX,YY,0.065,1,-1,0.)
      IF(ALAT(I).GE.0.) CALL SYSSYM(XX,YY,0.04,1,-1,0.)
C     ANNOTATE PTS., IF DESIRED
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      IF(FLAG9.EQ.1.) CALL SYSSYM(XX,YY+0.06,0.07,ASAMP2(I),8,90.)
 525  CONTINUE
C     DRAW CIRCLE OF DESIRED RADIUS.
 526  RADMAX=XLNGTH/2.
      CALL CIRCLE(RADMAX,RADMAX,RADMAX,1)
C     DRAW TICK MARK AT CENTER.
      CALL SYSSYM(RADMAX,RADMAX,0.15,3,-1,0.)
      XLNGT2=XLNGTH+0.5
C     WRITE TITLE BENEATH EACH PLOT
      SYMHT=(XLNGTH-1.)*7./372.
      CALL SYSSYM(XLNGT2+0.5,1.,SYMHT,RCORD,72,90.)
C     LABEL TYPE OF COORDINATES BEING PLOTTED
      CALL SYSSYM(XLNGT2+0.750,1.2,SYMHT,XTTL,16,90.)
C     DRAW TICK MARKS AT QUADRANT BOUNDARIES.  LONG. OF CONTOUR PLOTS
C     CENTERED ON MEAN IS NOT DEMARCATED.
      IF(CEN.EQ.1.) GO TO 528
      CALL SYSSYM(0.,RADMAX,0.15,13,-1,90.)
      CALL SYSSYM(RADMAX,0.,0.15,13,-1,0.)
      CALL SYSSYM(RADMAX,XLNGTH,0.15,13,-1,0.)
      CALL SYSSYM(XLNGTH,RADMAX,0.15,13,-1,90.)
      YLNGTH=XLNGTH/2.
C     LABEL LONG/DEC FOR PT. PLOT
      IF(FLAG5.EQ.3.) CALL SYSSYM(XLNGT2,YLNGTH,0.125,'0',1,90.)
      IF(FLAG5.LE.2.)CALL SYSSYM(XLNGT2,YLNGTH-.125,0.125,'180',3,90.)
 528  CALL SYSEND(-1,0)
C     SET ORIGIN FOR NEXT PLOT
      CALL SYSPAT(XLNGTH+4.)
C     RETURN TO BEGINNING FOR NEW DATA SET
 529  GO TO 3
 600  STOP
      END
C
C     THIS SUBROUTINE IS ONE OF THE CALTECH CONTOURING PROGRAMS,
C     MODIFIED SLIGHTLY TO BE COMPATIBLE WITH THE MAIN PROGRAM.
C     ONE DISADVANTAGE IS THAT IF THE PEN DRAGS, THE CONTOUR PLOT IS
C     COVERED WITH A SERIES OF PARALLEL LINES.  ALSO, OCCASIONALLY
C     (FOR NO OBVIOUS REASON) THE 95% CONTOUR IS SKIPPED--BEWARE.
C     EITHER TRY AGAIN WITH THE SAME PARAMETERS, OR TRY CHANGING THE
C     SMOOTHING PARAMETER SLIGHTLY.
C
      SUBROUTINE TOPOG(V,N,M,XMIN,XMAX,YMIN,YMAX,CNTRLO,CMAX,CNTRVL,
     1   WORDS,NCHAR,RA,RB,X,Y)
      DIMENSION V(N,M),RA(2),RB(2),X(2),Y(2),WORDS(2)
      SX=XMAX-XMIN
      SY=YMAX-YMIN
      IF(SX-SY)22,24,24
   22 SMAX=SY
      GO TO 1124
   24 SMAX=SX
 1124 SS=SX/SMAX
      SSS=SY/(2.0*SMAX)
      SYS=2.0*SSS
      Y1 = SY/SMAX - .001
      X2 = SX/SMAX - .001
      IF(NCHAR.GT.99) GO TO 1125
      DIMENSION FORMAT(2)
      DATA FORMAT /4H(E10,4H.3) /
 1125 YCONV=1.0/SMAX
      DELTAX=N-1
      DELTAX=(XMAX-XMIN)/DELTAX
      X(1) = 0.
      RB(1) = V(1,1)
      DO 27 J=2,N                                                    ,2
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     RB(J)=V(J,1)
 27  X(J)=X(J-1)+DELTAX
     DELTAY=M-1
     DELTAY=(YMAX-YMIN)/DELTAY
     Y(1) = 0.
    DO 28 J=2,M
 28  Y(J)=Y(J-1)+DELTAY
    DO 118  K=2,M
     DO 30 J=1,N
     RA(J)=RB(J)
 30  RB(J)=V(J,K)
    DO 118  J=2,N
 35  ASSIGN 112 TO L
    RR=RA(J)
    XX=X(J)
    YY=Y(K-1)
 37 RL=RR
    XL=XX
    YL=YY
 39 IF(RL-RA(J-1))   41,40 ,40
 40 IF(RL-RB(J))42,50 ,50
 41 RL=RA(J-1)
    XL=X (J-1)
    YL= Y(K-1)
    GO TO 40
 42 RL=RB(J)
    XL=X (J)
    YL=Y(K)
    GO TO 50
 50 RS=RR
    XS=XX
    YS=YY
    IF(RS-RA(J-1)) 52, 52,53
 52 IF(RS-RB(J)) 60,60,54
 53 RS=RA(J-1)
    XS=X (J-1)
    YS =Y(K-1)
    GO TO 52
 54 RS=RB(J)
    XS=X (J)
    YS=Y (K)
    GO TO 60
 60 RM=RR
    XM=XX
    YM=YY
    IF(RM-RS) 62, 62,61
 61 IF(RM-RL)70 ,62 ,62
 62 RM=RA(J-1)
    XM=X (J-1)
    YM=Y (K-1)
    IF(RM-RS) 64,64,63
 63 IF(RM-RL) 70,64,64
 64 RM = RB(J)
    XM=X (J)
    YM=Y (K)
 70 YCS=YS*YCONV
    YCM=YM*YCONV
    YCL=YL*YCONV
 71 YS=YS-YMAX+YMIN
    YM=YM-YMAX+YMIN
    YL=YL-YMAX+YMIN
 72  XCS=XS/SMAX
     XCM=XM/SMAX
```

```
      XCL=XL/SMAX
     RC = CNTRLO
  80 IF (RC.GT.CMAX )      GO TO 110
     IF ( RC .NE. RM ) GO TO 91
  81 IF ( RM .NE. RS ) GO TO 91
  82 IF ( RL .EQ. RM ) GO TO 100
  91 IF(RC-RS)100,95,92
  92 IF(RC-RM)96,93,94
  93 XPA=XCM
     YPA=YCM
     GO TO 99
  94 IF(RC-RL)106,103,110
  95 Q=0.0
     GO TO 97
  96 Q = (RC-RS)/(RM-RS)
  97 XPA = XCS-Q*(XCS-XCM)
     YPA = YCS-Q*(YCS-YCM)
  99 Q = (RC-RS)/(RL-RS)
     XPB = XCS-Q*(XCS-XCL)
     YPB = YCS-Q*(YCS-YCL)
     IF(RC)10115,10116,10116
10115 XPB1=0.5*(XPA+XPB)
     YPB1=0.5*(YPA+YPB)
     IF(ABS (XPA-XPB1)-.001)5001,5002,5002
 5001 IF(ABS (YPA-YPB1)-.001)100,5002,5002
 5002 CALL PLOTL(XPA,YPA,XPB1,YPB1)
     GO TO 100
10116 IF(ABS (XPA-XPB)-.001)5003,5004,5004
 5003 IF(ABS (YPA-YPB)-.001)100,5004,5004
 5004 CALL    PLOTL(XPA,YPA,XPB,YPB)
  100 RC = RC + CNTRVL
     GO TO 80
  103 XPA = XCL
     YPA = YCL
     GO TO 99
  106 Q=(RC-RM)/(RL-RM)
     XPA=XCM-Q*(XCM-XCL)
     YPA=YCM-Q*(YCM-YCL)
     GO TO 99
  110  GO TO L,(112,118)
  112  ASSIGN 118 TO L
     RR   =RB(J-1)
     XX   =X (J-1)
     YY   =Y (K)
     GO TO 37
  118 CONTINUE
      L=MOD(NCHAR,100)
      IF(L) 5006,5007,5006
 5006  CALL SYSSYM(1.0,0.0,0.15,WORDS,L,90.0)
      CALL SYSEND(0,0)
      RETURN
 5007  CALL SYSEND(0,0)
      RETURN
      END
C
C     THIS IS ALSO A MODIFIED CALTECH LIBRARY SUBROUTINE
      SUBROUTINE PLOTL(X1,Y1,X2,Y2)
      DIMENSION X(2),Y(2)
      COMMON/PLTDM/XLNGTH
      X(1)=X1*XLNGTH
      X(2)=X2*XLNGTH
      Y(1)=Y1*XLNGTH
      Y(2)=Y2*XLNGTH
```

```
      CALL SYSPLT(X(1),Y(1),3)
      CALL SYSPLT(X(2),Y(2),2)
      RETURN
      END
C
C     'CIRCLE' DRAWS NCIR CIRCLES CENTERED AT XCEN,YCEN INCHES.
C     THE NTH CIRCLE HAS RADIUS=RADMAX/NCIR.
C
      SUBROUTINE CIRCLE  (RADMAX,XCEN,YCEN,NCIR)
      DIMENSION X(1000),Y(1000)
      COMMON/COMPLO/ITEST,XLNGTH,YLNGTH
      ITEST=1
      XLNGTH=10.
      YLNGTH=10.
      RADIA2=3.1415963/120.
C     FOR PLOTTING SYMBOLS, COMPUTE 25 POINTS/CIRCLE
C     FOR DRAWING CIRCLES WITH RADIUS GT. 1'', COMPUTE 250 PTS./CIRCLE
      IF(RADMAX.LT.1.) RADIA2=3.1416/12.
      DO 100 I=1,NCIR
        FI=I
        FNCIR=NCIR
        RAD=RADMAX*FI/FNCIR
      DO 50 J=1,1000
          FJ=J
          THETA=RADIA2*(FJ-1.)/FI
      IF(THETA.GT.6.5) GO TO 55
          X(J)=RAD*COS(THETA) + XCEN
        Y(J)=RAD*SIN(THETA) + YCEN
      MAXJ=J
 50     CONTINUE
 55     CALL XYPLOT (MAXJ,X,Y, 0.,10.,0.,10.,0,0)
 100  CONTINUE
      RETURN
      END
C
C     'ROTATE' COMPUTES THE CARTESIAN COORDS (XROT,YROT) OF THE POINT
C     LAT=ALA,LONG=FI AFTER ROTATION TO AN AZ. EQ. PROJ. CENTERED AT
C     LAT=POLAT,LONG=POLONG. ALGORITHM FROM R. L. PARKER'S 'HYPERMAP'.
C
      SUBROUTINE ROTATE (ALA,FI,POLAT,POLONG)
      COMMON/ROTUV/XROT,YROT
      F=.0174533
      DEGRE=1./F
      ROT=180.
 899  SINRT=-COS(F*ROT)
      COSRT=SIN(F*ROT)
      SINLO=SIN(F*POLAT)
      COSLO=COS(F*POLAT)
      SINPH=SIN(F*(FI-POLONG))
      COSPH=COS(F*(FI-POLONG))
      SINLA=SIN(F*ALA)
      COSLA=SQRT(1.-SINLA*SINLA)
      COSA=SINLA*SINLO+COSLA*COSLO*COSPH
      SINA=SQRT(1.0001-COSA*COSA)
      SINB=COSLA*SINPH/SINA
      COSB=(SINLA*COSLO-COSLA*SINLO*COSPH)/SINA
      R=ATAN(SINA/COSA)
      U=R*(COSB*COSRT-SINB*SINRT)
      V=-R*(SINB*COSRT+COSB*SINRT)
      XROT=U*DEGRE+90.
      YROT=V*DEGRE+90.
 950  RETURN
      END
```