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Paleomagnetism of Late Pleistocene Sediments, Puget Lowland, Washington

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PALEOMAGNETISM OF LATE PLEISTOCENE SEDIMENTS, PUGET LOWLAND, WASHINGTON

A Thesis

Presented to

the Faculty of

Western Washington State College

In Partial Fulfillment

of the Requirements for the Degree

Master of Science

by

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ABSTRACT

Establishment of late Pleistocene geomagnetic polarity events would provide useful time indices for regional and interregional stratigraphic correlations. Development of a geomagnetic polarity scale tied to radiocarbon dating for the Puget Lowland helps to elucidate the possible occurrence of world-wide (dipole) reversed events during the last 50,000 years. Stability and reliability tests indicate that late Pleistocene sediments exposed in the Puget Lowland record valid paleomagnetic directions. Puget Lowland geomagnetic polarity for the interval 11,000 years to 30,000 years BP was normal except for a short reversed period between 20,000 years and 15,000 years BP. In addition to sampling errors and magnetic remanence directional errors, variations in paleomagnetic directions over short time periods may be due to secular variation. The brief period of reversed polarity could have been due to the non-dipole field rather than the dipole field. Interregional comparison of available geomagnetic polarity results for the last 50,000 years does not establish the occurrence of a dipole reversed event. Recorded reversed paleomagnetic directions during the last 50,000 years may represent localized geomagnetic excursions. Resolution of the details of geomagnetic behavior could result in the development of regional geomagnetic time scales for the late Quaternary.

Kurt Othberg

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ACKNOWLEDGEMENTS

I am grateful to Dr. Don J. Easterbrook for suggesting the project and for invaluable assistance and encouragement during sampling and writing. Thanks are due Dr. Myrl Beck for providing the paleomagnetic laboratory and materials and for critically reading the manuscript. The suggestions of Dr. Dave Rahm and Mr. John Whitney, who read the manuscript critically, are gratefully acknowledged. A special thanks goes to my wife, Penny, for encouragement and for typing the final draft.

INTRODUCTION

Purpose

Establishment of a time scale for the Quaternary based on geomagnetic polarity has provided a useful tool for regional and world-wide stratigraphic correlations. However, geomagnetism has so far not been useful for studies of the last 700,000 years because no well-defined geomagnetic events within the Brunhes normal epoch have been recognized. When compared statistically with other polarity epochs, the probability is high that short polarity events should have occurred during the 700,000 year-long Brunhes epoch (Cox, 1968; 1969). Two reversals of the geomagnetic field during the late Pleistocene have been studied: (1) the Blake reversal at approximately 110,000 years before present (BP) (Smith and Foster, 1969), and (2) the Laschamp reversal between 50,000 years and 8,700 years BP (Bonhommet and Babkine, 1967; Bonhommet and Zahringer, 1969). Whether or not the Blake and Laschamp reversals should be classified as true world-wide polarity events remains to be demonstrated. If these reversals, measured in lavas and sediments, represent dipole field reversals, then they may be used as world-wide time indices for late Pleistocene stratigraphy.

This investigation was confined to an examination of geomagnetic polarity within the time range of accurate radiocarbon dating, i.e., the last 50,000 years BP, and therefore, bears only on the existence of the Laschamp reversal as a world-wide polarity event.

The purpose of this research was fourfold: (1) to determine whether late Pleistocene fine-grained sediments exposed in the Puget Lowland record valid paleomagnetic directions, (2) to begin establishment of a geomagnetic polarity time scale for the Puget Lowland tied to radiocarbon dating,

(3) to provide information about short term variations in the geomagnetic field, and (4) to make comparisons with polarity results from elsewhere in the world for possible elucidation of the existence of world-wide polarity events.

Previous Work

No previous investigations of the paleomagnetism of the late Pleistocene sediments in the Puget Lowland are known. However, some late Pleistocene paleomagnetic data are available from Europe, North America, the Pacific, Atlantic and Indian Oceans, the Mediterranean Sea, the Gulf of Mexico and the Bering Sea.

Paleomagnetic results from this investigation are fit into a previously established late Pleistocene stratigraphic and chronologic framework for the Puget Lowland (Crandell and others, 1958; Easterbrook, 1963; Armstrong and others, 1965; Mullineaux and others, 1965; Easterbrook and others, 1967; Easterbrook, 1969).

Geologic Setting

The Puget Lowland lies in the Puget Trough between the Cascade Range on the east and the Olympic Mountains and Vancouver Island on the West (Fig. 1). The lowland is an elongate structural trough, modified by Pleistocene deposition and erosion, lying within approximately 500 feet of present sea level. The western portion of the lowland is composed of marine waterways between numerous peninsulas and islands. The eastern portion consists mostly of broad alluvial valleys. A large part of the region is directly underlain by unconsolidated Pleistocene deposits. Excellent exposures of these deposits may be found along many miles of sea cliffs, numerous road cuts and gravel excavations.

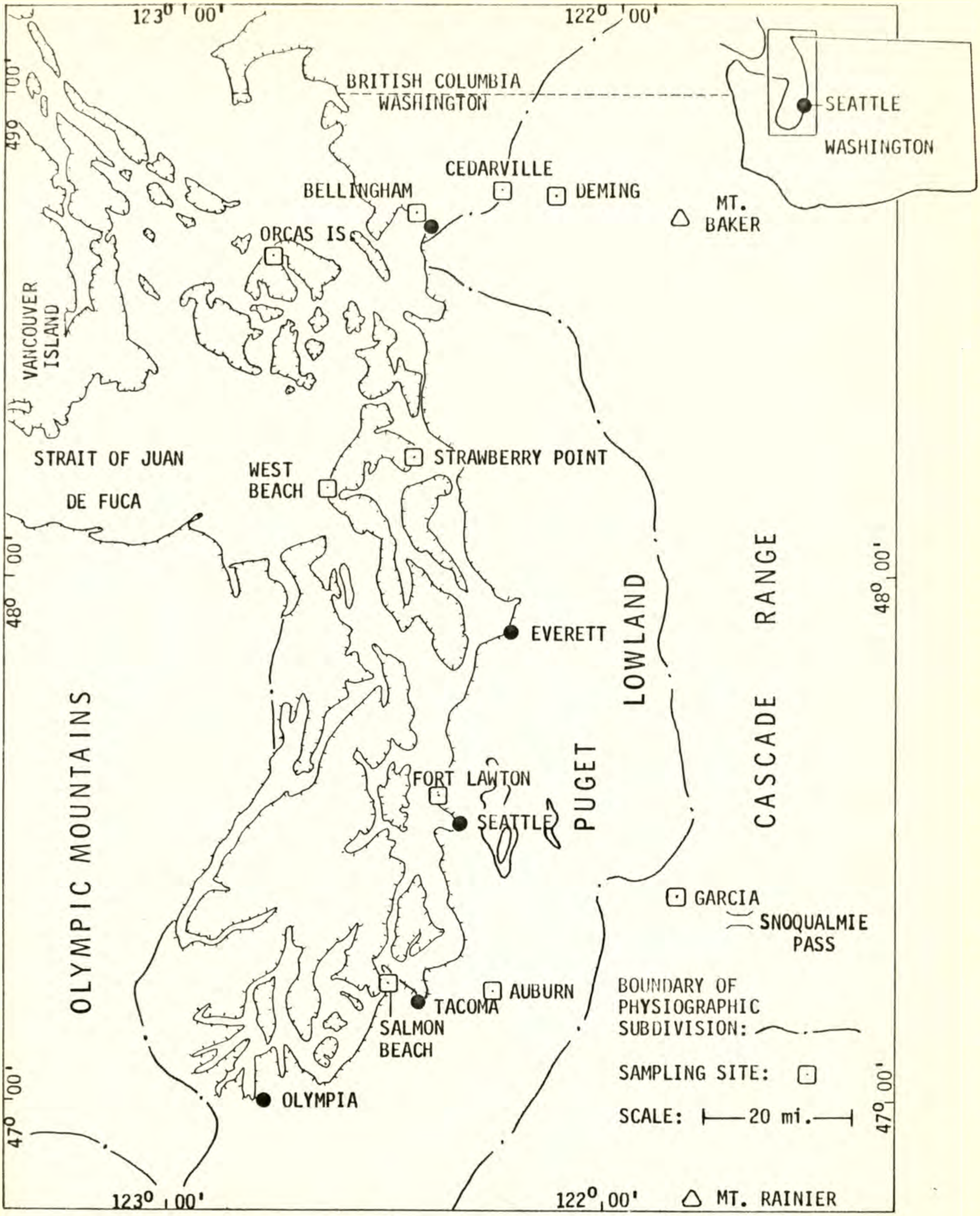


Figure 1. Index map of the Puget Lowland showing site locations.

The Puget Lowland was glaciated several times by Pleistocene continental glacier lobes extending southward from British Columbia. The stratigraphy of the unconsolidated deposits is therefore directly related to episodic glaciations. This investigation utilized sediment deposited just prior to, during and in the waning stages of the last major continental glaciation (Fig. 2).

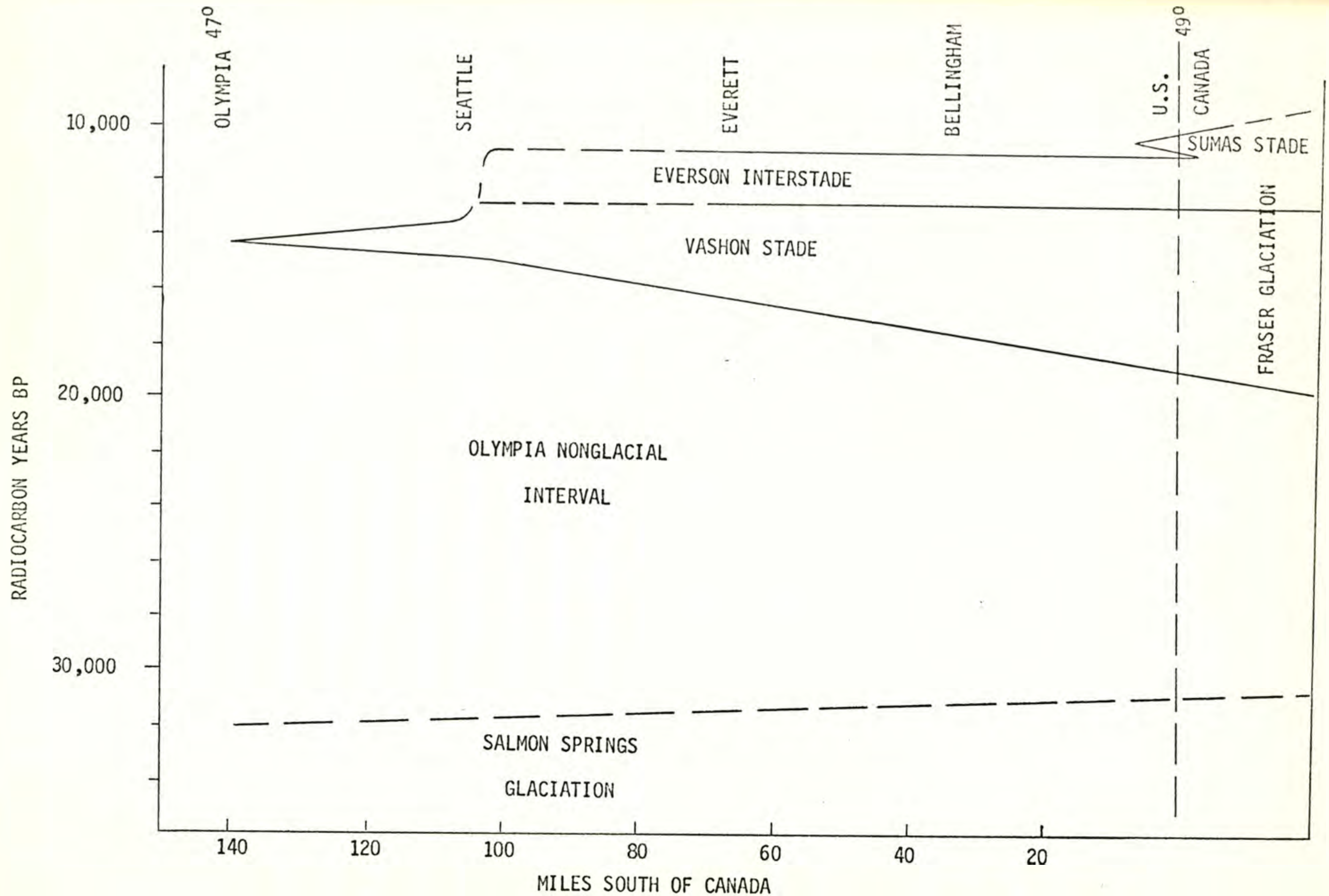


Figure 2. Late Pleistocene glacial and nonglacial events in the Puget Lowland (adapted from Easterbrook, 1970).

VALIDITY OF MEASUREMENTS

The selection of sediment types for paleomagnetic measurements is vital to the validity of the results. Published empirical and theoretical studies indicate that in order to acquire a detrital remanent magnetization (DRM) a sediment deposited in water must have a grain size of approximately 50 microns or smaller, i.e., silt and clay. The majority of sites in this investigation consisted of silts and clays of lacustrine and fluvial origin. However, sediment at sites representing the Everson Interstade consisted of glaciomarine drift, a poorly sorted, pebbly, silty clay.

Detrital Remanent Magnetization

The results of what may be the first systematic measurements of the natural remanent magnetization of sediments were published in 1938 by McNish and Johnson. They measured the remanent magnetization of New England post-glacial varved clays. Similar measurements were made on Swedish varves by Ising in 1942. He recognized that the remanence was due to the alignment of magnetic particles comprising a small fraction of the sediment. Using artificially deposited magnetic particles, Benedict (1943) demonstrated the ability of small particles to align themselves in a magnetic field. Further laboratory work (Johnson and others, 1948) demonstrated the importance of gravitational coupling and hydrodynamic forces in determining the perfection of alignment of magnetic grain moments with an ambient magnetic field direction. Studies in the 1950's and 1960's culminated with the establishment of models of detrital remanent magnetization (DRM) (Griffiths and others, 1960; Nagata, 1961, 1962; Rees, 1961; King and Rees, 1966). Stacey (1972) has reviewed the variables that determine the perfection of DRM acquisition.

Unconsolidated sediments useful to studies of the geomagnetic field are water deposited clays and silts. Magnetite is apparently the magnetic particle most responsible for DRM in sediments of Pleistocene age (King and Rees, 1966).

Contributing to the degree of DRM acquisition in a clay or silt are 1) systematic factors: ambient magnetic field, gravity, and hydrodynamic forces; and 2) randomizing factors: Brownian motion, initial randomness, and particle inhomogeneity (King and Rees, 1966). The net effect of these factors produces the relative perfection of alignment of magnetic grain moments with the direction of the ambient field.

The randomizing factors serve to decrease bulk sediment magnetic intensity and increase the scatter of DRM directions between samples. If the randomizing factors become dominant, DRM ceases to exist. However, the importance of the randomizing factors to DRM is grain-size dependent.

Because a range of grain sizes may exist in a sediment, size limits may be determined, within which magnetic grains will orient themselves parallel to an ambient magnetic field. The lower size limit is controlled by the effect of Brownian motion. Although insignificant in larger grain sizes, Brownian movement increases so rapidly with decreasing particle size that grains of 0.1 microns or less probably do not make any appreciable contribution to remanence in a sediment. This may be the ultimate lower limit for grain sizes important in DRM (King and Rees, 1966), although this critical size limit varies with the earth's ambient magnetic field intensity (Stacey, 1972).

Initial randomness and particle inhomogeneity serve to disturb grain orientations of larger grains, and therefore, are important to the determination of an upper grain size limit for DRM. Particle inhomogeneity

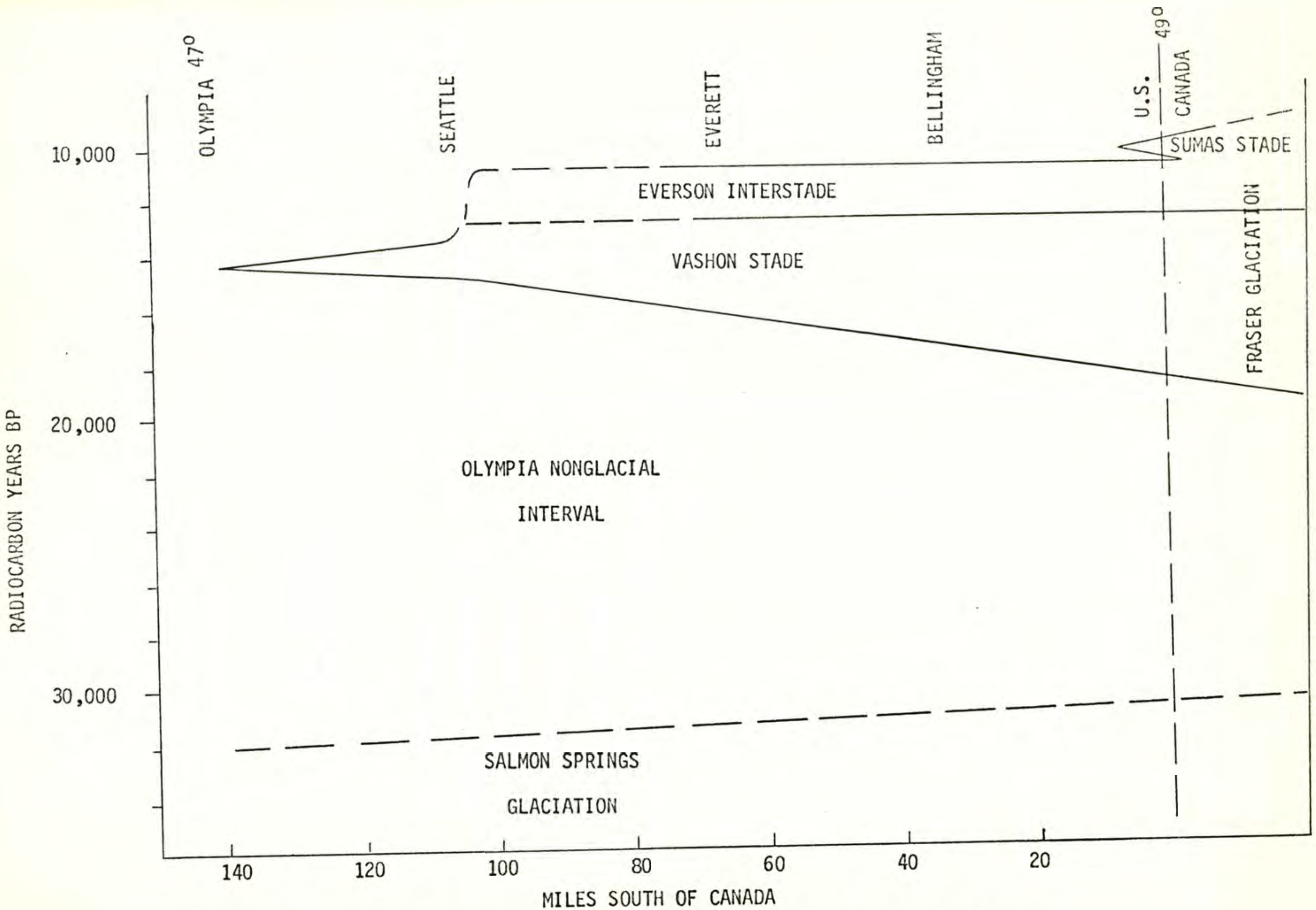


Figure 2. Late Pleistocene glacial and nonglacial events in the Puget Lowland (adapted from Easterbrook, 1970).

is the dislocation of the center of gravity from the grain geometric center. It results in the tendency of a grain to rotate in response to torques produced when net hydraulic drag does not pass through the center of gravity of a grain. Randomization due to particle inhomogeneity may be important for grain sizes greater than 10 microns (King and Rees, 1966). Initial randomness results from the chaotic motion of turbulent water. The degree of turbulence in water necessary to suspend a grain is proportional to the square of the grain radius. It should be possible, then, to determine the grain size above which the depositing body of water was too turbulent for acquisition of an accurate DRM. King and Rees (1966) and Stacey (1972) have shown that any size magnetite grain can orient parallel to an ambient field in approximately one second, given no disturbing forces. Most alignment probably occurs within a thin zone of no flow or laminar flow at the bottom of a body of water. However, the laminar flow layer is not thick enough to allow large, rapidly falling grains time to orient.

Based on this information, King and Rees (1966) estimated that the critical grain diameter for which the effect of turbulence significantly disrupts the acquisition of accurate DRM lies between 10 microns and 180 microns. Assuming that silt particles (62 microns) relate to fluid stress in the same way as sand particles (62 microns-2mm), the laminar layer would be too thin for magnetic orientation of grains over 50 microns in size. Supporting evidence of this critical size determination comes from Keen's (1963) study of turbidite graded sequences, in which consistent DRM directions were obtained only from the less than 40 micron sediments.

In addition to minimal effects of the randomizing factors, perfect alignment of magnetite grains parallel to an ambient magnetic field may only occur when the systematic factors are dominated by magnetic torques.

As gravitational and laminar hydrodynamic forces (currents) become increasingly important in the determination of the orientation of magnetic particles, systematic deviations from the ambient field direction occur and the DRM direction is not parallel to the ambient field direction. The deviations become more pronounced as the grain size of the particles increases. The observed inclination deviations (Johnson and others, 1948; Griffiths and others, 1960; Irving, 1964) have been attributed to gravitationally induced rolling when a particle touches bottom. Gravitational rolling does not deflect the declination since the average of the grain moments will still be parallel to the declination of the ambient magnetic field. If, however, the laminar flow layer has a current of between 5-30 cm/sec DRM declinations may be deflected up to 10^{o} as a result of hydraulic drag upon non-spherical magnetite grains (Rees, 1961; Irving, 1964). Rees (1965) and King and Rees (1966) have shown that the gravitational and hydrodynamically induced errors in DRM are grain size dependent, such that magnetite particles on the order of 1 micron in size may acquire a nearly deviation-free DRM.

Theoretically and empirically, sediments of silt size and smaller contain magnetite partcles which are capable of acquiring a reliable DRM. Even if the depositional process did not dictate a critical upper grain size for DRM acquisition, the grain size dependence of rock magnetism does. The grains that are thought to carry the "hard" remanence are very small in size. The range in size of single domain and pseudo-single domain grains that comprise the "hard" remanence is the critical factor. Although critical single and pseudo-single domain sizes are very shape dependent, single domain behavior has been observed for elongate grains up to 17 microns in length (Evans and others, 1968). Considering relative sorting during deposition due to differences in specific gravities, one would expect a coarse silt sediment (median diameter of the predominant, silicate grains) to contain magnetite particles which behave as pseudo-single domain grains, producing a highly coercive remanence.

In summary, for water-deposited sediments, the acquisition of a reliable DRM can be considered to be restricted to silts and clays. These sediments may be poorly sorted (Griffiths, 1951) such that a range of sizes of magnetite grains exist in the sediment, some of which lie within the critical size range for DRM acquisition. Althouth the critical size range for DRM will change with ambient field intensity (Stacey, 1972), magnetite grains of sizes larger and smaller than the critical size range may only represent a randomized component of magnetization. Glaciomarine drift is perhaps the best example of how the critically sized magnetite grains may control the DRM while a significant portion of larger grains contribute only a randomized component of magnetization. (see Reliability)

Stability of Remanent Magnetization

A standard stepwise alternating field demagnetization was used in assessing magnetic stability of twenty samples of silts, clays and glaciomarine drift. Results from four samples are shown in Figures 3 and 4. Remanent magnetization is comprised of many individual magnetic moments. If the "hard" magnetic moments are dominant over the "soft" magnetic moments, the remanent magnetization will be stable and easily measured.

Typically, the silts, clays and glaciomarine drifts sampled for this investigation have high coercivity magnetic components. This is shown by the high demagnetization levels required to randomize the magnetic moments. A sterographic plot of paleomagnetic directions for each demagnetization level (Figs. 3 and 4) demonstrates that alternating frequency demagnetization levels of 400 - 800 oe. are required to destroy the

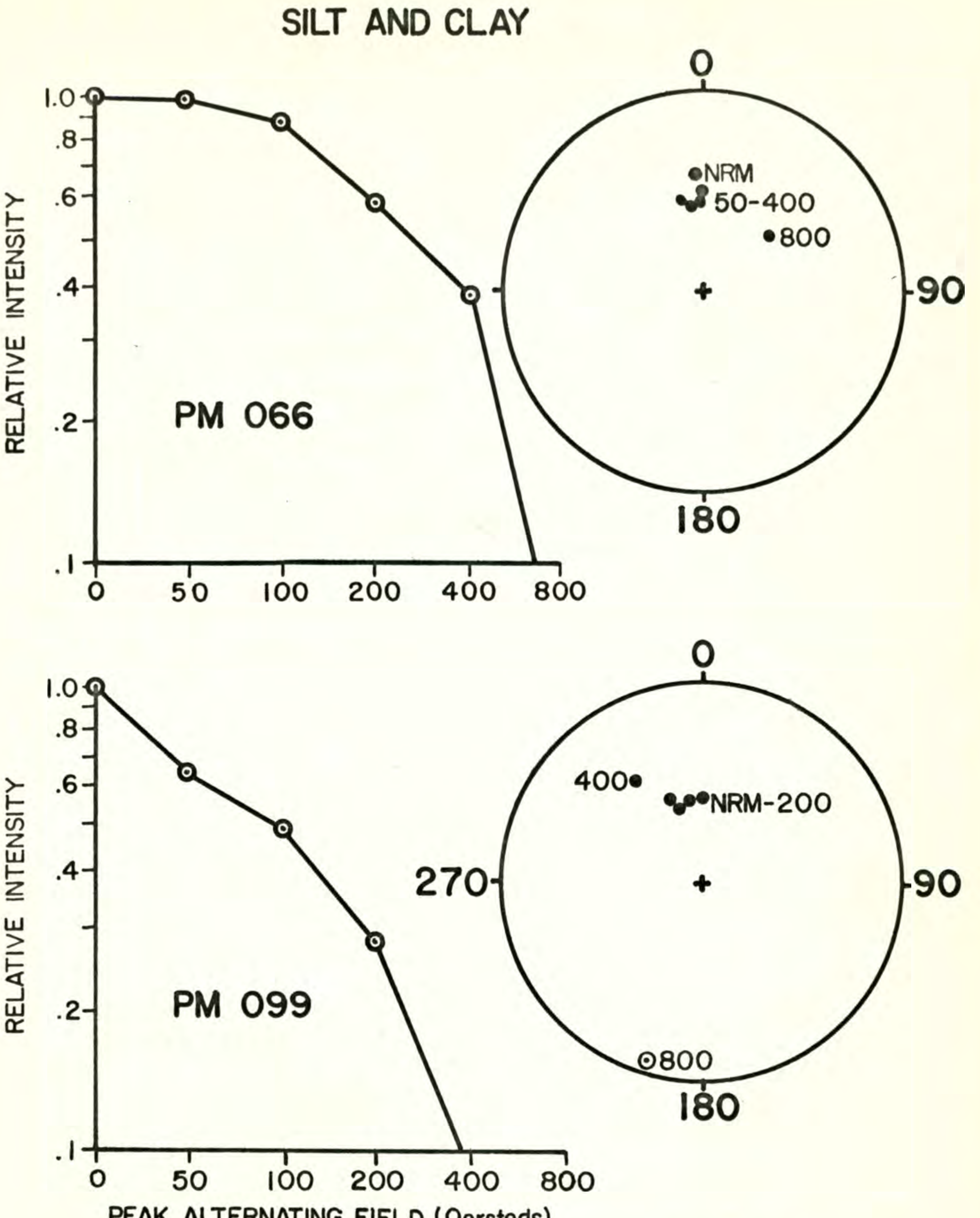


Figure 3. Stability of paleomagnetic directions of silt and clay. Normalized intensities of magnetization plotted against peak a.f. demagnetization field; and stereographic projections showing corresponding remanent magnetization directions.

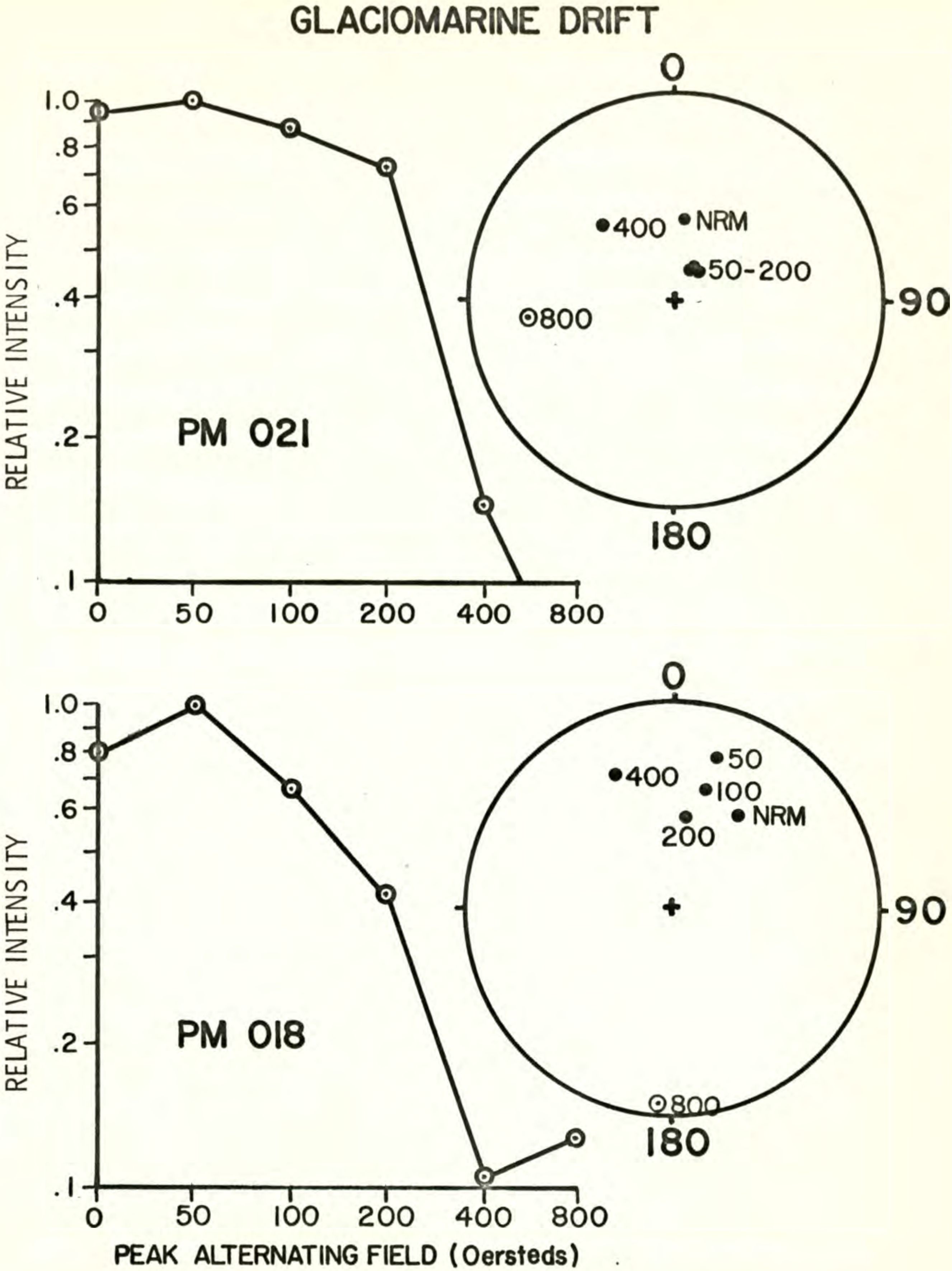


Figure 4. Stability of paleomagnetic directions of glaciomarine drift. Normalized intensities of magnetization plotted against peak a.f. demagnetizing field; and stereographic projections showing corresponding remanent magnetization directions.

remanent direction. Usually, a 100 - 200 oe. demagnetization level appeared to randomize "soft" magnetic components, while at least 50% of the natural remanent magnetization (NRM) remained. Based on this finding, all samples were cleaned at 100 oe. or 200 oe.

Studies of the magnetic properties of similar sediment by other investigators (Johnson and others, 1948; Griffiths and others, 1960; Denham and Cox, 1971), indicate that the strong, stable remanences are due to magnetite detritus from igneous rocks. The remanent magnetizations measured in this investigation are similarly strong and stable, and source areas for magnetite detritus are plentiful in the mountain ranges adjoining the Puget Lowland.

Reliability of Remanent Magnetization

A second requirement for the validity of the paleomagnetic directions is that they are reliable, that is, the detrital remanent magnetizations are reproducible measurements of the geomagnetic field at the time of deposition. That DRM's yield a declination identical to an external field has been fully demonstrated by Benedict (1943), Johnson and others (1948) and Griffiths and others (1960). Two methods were used to check reliability of specimens used in this investigation. Firstly, by sampling vertical intervals that represent short periods relative to the known variations in the geomagnetic field, the paleomagnetic directions can be expected to vary within known limits. In a vertical (time) direction, declinations of successive samples typically varied $\pm 20^{\circ}$, but as high as $\pm 45^{\circ}$. These variations are consistent with the known effects of the drifting non-dipole field (Stacey, 1969). Secondly, a test of reliability was performed on silt and glaciomarine drift. This consisted of a check on the internal consistancy of contemporaneously deposited samples. Nine to ten samples from one site were collected along a single horizon where the depostition of all samples should have been contemporaneous, recording precisely the same paleomagnetic directions. The NRM and "cleaned" directions of each sample, for each sediment type were plotted stereographically (Figs. 5 and 6). Fisher's (1953) statistics were calculated in order to provide a quantitative measure of the reliability. The NRM directions and "cleaned" directions both show tight groupings as exemplified by small circles of confidence (α 95's of 7.1^{o} and 4.2^{o}) and high precision parameter (k) values (k's of 54 and 134). The decreased dispersion after cleaning indicates further that a small component of "soft" magnetic moments is removed at 200 oe.

The stability and reliability tests performed on these sediments demonstrate that the paleomagnetic directions measured are valid. More specifically, the tests indicate that (1) variability in direction between contemporaneous samples may be due mostly to sampling and laboratory errors, (2) any one paleomagnetic declination probably is a measure of the declination of the geomagnetic field at the time of deposition, and (3) although the sand and pebble grains present in glaciomarine drift might be expected to cause variable magnetic directions, glaciomarine drift does preserve reliable paleomagnetic directions.

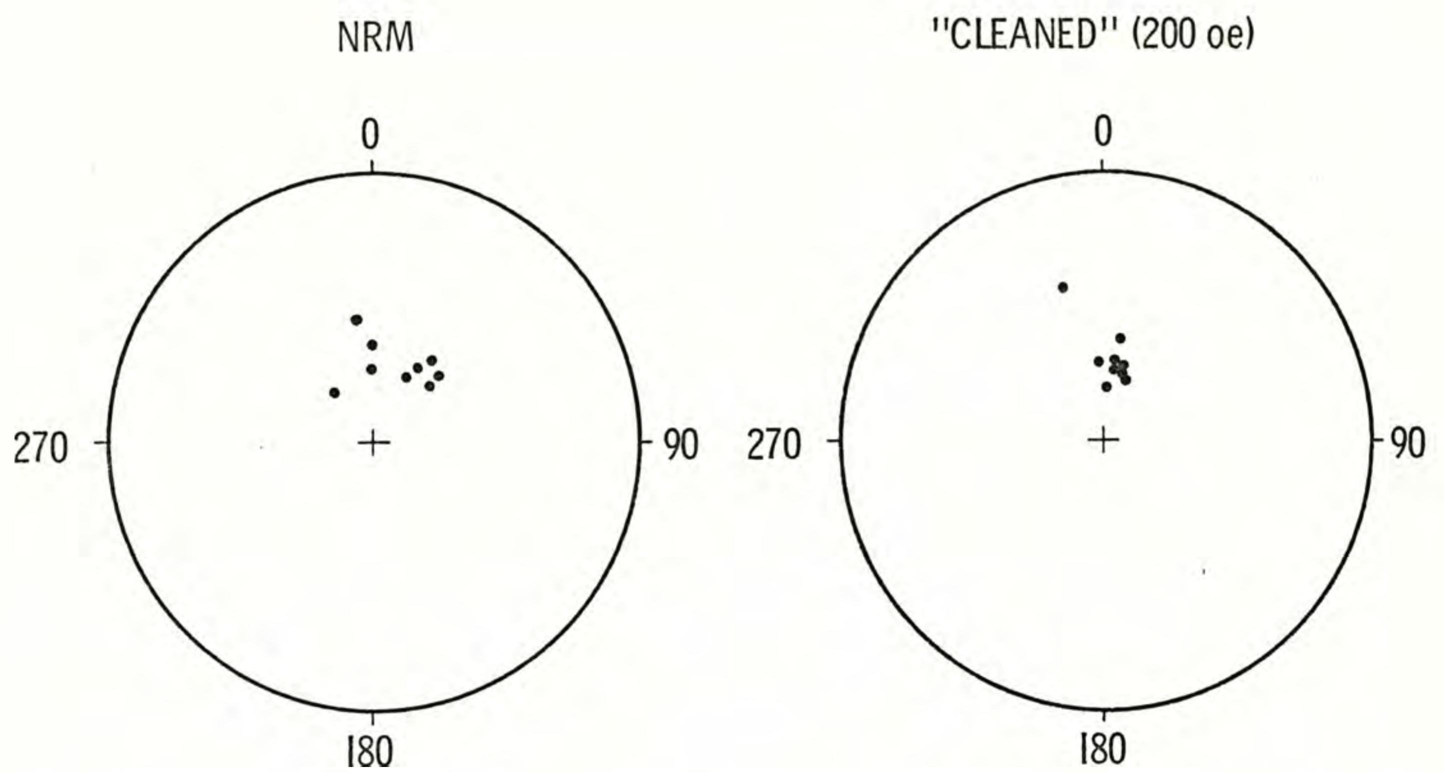


Figure 5. Reliability of paleomagnetic directions in silt; nine samples from one horizon. Stereographic projection of the remanent magnetization directions.

Statistics ("cleaned"): Circle of confidence (95%) = 7.1°
Angular standard deviation = 10.4°
Precision parameter = 54

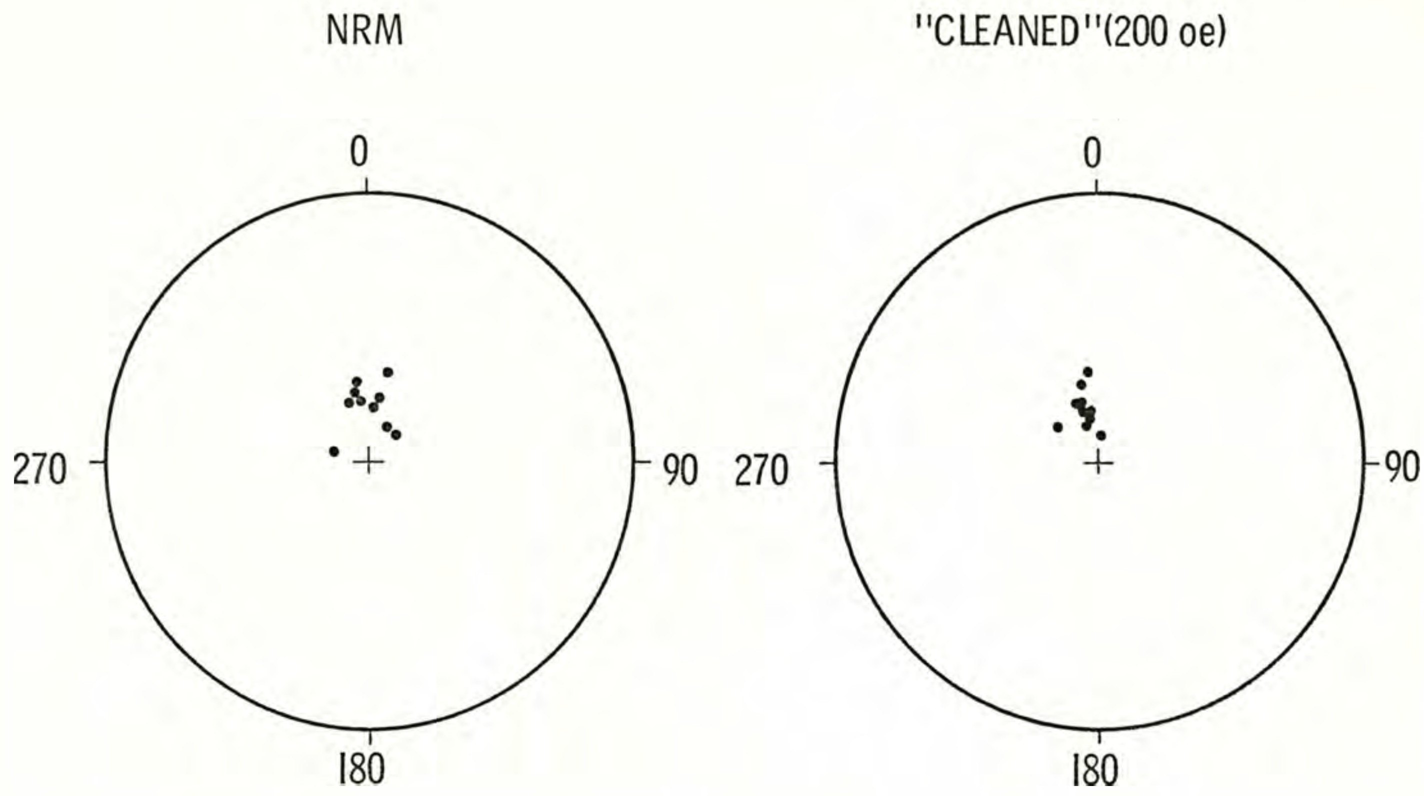


Figure 6. Reliability of paleomagnetic directions in glaciomarine drift; ten samples from one horizon. Stereographic projection of the remanent magnetization directions.

Statistics ("cleaned"): Circle of confidence (95%) = 4.2°
Angular standard deviation = 6.7°
Precision parameter = 134

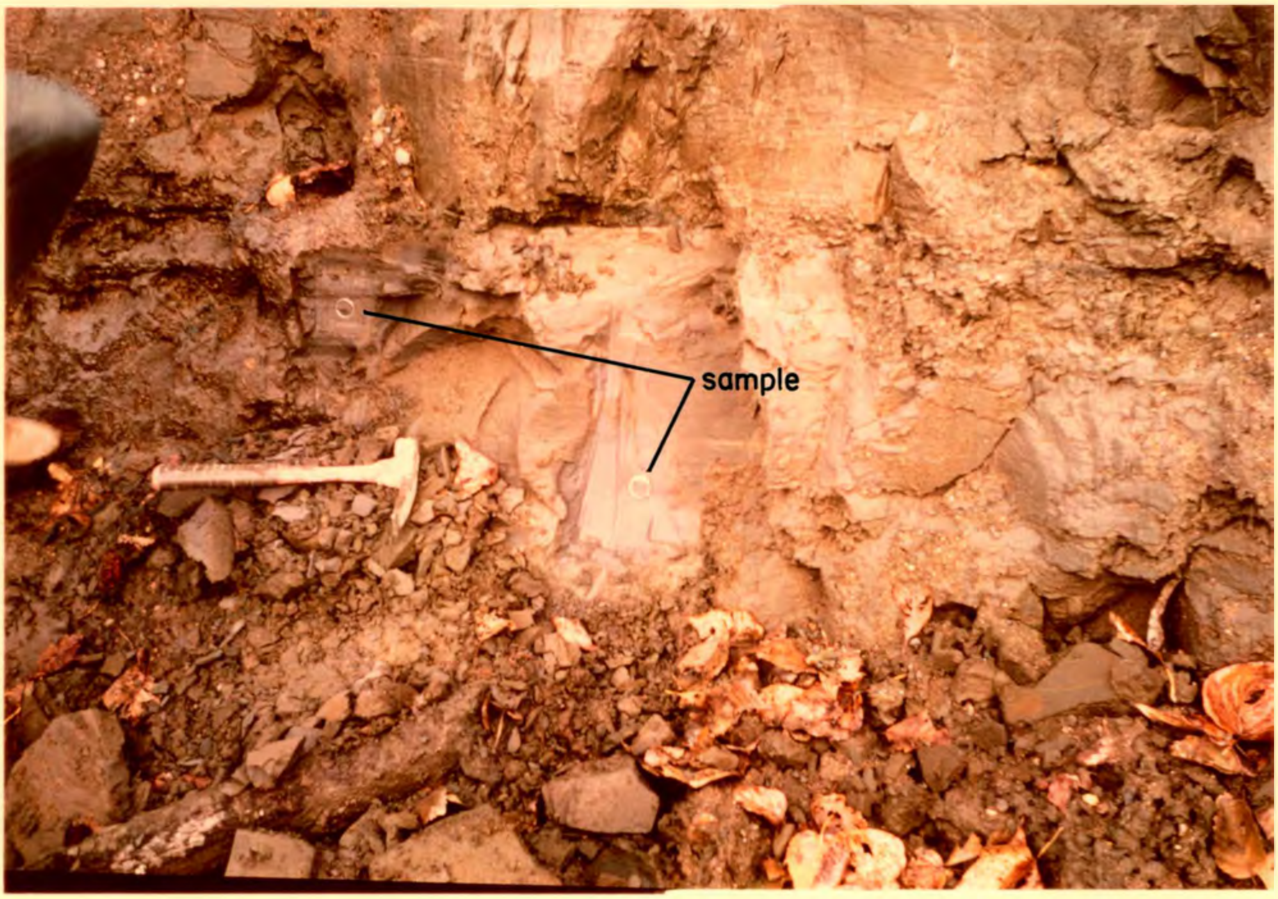
SAMPLING AND MEASUREMENTS

Procedures

Samples were collected only from undisturbed sediment, obtained by removal of outer, modified materials, and avoidance of slumped areas. Two methods of sampling were employed. 1) One-inch by one-inch sediment cores were carved from an oriented slab, and retained within a non-magnetic cylinder which was sealed in the lab. The orientation of the X, Y and Z axes of these cores (Fig. 7) was coincident with north, east and down directions respectively. 2) A non-magnetic cylinder of the proper size was driven into freshly exposed sediment in any orientation. The orientation of the core axes was, therefore, not coincident with north, east and down. In order to later correct for the discrepancy, the orientation of the X, Y and Z axes with respect to geographic north and vertical was measured with a Brunton compass and recorded. Essentially, the orientation recorded was the dip of the Z axis of the core, and the compass direction of the dip.

The remanent magnetization of the cores was measured using a Schoenstedt SM-1A spinner magnetometer. The intensity of magnetization parallel to each of the three core axes was measured in both the positive and negative directions. The resultant magnetization direction and intensity were calculated using a computer program. The computer program, in addition, calculated the corrected resultant direction for those samples whose axes were not oriented coincident with north, east and down in the field.

For most samples, two remanent magnetizations were measured, the natural remanent magnetization (NRM) and the remanent magnetization remaining after either 100 oersted (oe.) or 200 oe. alternating frequency



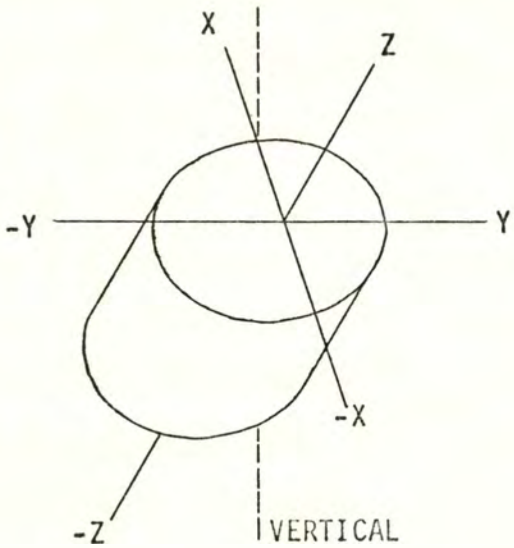


Figure 7. Sample core orientation before removal.

demagnetization, commonly referred to as "cleaning". The resulting data is contained in Appendix A.

The demagnetization was accomplished by tumbling each sample in an alternating polarity magnetic field which slowly decreased to zero. The equipment is designed to effectively randomize the magnetic moments of those grains which are less coercive than the "cleaning" level chosen. About twenty samples were "cleaned" at 50, 100, 200, 400 and 800 oe. each. This provided data upon which was based the choice of a 100 - 200 oe. "cleaning" level for all samples. For those samples which recorded normal geomagnetic directions, the NRM's differed little from the "cleaned" directions. However, the fact that the samples with reversed directions often changed markedly after "cleaning" was important evidence that probably all samples should be "cleaned". The interpretation of the ancient geomagnetic field directions was based on the "cleaned" remanent magnetizations.

Site Selection and Results of Measurements

A prime concern in this study was to sample an essentially continuous column of fine-grained sediment deposited during the past 50,000 years. The selection of sites was, therefore, based on two criteria: (1) well established stratigraphic sections containing organic material suitable for radiocarbon dates, and (2) stratigraphic sequences comprised of sediment suitable for paleomagnetic measurements.

The late Pleistocene stratigraphic sequences have been worked out for most of the Puget Lowland and a large number of radiocarbon dates have already been obtained from many localities in the region (Mullineaux and others, 1965; Armstrong and others, 1965; Easterbrook, 1969), thus providing an excellent framework for paleomagnetic study. Ten sites were selected

(Fig. 1), that were composed of sediments representing three climatic episodes during the late Pleistocene in the Puget Lowland. The climatic episodes, from youngest to oldest, are (1) the Everson Interstade, (2) the Vashon Stade of the Fraser Glaciation, and (3) the Olympia Nonglacial Interval (Fig. 2).

Everson Interstade

Sediments of the Everson Interstade at the type locality along the Nooksack River near Cedarville consist of two glaciomarine drift units separated by interbedded sand and clay (Fig. 8). Radiocarbon dates from Everson sediments range in age from about 10,400 years BP to approximately 13,000 years BP.

Samples were also collected from correlative exposures of Everson Glaciomarine Drift at Orcas Island, West Beach (Whidbey Island), Deming and Bellingham, Washington. Radiocarbon dates were available from each of these localities. Results of the twenty-two paleomagnetic measurements using Everson sediments are shown in Fig. 9.

Vashon Stade

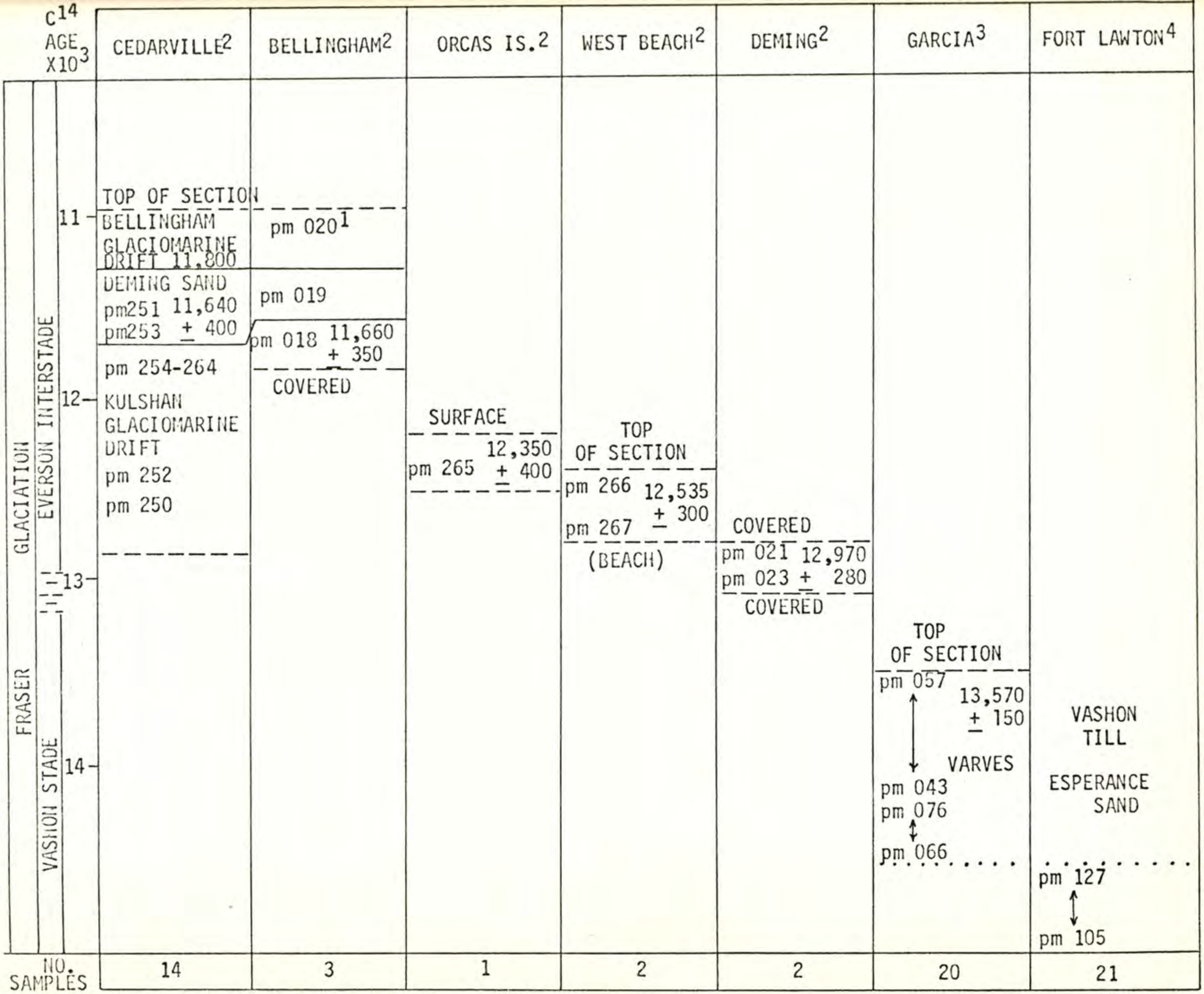
Ice-dammed lake silts of the Vashon Stade, representing the interval 13,500 to 15,000 years BP were sampled for paleomagnetic measurements. The stratigraphic sections, radiocarbon dates and sample locations are shown in Fig. 8. Results of the forty-one paleomagnetic measurements are shown in Fig. 10.

Two localities were selected that probably represent sequential episodes of deposition during the advance of Vashon ice close to and over the Seattle area: (1) varved sediments, exposed by downcutting of the Snoqualmie River at Garcia, Washington, radiocarbon dated at 13,600 years BP (Fairhall and others, 1966) and probably spanning the time during which

Figure 8. Sample localities, stratigraphy and radiocarbon dates for Everson and Vashon sediments. All dates were determined by Isotopes, Inc., the United States Geological Survey radiocarbon lab, the University of Washington radiocarbon lab, and the Western Washington State College radiocarbon lab. Limits of error were determined in the following ways:

Isotopes: one standard deviation.
U.S.G.S.: slightly larger than one standard deviation, taking into account known laboratory factors which affect uncertainty.
U. of W.: one standard deviation.
W.W.S.C.: based on one standard deviation, but often larger, allowing for laboratory factors affecting uncertainty.

[1] Paleomagnetic sample identification number.
[2] Easterbrook (1963; 1969; and unpublished data)
[3] Fairhall and others (1966)
[4] Mullineaux and others (1965)



C^{14} AGE $X10^3$
CEDARVILLE[2]
BELLINGHAM[2]
ORCAS IS.[2]
WEST BEACH[2]
DEMING[2]
GARCIA[3]
FORT LAWTON[4]
FRASER GLACIATION
EVERSON INTERSTADE
VASHON STADE
11
12
13
14
TOP OF SECTION
BELLINGHAM GLACIOMARINE DRIFT 11,800
DEMING SAND
pm251 11,640
pm253 ± 400
pm 254-264
KULSHAN GLACIOMARINE DRIFT
pm 252
pm 250
pm 020[1]
pm 019
pm 018 11,660 ± 350
COVERED
SURFACE
pm 265 12,350 ± 400
TOP OF SECTION
pm 266 12,535
pm 267 ± 300
(BEACH)
COVERED
pm 021 12,970
pm 023 ± 280
COVERED
TOP OF SECTION
pm 057
13,570 ± 150
VARVES
pm 043
pm 076
pm 066
VASHON TILL
ESPERANCE SAND
pm 127
pm 105
NO. SAMPLES
14
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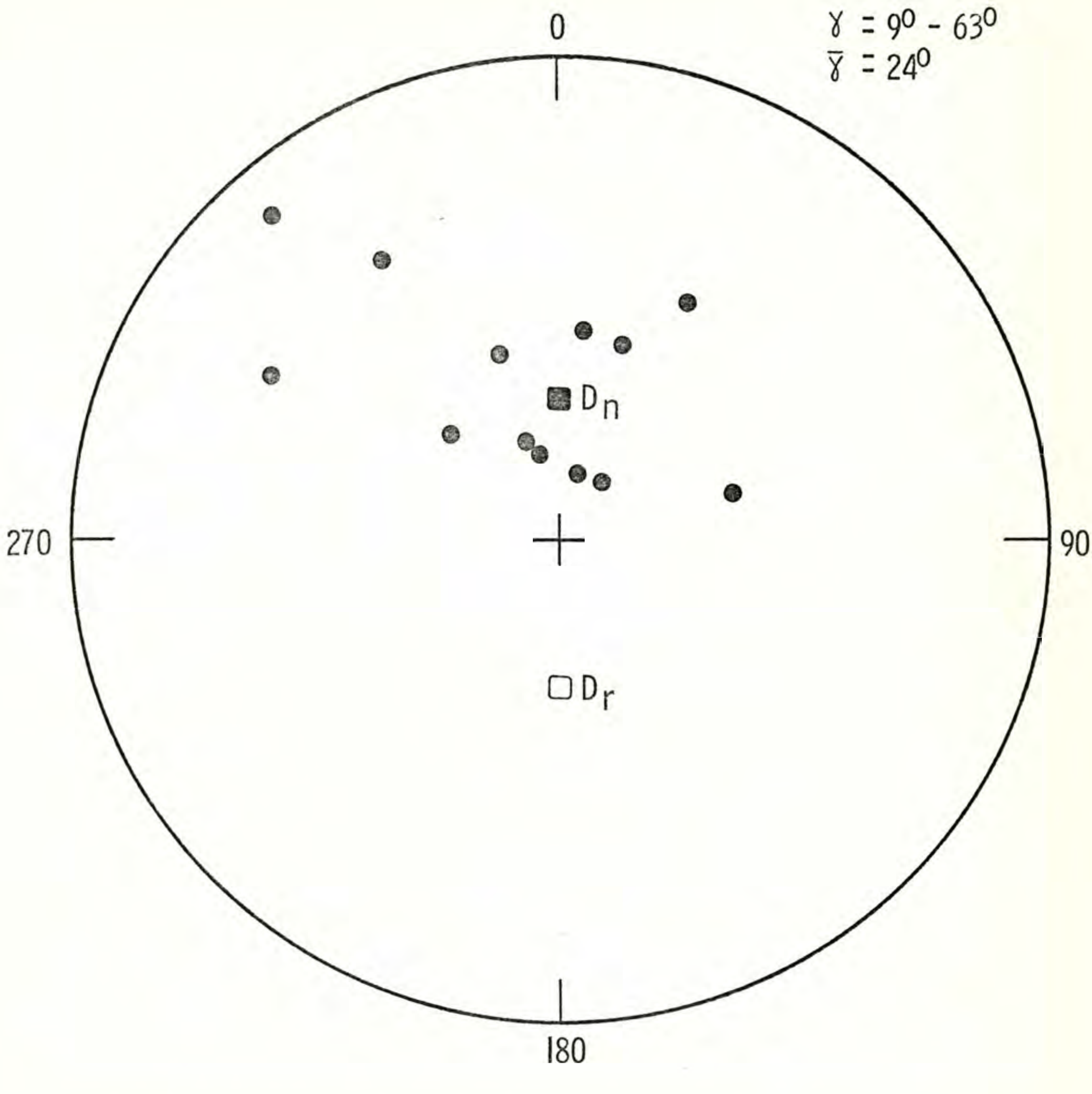


Figure 9. Paleomagnetic directions in sediments of Everson age (≅ 11,000 - 13,000 years BP). Stereographic projection in which downward-pointing inclinations (normal polarity) are plotted with closed circles, and upward-pointing inclinations (reverse polarity) are plotted with open circles. D_n and D_r are the axial centered dipole directions for the Puget Lowland, normal and reverse polarity, respectively. γ is the divergence from D_n or D_r. $\bar{\gamma}$ is the mean divergence.

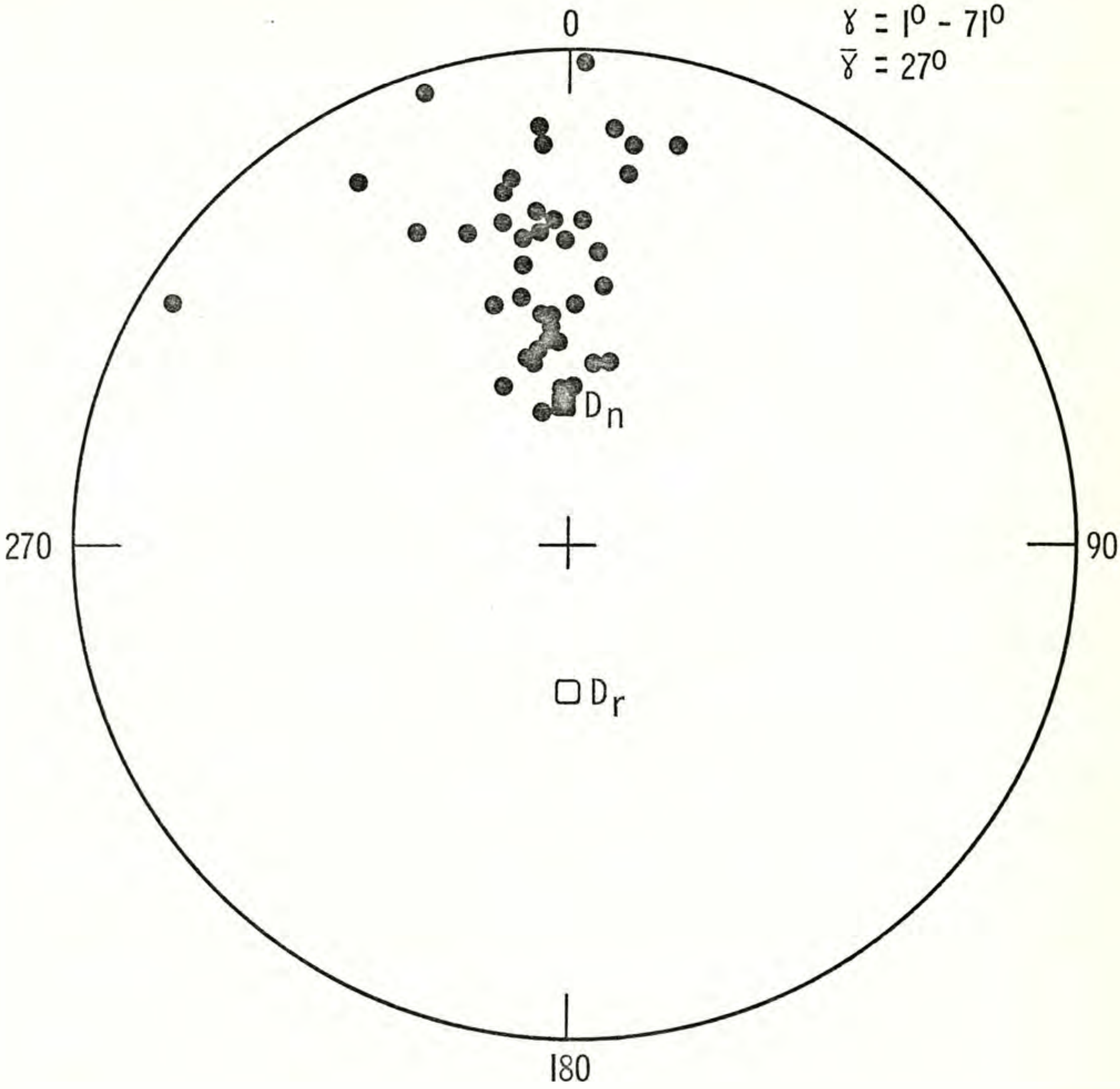


Figure 10. Paleomagnetic directions in sediments of Vashon age (≅ 13,500 - 15,000 years BP). Explanation of symbols as in Figure 9.

the Seattle area was covered by Vashon ice, and (2) the Lawton Clay, pro-glacial lake sediments exposed at Fort Lawton. Radiocarbon dates indicate that the Seattle area was covered by ice from approximately 14,500 years to approximately 13,500 years BP, and that the youngest Olympia age sediments are 15,000 years old (Rigg and Gould, 1957; Mullineaux and others, 1965). Vashon ice passed over the Seattle area after 14,500 years BP and had disappeared by about 13,500 years BP. The Fort Lawton locality was probably overridden at about the same time that continental ice dammed the Snoqualmie River and lake sediment was deposited at Garcia.

Olympia Nonglacial Interval

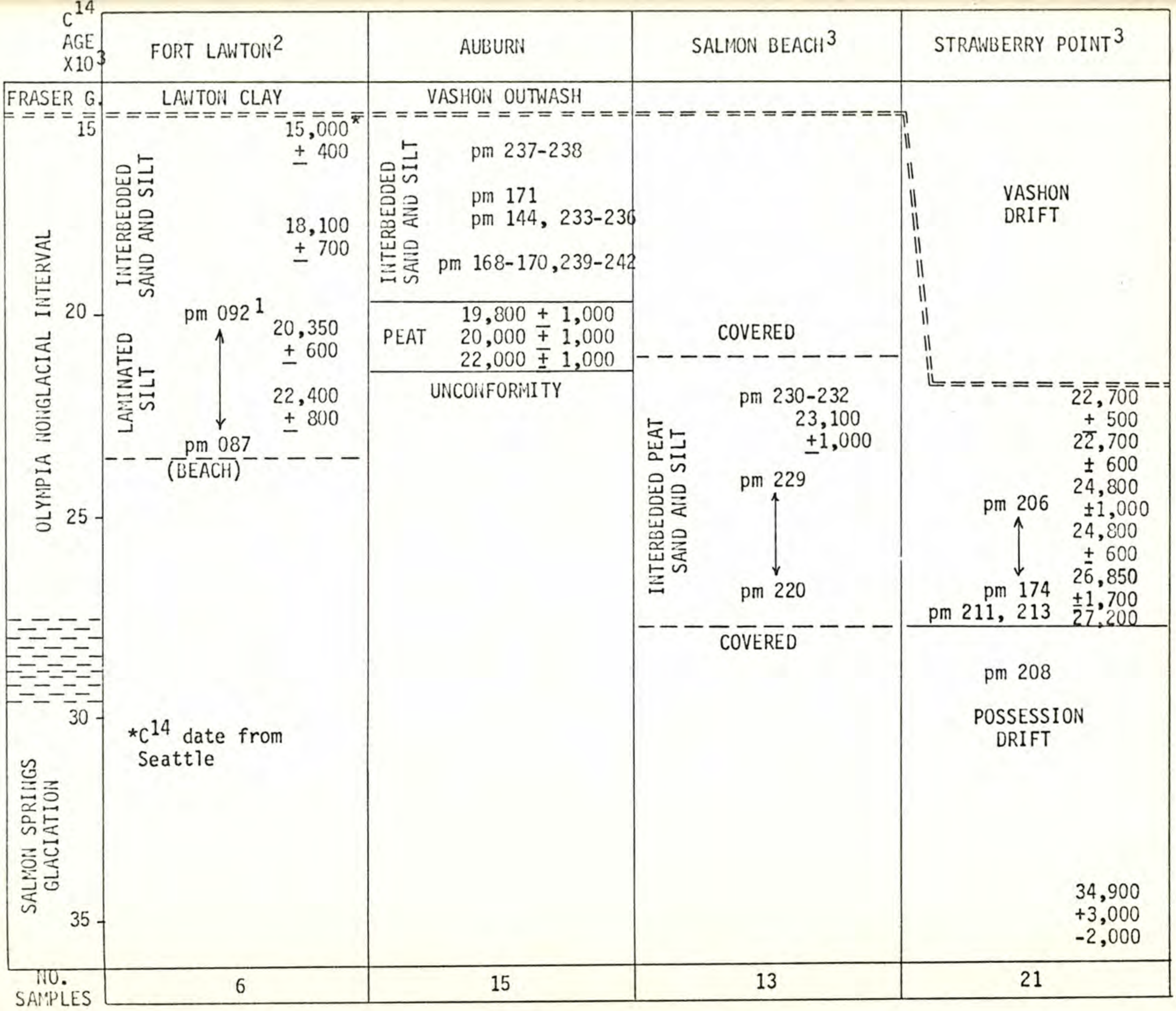
The time interval 15,000 years to approximately 30,000 years BP is represented by sediment of the Olympia nonglacial interval (Armstrong and others, 1965; Mullineaux and others, 1965; Easterbrook, 1969; Easterbrook and Smith, 1973). Four sites were selected: (1) the Fort Lawton sea cliff, (2) a borrow pit at Auburn, (3) a sea cliff at Salmon Beach, and (4) a sea cliff at Strawberry Point (Fig. 11).

Fort Lawton nonglacial sediments range in radiocarbon age from approximately 15,000 years BP to approximately 22,400 years BP (Mullineaux and others, 1965). However, sand phases, unsuitable for paleomagnetic measurement, prevented sampling the section representing about 20,000 years to 15,000 years BP. Paleomagnetic results from the Fort Lawton nonglacial sediments are shown in Fig. 12.

Olympia nonglacial sediments at the Auburn borrow pit consist of a peat layer and five thin silt layers interbedded with a coarse, black sand (Fig. 13). Fifteen samples were taken from the silt layers. The paleomagnetic results are shown in Fig. 14.

Figure 11. Sample localities, stratigraphy and radiocarbon dates for Olympia sediments.

[1]Paleomagnetic sample identification number.
[2]Mullineaux and others (1965)
[3]Easterbrook (1969; and unpublished data)


C14 AGE X10^3
FORT LAWTON2
AUBURN
SALMON BEACH3
STRAWBERRY POINT3
FRASER G.
OLYMPIA NONGLACIAL INTERVAL
SALMON SPRINGS GLACIATION
15
20
25
30
35
LAWTON CLAY
INTERBEDDED SAND AND SILT
LAMINATED SILT
15,000*
± 400
18,100
± 700
pm 092 1
20,350
± 600
22,400
± 800
pm 087
(BEACH)
*C14 date from Seattle
VASHON OUTWASH
INTERBEDDED SAND AND SILT
pm 237-238
pm 171
pm 144, 233-236
pm 168-170,239-242
PEAT
19,800 ± 1,000
20,000 ± 1,000
22,000 ± 1,000
UNCONFORMITY
COVERED
INTERBEDDED PEAT SAND AND SILT
pm 230-232
23,100
±1,000
pm 229
pm 220
COVERED
VASHON DRIFT
22,700
± 500
22,700
± 600
24,800
±1,000
24,800
± 600
26,850
±1,700
27,200
pm 206
pm 174
pm 211, 213
pm 208
POSSESSION DRIFT
34,900
+3,000
-2,000
NO. SAMPLES
6
15
13
21

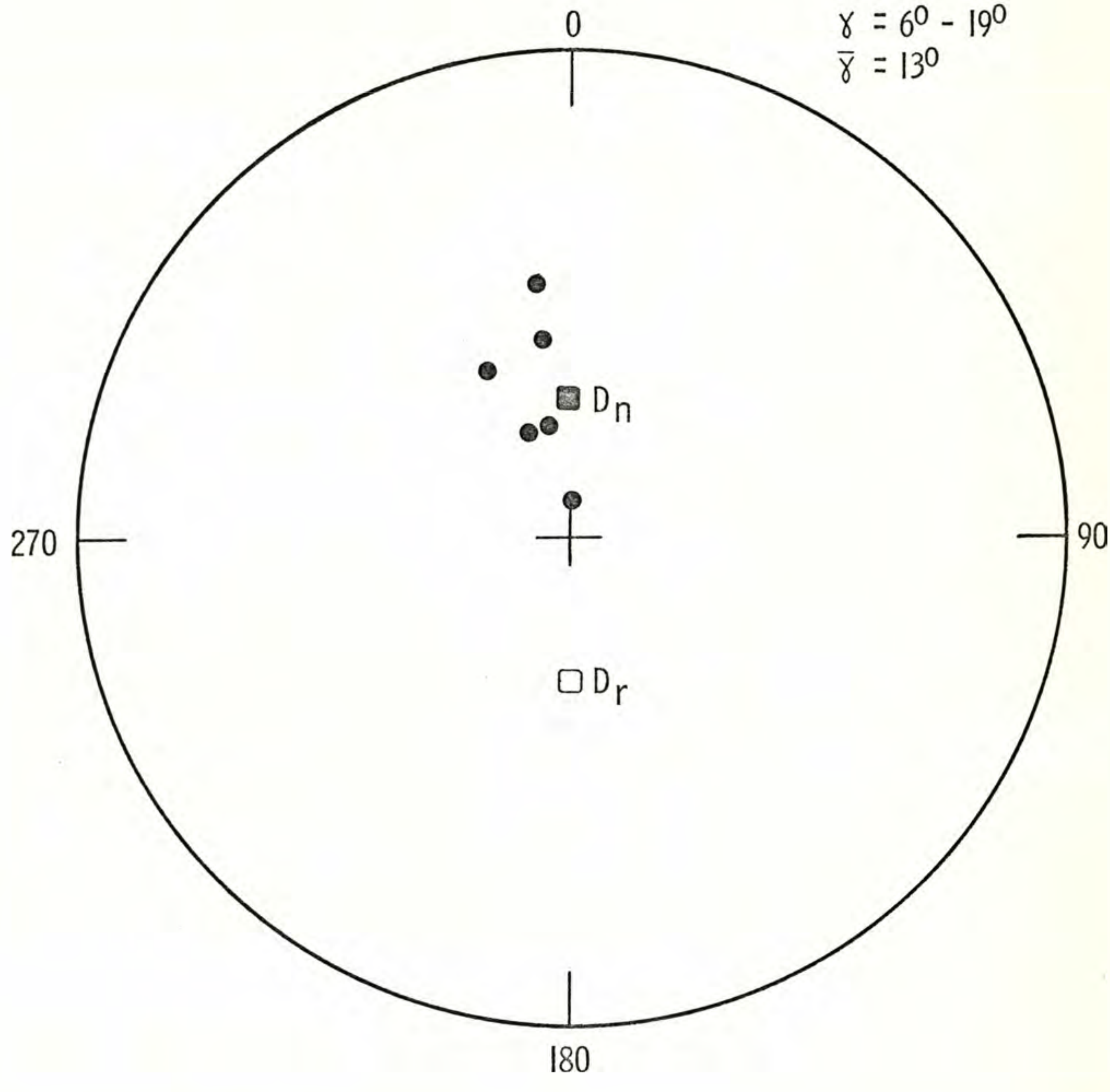


Figure 12. Paleomagnetic directions in Fort Lawton sediments of Olympia age (≅ 20,000 - 22,400 years BP). Explanation of symbols as in Figure 9.

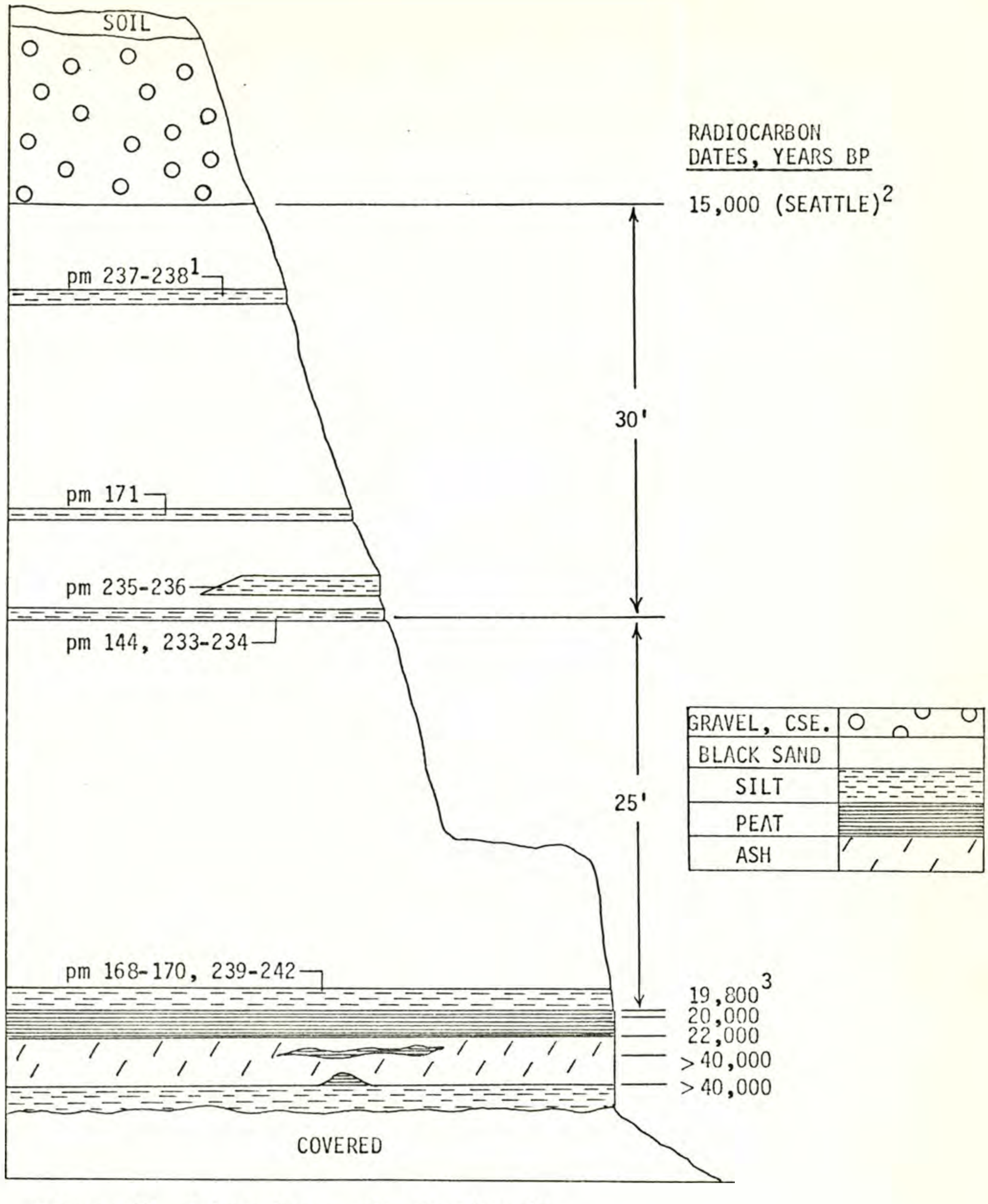


Figure 13. Auburn borrow pit stratigraphy.

[1]Paleomagnetic sample identification number.
[2]Mullineaux and others (1965).
[3]Easterbrook, unpublished dates.

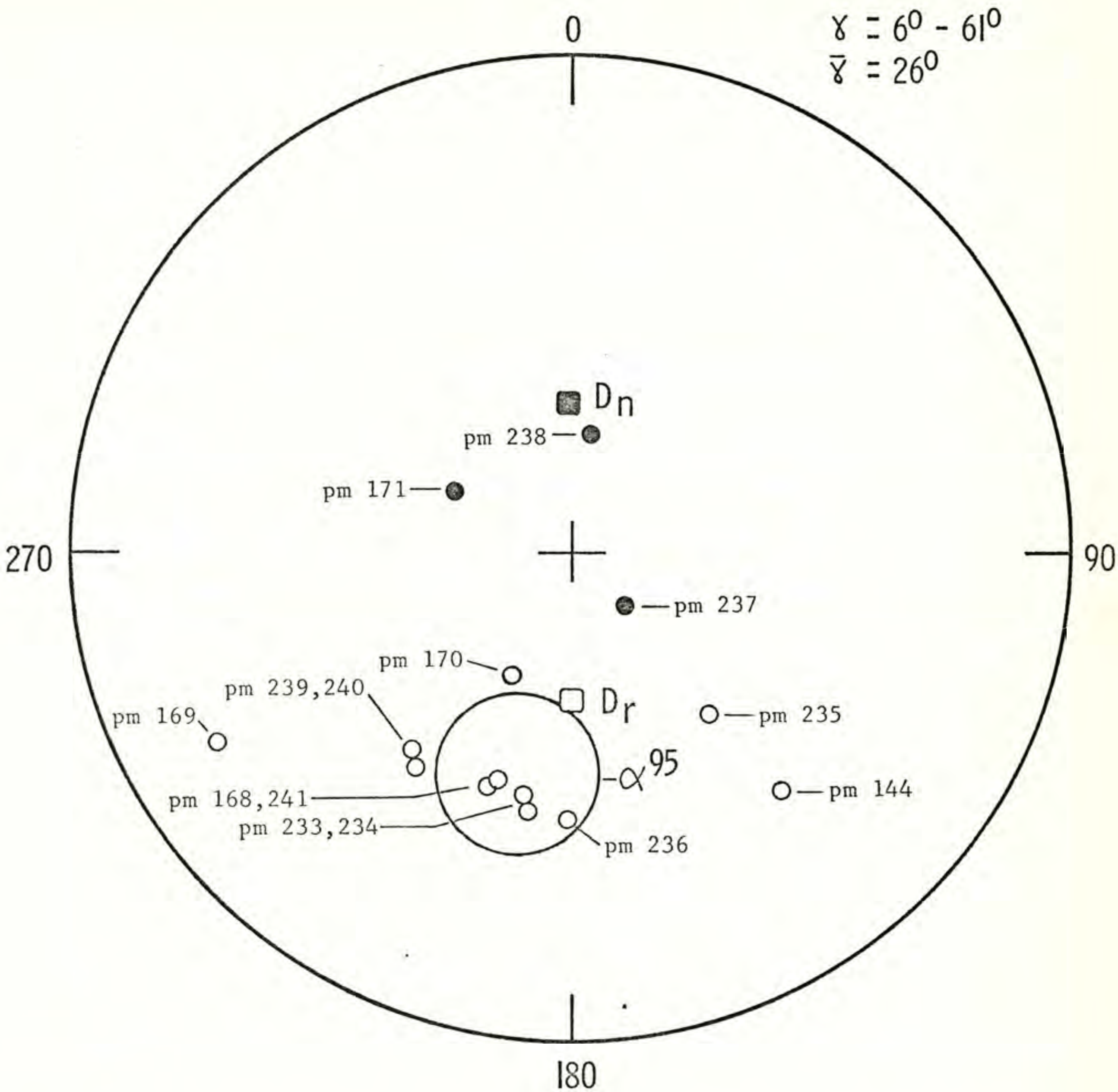


Figure 14. Paleomagnetic directions in Auburn sediments of Olympia age (≅ 15,000 - 20,000 years BP). Explanation of symbols as in Figure 9. Sample numbers are paired with directional symbols. The circle of confidence (α^{95}) for the 10 reversed directions is shown, and the statistics are as follows:

Circle of confidence (α^{95}) = 14.3^o
Angular standard deviation = 23.3^o
Precision parameter = 11.2

The Auburn borrow pit site yielded radiocarbon dates ranging from 19,800 years BP to >40,000 years BP. Vashon outwash gravels that top the section provide an upper age limit of 15,000 years BP for the underlying nonglacial sediments. The geologic map that includes the Auburn site (Waldron, 1961) indicates that all material underlying Vashon Drift is Salmon Springs age or older (>40,000 years). However, two lines of evidence utilized in this investigation indicate the presence of Olympia age sediments at the Auburn site: (1) three radiocarbon dates, all within limits of error of each other, place the age of the peat overlying the ash at 22,000 - 19,800 years BP, and (2) analysis of the mafic heavy minerals from the black sands suggests that these sediments are post Salmon Springs in age. Mullineaux (1970) demonstrates that pyroxene-amphibole ratios of volcanic sands of Mt. Rainier provenance vary with age (Table 1). Pyroxene was found to dominate over amphibole in Vashon outwash and modern alluvium, but pyroxene and amphibole were found to comprise approximately equal amounts in Salmon Springs and older drift. Black sands overlying the 19,800 years old peat at the Auburn site have a ratio of 2.7 pyroxene to 1 amphibole. Therefore, the black sands and the intercalated silt layers are probably younger than Salmon Springs age, corroborating the 19,800, 20,000 and 22,000 years BP radiocarbon dates.

The Salmon Beach sea cliff yielded one radiocarbon date of 23,100 years BP from one of four peat layers intercalated with silt and sand. The whole section may represent an interval of at least 4,000 years based on a comparison of the thickness of this section with similar Olympia nonglacial sections. More radiocarbon dates are needed, however, to establish the time span covered. Paleomagnetic results from Salmon Beach are shown in Fig. 15.

	Ratio of pyroxene to amphibole	
Modern alluvium	5.3/1	(21 determinations)[1]
Vashon drift	approx. 2.5/1	(16 determinations)[1]
Salmon springs and older drift	approx. 1/1	(17 determinations)[1]
Black sands from the Auburn site	2.7/1	(433 grains counted)

[1]Each determination based on 100 grain counts of heavy mineral suites (Mullineaux, 1970).

Table 1. Pyroxene and amphibole ratios for Pleistocene and Holocene sediments in the Auburn area.

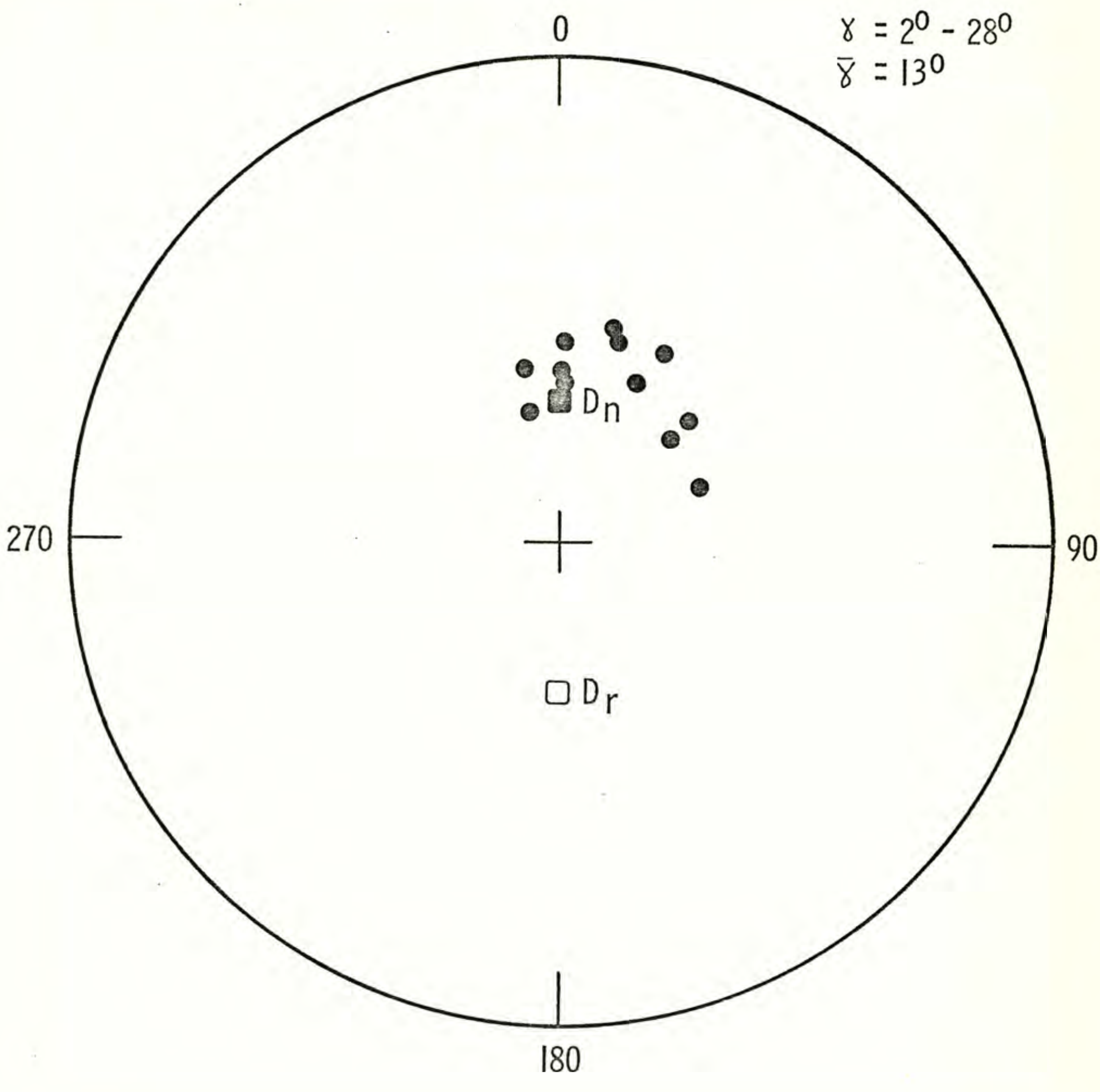


Figure 15. Paleomagnetic directions in Salmon Beach sediments of Olympia age (≅ 23,000 - 27,000 years BP). Explanation of symbols as in Figure 9.

The Strawberry Point sea cliff on Whidbey Island exposes sediments of the Olympia nonglacial interval consisting of silt and peat layers. Radiocarbon dates from the peats range from 22,700 years to 27,600 years BP. Suitable samples were not available from the uppermost part of the section. The section sampled probably covers a time range of approximately 24,000 years to 28,000 years BP. One sample was obtained from the underlying Possession Drift dated between 28,000 years and 35,000 years BP. Paleomagnetic results from Strawberry Point are shown in Fig. 16.

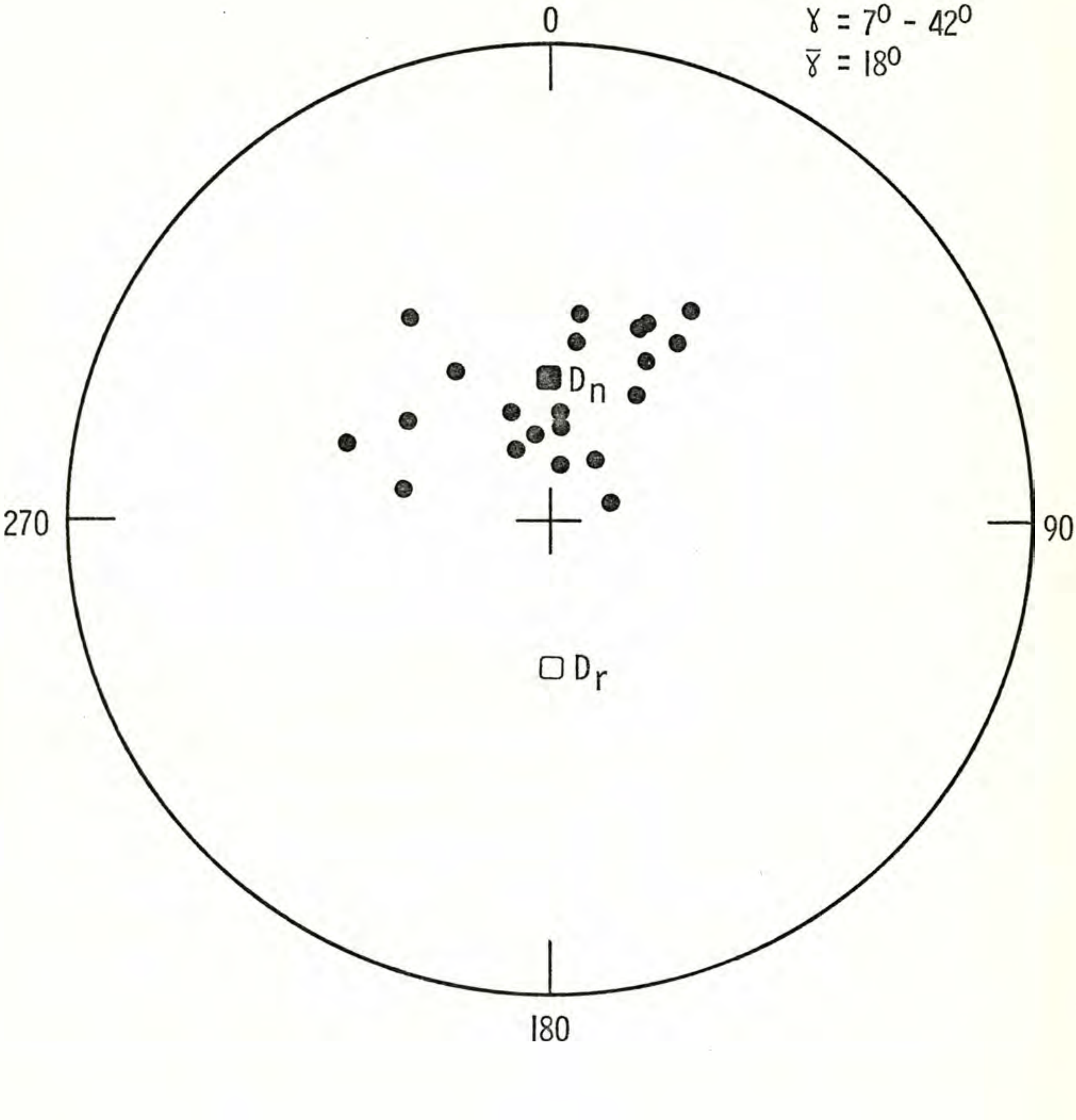


Figure 16. Paleomagnetic directions in Strawberry Point sediments of Olympia age (≅ 24,000 - 30,000 years BP). Explanation of symbols as in Figure 9.

DISCUSSION OF RESULTS

Geomagnetic Polarity

Paleomagnetic directions from 118 samples representing a time range of approximately 11,000 years to 30,000 years BP are plotted on sterographic projections in Figures 9, 10, 12, 14, 15 and 16. It may be observed that polarities are normal, with the exception of those from the Auburn site. All of the samples from the lower two silt layers at Auburn provided reversed directions. More complete vertical sampling may have provided transitional directions, but thicker beds were not encountered. The three younger samples from the upper silt layers have downward pointing inclinations (normal). One of these, pm 237, has a southeasterly declination, but may have been rotated about the vertical axis during collection. It was collected from the same slab as pm 238 and some difficulties were encountered obtaining an accurate orientation scribe.

The interpretation of Puget Lowland geomagnetic polarity for the time interval studied is shown in Figure 17. The paleomagnetic results indicate that the local geomagnetic field polarity was as follows: normal at about 30,000 years BP; normal from approximately 28,000 years to 20,000 years BP; reverse polarity, but returning to normal (no transition recorded), during the interval 20,000 years to 15,000 years BP; normal from 15,000 years to 13,500 years BP; and normal from 13,000 years to 11,000 years BP. No data was obtained for the short interval between 13,500 years to 13,000 years BP.

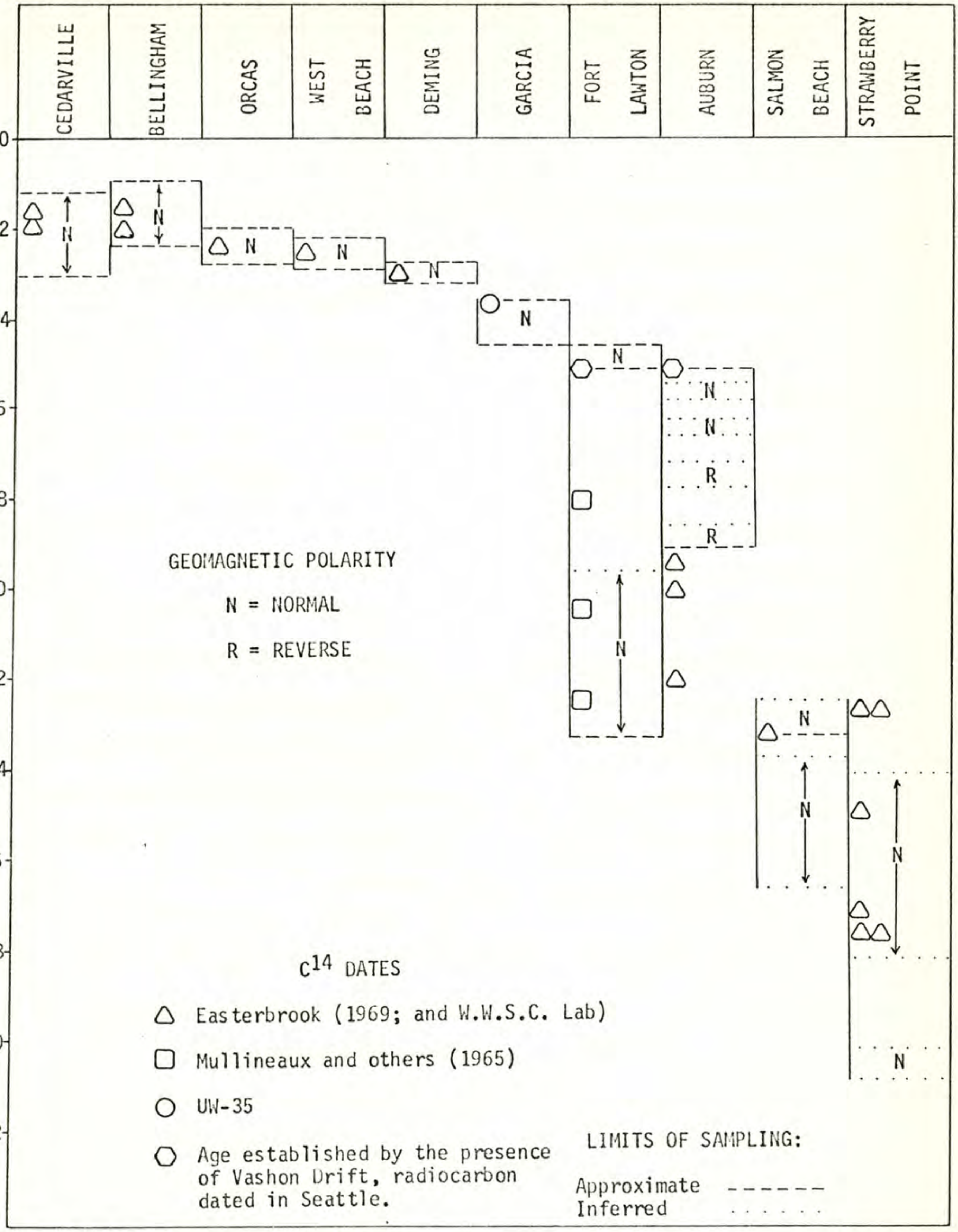


Figure 17. Geomagnetic polarity interpretation for the Puget Lowland.

Geomagnetic Field Variation

The variations between paleomagnetic directions determined in this investigation probably result from three very different causes: 1) sampling and laboratory errors, 2) systematic errors causing the detrital remanence to diverge from the ambient field direction, and 3) variations in the local geomagnetic field directions as a function of time. The degree to which the paleomagnetic directions are an accurate measure of geomagnetic field variation depends on the two sources of error, 1) and 2), above.

Samples which were deposited at the same time, along a single horizon, should have recorded the same ambient field direction. Therefore, examination of the maximum directional divergence of samples collected for reliability studies (Figs. 5 and 6), provides an estimate of the possible maximum variation due to errors. The divergences from the mean direction (γ') for silt and glaciomarine drift are 1° - 26° (average 7°) and 3° - 11° (average 6°), respectively.[1] Therefore, one may expect deviations between samples of differing age to be as great as 26° and averaging approximately 10° even if no geomagnetic field variations occur. However, these values should probably not be considered upper limits for errors, but order-of-magnitude approximations of the range of errors possible.

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[1] Adapting techniques utilized by Lawley (1970), γ' is defined as the divergence of the direction of magnetization of one sample from the mean direction of all samples of the same age. γ is defined as the divergence of the direction of magnetization of one sample from the normal or reverse axial centered dipole direction (0° or 180°, $+66^{\circ}$ or -66° for Lat. 48° N.) for any one age group (Figs. 9, 10, 12, 14, 15 and 16). $\bar{\gamma}$ is the mean divergence, for any one group, from the normal or reverse axial centered dipole direction.

Observing the values of γ and $\bar{\gamma}$ (Figs. 9, 10, 12, 14, 15 and 16), in general the variations in directions within each group may be considered to be partly due to errors. However, divergences may be due to the secular variation of the geomagnetic field direction, which could result in values of γ ranging from 0^{o} - 20^{o} (archaeomagnetic data for the last 2,000 years, Nagata, 1961), and much greater if the geomagnetic field intensity drops (Lawley, 1970). Secular variation of the geomagnetic field direction is comprised of slowly migrating features. The most prominent of these features, the non-dipole field, is characterized today by westward drifting magnetic anomalies of approximately continental dimension with intensities varying from 0% to approximately 30% of the present dipole field intensity (Fig. 18). The effects of secular variation of the geomagnetic field direction are believed to average to an axial dipole field over periods of 10,000 years, and perhaps as short as 2,000 years (Stacey, 1969; Tarling, 1971; Strangway, 1970). The character of the secular variation is thought to have been the same throughout the Pleistocene (Kono, 1972). Secular variation results in deviations in paleomagnetic directions between samples which are separated by short time intervals. If the time interval is less than 2,000 years, samples separated by large geographic distances may give divergent paleomagnetic directions.

Another secular variation in the geomagnetic field is that of the intensity, which has been shown to vary by a factor of 2 between 2,000 years BP and 5,000 years BP (Bucha and others, 1970). Although this investigation does not include paleointensity data, some indication of past reduced field intensity may have resulted in directional variations in the paleomagnetic data. Two types of directional variation can be expected to

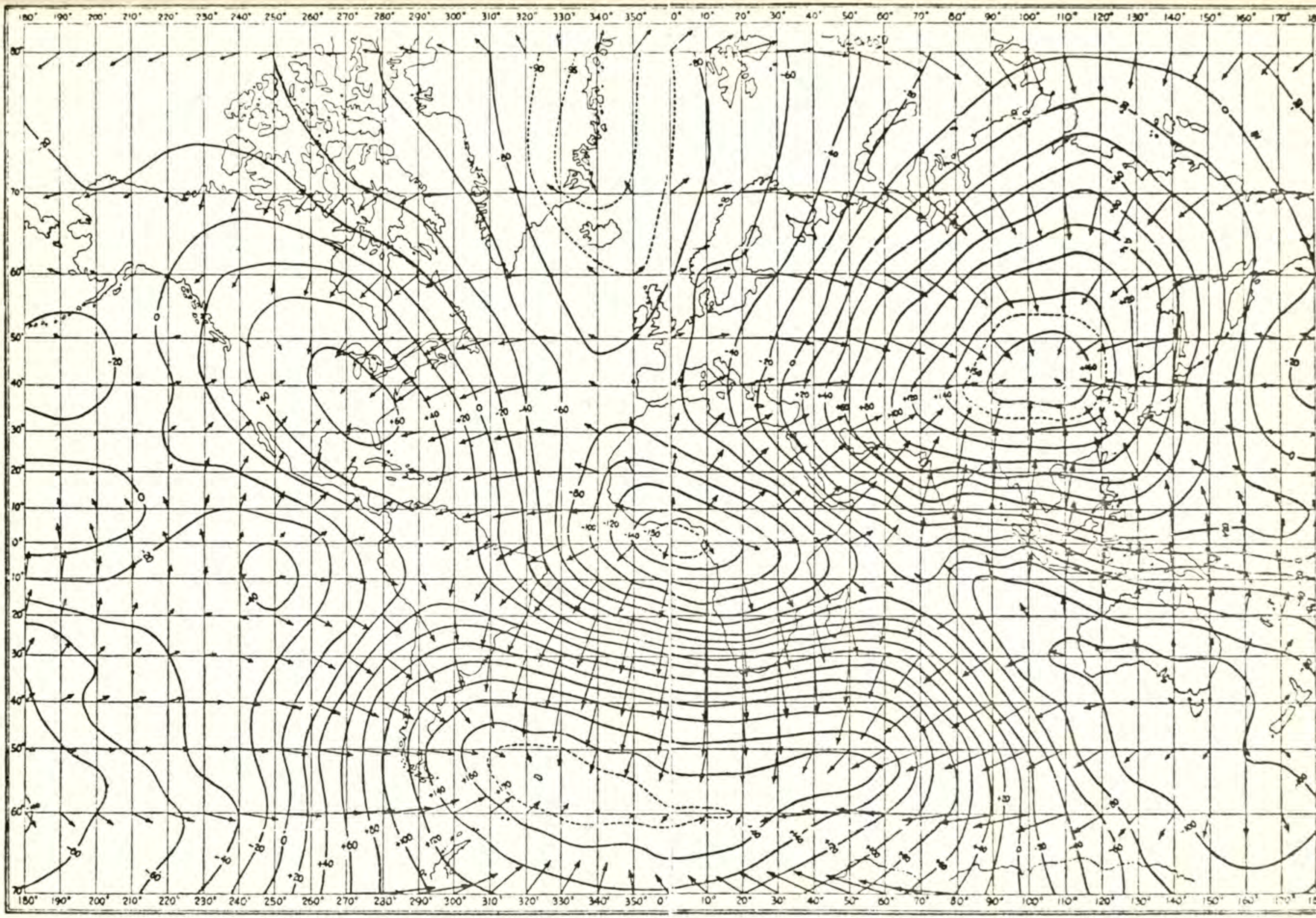

Figure 18. Non-dipole field for 1945. Vertical field intensity with contour intervals at 0.02 oersted. Arrows give horizontal component (from Bullard and others, 1950).

occur as a result of reduced geomagnetic field intensity. Firstly, large values of divergence (γ) would occur because the effect of secular variation increases. The paleomagnetic directions would, therefore, deviate greatly from the axial centered dipole direction (Lawley, 1970; Watkins, 1969). Secondly, the net geomagnetic field polarity may change such that the field polarity is reversed after the period of reduced intensity. Cox (1968, 1969) estimates that if a reduced dipole intensity has occurred every 10^4 years, there is a probability of 0.05% that a reversed polarity will occur during any one 10^4 year period (based on occurrence of geomagnetic polarity events during the late Tertiary and Quaternary Periods.

The first type of variation, large divergences from an axial centered dipole, probably occurs every time total geomagnetic field intensity reduces to low values ($\cong$16% of present value), and will result in divergences of 40° - 90° (Lawley, 1970). None of the paleomagnetic directions from this investigation show divergences that great, although both reversed and normal directions were recorded.

The question arises, if a geomagnetic dipole reversal is the cause of the reversed directions at Auburn, why are no divergent, transitional direction recorded? Two possible explanations exist. Firstly, if sedimentation rates were slow, each sample would have recorded a direction in which secular variation was averaged, and no divergent directions occur. For example, ocean sediments have sedimentation rates generally much less than 1 cm/1000 years. This can be discounted for the Puget Lowland sediments studied. Estimates of sedimentation rates range from greater than 1 cm/year for pro-glacial lake varves and glaciomarine drift, to no less than 1 cm/50 years for nonglacial silts (1 cm/50 years is an underestimate because often the measured sections contain peat beds which have been compressed vertically). Therefore, the samples used in this investigation should have recorded any divergent directions due to secular variation.

The second possibility is that the polarity transition occurred very rapidly, on the order of 1,000 years or less, such that sand phases at the Fort Lawton site, about 20,000 - 15,000 years BP, prevented the sampling of silt which would have recorded divergent, transitional directions. Similarly at the Auburn site, where sands occupy much of the section, silt layers comprised only about 1 meter out of 15 meters between the peat layer and Vashon gravels. Rapid transition of the local geomagnetic field polarity is considered to be the best explanation for the lack of observed transitional directions in this investigation.

Rapid transition of the local geomagnetic field polarity may be due to either drifting non-dipole anomalies, or short term, flip-flop behavior of the dipole field. If the total geomagnetic intensity is so low, such that the non-dipole generates all or most of the net geomagnetic field, paleomagnetic directions recorded in less than 100 years duration (lavas, sediments with high rates of deposition) would vary greatly on the surface of the earth (Fig. 18). Fully normal, transitional, and fully reverse directions could be recorded depending on the geographic location. Extrapolating from observed geomagnetic intensity variations in the last 9,000 years (Bucha and others, 1970), it is estimated that the net geomagnetic intensity may remain low enough for the above phenomenon to occur during intervals of about 2,000 years, which is well within the interval required by the data in this investigation.

Alternatively, the dipole field may change polarity during an interval in which geomagnetic field intensity varies from high to low, then low to high (Cox, 1968).[1] Two such intervals (one cycle) would have to have

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[1] In Cox's model, a polarity change occurs whenever the quantity ($M_A + M_{A'}$) changes sign (M_A is the axial moment of the dipole field and $M_{A'}$ is a measure of the non-dipole field).

occurred within 5,000 years (15,000 - 20,000) if the Puget Lowland data are correct, and Cox's model is correct. Extrapolation of archeomagnetically observed paleointensity fluctuations (Bucha and others, 1970) indicates periods of 8,000 years just to complete one cycle which would be required to explain the observed double transition. Therefore, if the reversed paleomagnetic directions are the result of changes in dipole polarity, either Cox's model is incorrect, or the geomagnetic field intensity may have varied at greater rates 15,000 - 20,000 years BP than observed for the last 9,000 years. However, Cox's model is probably correct as data from Lawley (1970) support the hypothesis of low field intensity-dependent polarity changes. Unfortunately, extrapolations of geomagnetic field intensity fluctuations beyond 9,000 years BP can only be considered tenuous because there is insufficient paleointensity data.

Based on the above examination, the observed polarity change between 15,000 years and 20,000 years BP in the Puget Lowland could be the result of drifting non-dipole anomalies rather than dipole polarity reversals. Two important tests of this conclusion will be 1) paleointensity data, and 2) a comparison of paleomagnetic results from other, distant localities covering the same time range. Paleointensity measurements using sediments is presently unfeasible, but a world-wide paleomagnetic comparison may show whether or not a dipole polarity change occurred during the period 20,000 - 15,000 years BP.

Comparison with other Regions

Some paleomagnetic data for the last 50,000 years is available from Europe, North America, the three main oceans, the Gulf of Mexico, the Mediterranean Sea and the Bering Sea. The results include several records of reversed polarities at a variety of locations. The most notable of these is the Laschamp reversed event (see discussion below of the use of the term "event").

Bonhommet and Babkine (1967) and Bonhommet (1970) reported reversely magnetized lavas in France whose age was estimated to be less than 50,000 years BP. Potassium-argon and radiocarbon dates by Bonhommet and Zarhinger (1969) indicate that the lavas may be as young as 20,000 years, but no younger than approximately 8,000 years. Mean declinations and inclinations for the Laschamp and Olby flows were 231^o, -65^o and 234^o, -71.5^o, respectively. The divergence (γ) from a reversed, axial centered dipole direction at 45^o N. is 22^o.

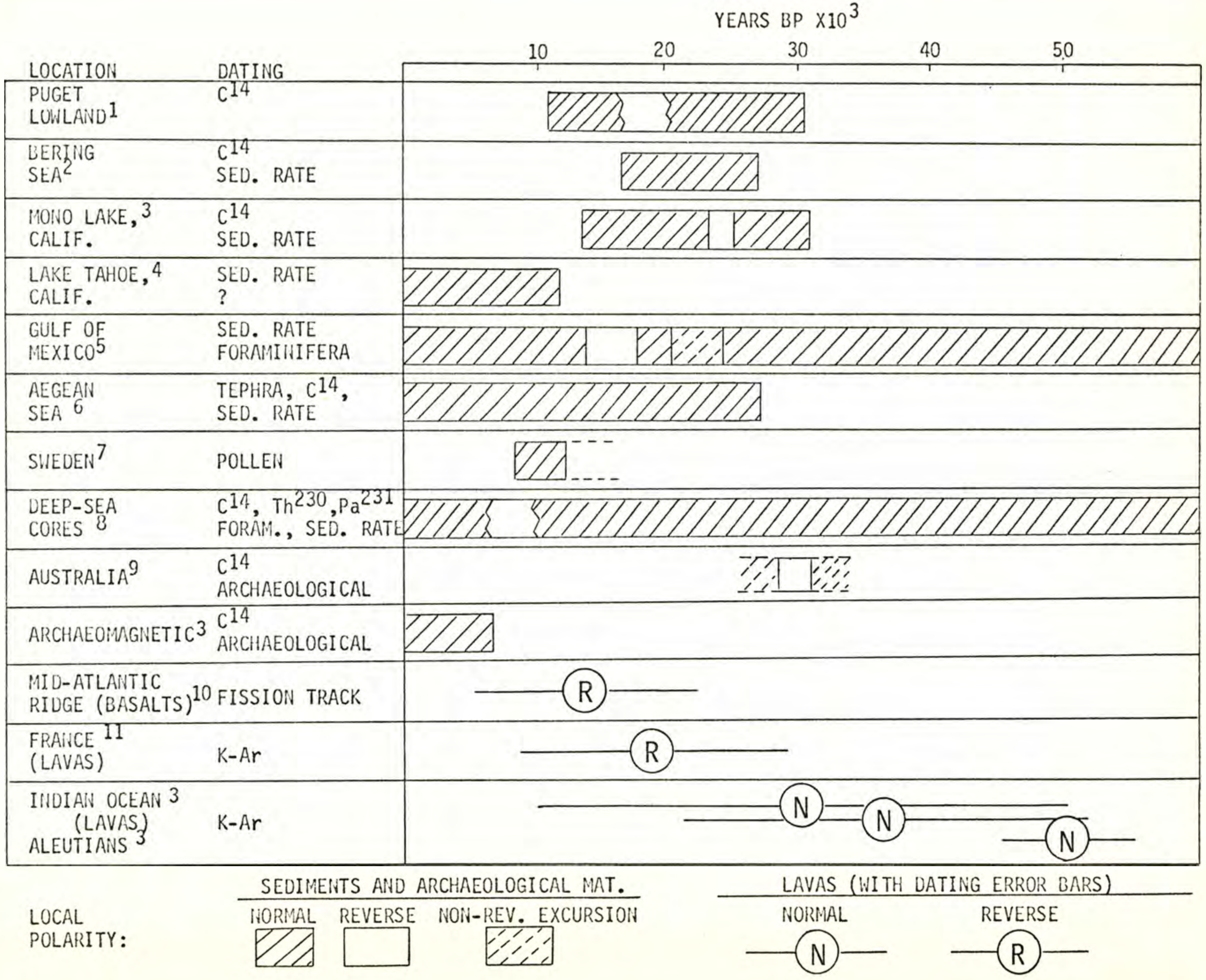
Whitney and others (1971) concluded that the magnetization of the Laschamp and Olby flows had not self reversed, but recorded valid paleomagnetic directions. Whether the reverse polarity was due to a dipole reversal or to a localized field reversal has not been confirmed.

Unfortunately, the polarity change recorded in the Laschamp and Olby flows was labeled the Laschamp "event" (Bonhommet, 1970; Bonhommet and Zarhinger, 1969), implying status equal to events within the geomagnetic polarity time scale; that is, world-wide confirmation. Cox (1969) included the Laschamp event in his summary paper on geomagnetic reversals, but Denham and Cox (1971) threw considerable doubt on its world-wide existence (Fig. 19). Nonetheless, Laschamp "event" is widely used, inferring that it is an established polarity event, similar to those within Matayuma and older polarity epochs.

Evidence from other parts of the world does not, at this time, prove or disprove the existence of the Laschamp as an event, including this investigation. Time correlations between reported paleomagnetic results from various localities are probably tenuous due to the varying

Figure 19. World-wide paleomagnetic results, 0-50,000 years BP.

[1]This investigation
[2]Whitney and others (1971)
[3]Denham and Cox (1971)
[4]Goldstein and Henyey (1970)
[5]Clark and Kennett (1972)
[6]Opdyke and others (1972)
[7]Morner and others (1971)
[8]Wollin and others (1971)
[9]Barbetti and Mc Elhinny (1972)
[10]Aumento and others (1973)
[11]Bonhommet and Babkine (1967); Bonhommet and Zarhinger (1969)


YEARS BP X10^3
10
20
30
40
50
LOCATION
DATING
PUGET LOWLAND1
C14
BERING SEA2
C14 SED. RATE
MONO LAKE,3 CALIF.
C14 SED. RATE
LAKE TAHOE,4 CALIF.
SED. RATE ?
GULF OF MEXICO5
SED. RATE FORAMINIFERA
AEGEAN SEA 6
TEPHRA, C14, SED. RATE
SWEDEN7
POLLEN
DEEP-SEA CORES 8
C14, Th230, Pa231 FORAM., SED. RATE
AUSTRALIA9
C14 ARCHAEOLOGICAL
ARCHAEOMAGNETIC3
C14 ARCHAEOLOGICAL
MID-ATLANTIC RIDGE (BASALTS)10
FISSION TRACK
FRANCE 11 (LAVAS)
K-Ar
INDIAN OCEAN 3 (LAVAS) ALEUTIANS 3
K-Ar
R
R
N
N
N
LOCAL POLARITY:
SEDIMENTS AND ARCHAEOLOGICAL MAT.
NORMAL
REVERSE
NON-REV. EXCURSION
LAVAS (WITH DATING ERROR BARS)
NORMAL
REVERSE
N
R

dating techniques utilized and their inherent inaccuracies (Fig. 19). In addition to potassium-argon, dating techniques used include radiocarbon dating where possible, fission track dating, sedimentation rates, and chronology based on foraminifera, tephra (ash), and pollen.

Some localities provide paleomagnetic evidence of continuously normal polarities for considerable periods of time, overlapping periods of reversed polarity recorded at other localities. This should not be the case if the reversed magnetizations were due to dipole polarity changes. However, such inconsistent evidence could result from a situation in which the Laschamp was a localized geomagnetic excursion, due to the non-dipole field, during a time of low total geomagnetic field intensity, as discussed earlier. In this situation attempts to confirm the world-wide occurrence of the Laschamp would be in part positive, and in part negative. Localized excursions of the field could have been recorded at various localities, with differing degrees of divergence, none of which would be exactly time equivalent. A similar hypothesis was expressed by Watkins (1972).

Measurements of the ancient geomagnetic field intensity would provide evidence as to the validity of this hypothesis. If the Laschamp and Olby lavas, indeed, cooled at a time of low geomagnetic field intensity, it is possible that the non-dipole was responsible for the Laschamp polarity excursion. Bonhommet and Zarhinger (1969) indicated that paleointensity measurements from the Laschamp and Olby flows would be forthcoming. At present, paleointensity measurements are not possible using sediments, but in some reported instances the NRM intensities dropped to low values when large directional divergences occurred (Denham and Cox, 1971; Wollin and others, 1971). Further experimental studies may show whether or not dipole polarity changes were responsible for any of the reversed polarities reported for the last 50,000 years.

An interregional paleomagnetic comparison does not, at this time, establish a geomagnetic polarity event during the last 50,000 years. However, because radiocarbon dating can provide very good resolution of the short period geomagnetic fluctuations, highly useful geomagnetic time scales for the last 50,000 years may possibly be developed which could provide an additional correlation tool for glacial geologists, archeologists, and other late Quaternary investigators. Time scales need to be developed for particular regions, comprised of polarity changes, secular variation curves and intensity variations.

SUMMARY

The purpose of this investigation was fourfold. The first purpose was to demonstrate whether or not fine-grained sediments exposed in the Puget Lowland have recorded valid paleomagnetic directions. Stability tests demonstrate that the detrital remanent magnetizations are strong and stable. Reliability tests show that any one paleomagnetic direction has a high probability of being the true direction, and that glaciomarine drift records reliable paleomagnetic directions.

The second purpose of this research was to begin development of a polarity time scale for the Puget Lowland. Figure 17 shows the results, and demonstrates that the local ambient field direction was predominantly normal from 30,000 years to approximately 11,000 years BP, with a short reverse polarity occurring between 20,000 years and 15,000 years BP.

The third purpose was to provide information about short term variations in the geomagnetic field. Divergence of paleomagnetic directions from those of an axial centered dipole are considered to be due to sampling errors, detrital remanence acquisition errors, and geomagnetic secular variation. The local polarity reversal between 20,000 years and 15,000 years BP could have been caused by the geomagnetic non-dipole field rather than the dipole field.

The fourth purpose was to make comparisons with polarity results from other regions for possible elucidation of the existence of polarity events during the last 50,000 years. At this time, it is not possible to establish an interregional geomagnetic polarity event for this period. Recorded reversed paleomagnetic directions for the last 50,000 years may represent localized geomagnetic excursions. Although the dominant polarity for the last 50,000 years was normal, resolution of the details of geomagnetic

behavior could result in the development of regional geomagnetic time scales for the late Quaternary.

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APPENDIX 1. Paleomagnetic Data

	NRM			"CLEANED"		
Sample	Dec.[1]	Inc.[2]	M[3]	Dec.	Inc.	M
CEDARVILLE:						
pm 250	315.5	+64.5	1.17	----	----	----
251	322.5	+29	1.19	318.5	+10	0.69
252	322.5	+69.5	1.12	347.5	+76	0.19
253	299.5	+36	1.56	299	+31	1.12
254	52	+64.5	2.00	346	+65	0.52
256	7	+71.5	2.30	345.5	+75	0.60
257	353	+68	1.91	335	+71	0.45
258	25.5	+76	2.08	343.5	+75	0.46
259	296	+78.5	1.41	352	+62	0.33
260	43	+77	1.64	6	+82.5	0.40
261	12	+60.5	1.76	338.5	+77.5	0.39
262	9	+69	1.89	342.5	+75.5	0.46
263	351.5	+67.5	2.02	341.5	+72.5	0.51
264	346	+69	1.98	309	+73	0.54
resultant						
pm 254-264	----	----	----	341.5	+73.5	----
BELLINGHAM:						
pm 018	33.5	+43.5	0.64	7	+54	0.41
019	340.5	+27.5	2.15	328	+31.5	0.86
020	11.4	+60.5	0.86	19.5	+54.5	0.39
ORCAS:						
pm 265	30.5	+46	1.67	29.5	+43.5	0.85
WEST BEACH:						
pm 266	8.5	+68.5	0.87	----	----	----
267	25	+63	0.89	76	+59	0.36
DEMING:						
pm 021	4.5	+66	0.51	18.5	+78.5	0.39
023	1	+53	0.21	342.5	+57	0.07
GARCIA:						
pm 043	0	+57.5	0.36	351	+42.5	0.14
044	358	+57	7.45	358	+56.5	3.83
045	351	+69.5	8.27	345	+68.5	4.63
046	358	+64	0.97	349	+59.5	0.38
047	354	+61	0.43	350.5	+60.5	2.25
048	359	+65	2.21	3	+64.5	1.30
049	4.5	+32.5	1.73	6	+40.5	0.97
050	347	+55.5	3.63	354	+57	2.17
051	4	+58	3.19	8	+60.5	1.61
052	7.5	+33.5	3.51	8.5	+46.5	2.17
053	5	+68	3.07	358.5	+70.5	1.83
054	350	+63	8.08	337.5	+63	6.55

[1]Dec. = declination; azimuth of the horizontal component of the remanent magnetization direction.
[2]Inc. = inclination; dip of the remanent magnetization direction.
[3]M = intensity of magnetization, emu/cm^3 X10^{-3}.

GARCIA, cont.						
pm 055	351	+53	0.19	----	----	----
056	0.5	+46	1.50	2	+50	1.01
057	341.5	+52.5	0.98	349	+48.5	0.73
066	356	+39.5	1.29	356.5	+52	0.86
067	6.5	+53	1.60	356.5	+54.5	0.70
070	2.5	+39	1.30	3	+35	0.70
071	355	+49	0.53	354.5	+51	0.26
074	357.5	+31	0.57	355.5	+37	0.31
076	9	+59.5	1.91	14.5	+59	1.02
FORT LAWTON:						
pm 087	354.5	+88	0.30	7.5	+84	0.08
088	343	+49.5	1.22	333.5	+59	0.40
089	351	+61	1.79	339	+70.5	0.52
090	359	+55.5	0.98	352.5	+56	0.27
091	4.5	+55.5	0.47	353	+46	0.16
092	0	+60.5	0.87	350	+60.5	0.18
105	344.5	+58	0.74	343	+48.5	0.32
107	344.5	+32.5	0.35	342.5	+4.5	0.17
108	319.5	+39	0.53	302.5	+7	0.23
109	4	+39	0.56	330	+16.5	0.26
110	357	+34	0.61	6.5	+17	0.19
111	355	+38.5	0.51	351	+26	0.25
112	5.5	+33	0.32	2	+3	0.32
113	6.5	+48.5	0.65	357.5	+34.5	0.25
114	2	+51.5	0.42	334.5	+30.5	0.17
116	0	+42	0.31	348	+34.5	0.38
117	9	+39.5	0.51	9.5	+25.5	0.34
118	15.5	+50	0.23	9.5	+19.5	0.23
119	12.5	+19	0.55	15	+17	3.24
120	351.5	+30	5.02	349.5	+28.5	2.95
121	343.5	+36.5	5.90	341.5	+34.5	3.34
122	354	+26.5	4.61	354.5	+33	1.96
123	4.5	+41	2.33	359	+38	1.19
124	347	+59.5	4.17	350.5	+58	2.28
125	353	+38	4.36	351.5	+37.5	2.39
126	353.5	+21	3.56	356	+19.5	2.02
127	359	+18.5	4.85	356.5	+16.5	2.77
AUBURN:						
pm 144	60	+55.5	0.41	140	-35	0.06
168	291.5	-7	0.45	240	-19	0.31
169	210.5	-52	0.61	197.5	-58.5	0.54
170	295	-58	0.73	204.5	-66.5	0.63
171	289	+67	0.72	297	+70	0.13
233	345	+75	0.45	189	-44	0.07
234	330	+68	0.28	190.5	-46.5	0.18
235	125	-19	0.40	140.5	-53.5	0.44
236	173	-34.5	0.82	181	-43.5	0.87
237	63.5	+81	0.09	135	+76.5	0.06
238	5	+66	0.18	10	+71	0.11
239	222	-37.5	0.03	216	-46.5	0.30
240	238	+29.5	0.07	198.5	-47	0.10
241	243	+15.5	0.20	214.5	-44	0.11
242	57.5	+62	0.34	----	----	----

SALMON BEACH:						
pm 220	331.5	+70.5	0.30	0.5	+64	0.09
221	350	+57.5	0.30	345.5	+60.5	0.11
222	28	+60.5	0.005	14	+50.5	0.002
224	34	+63.5	0.004	0.5	+62	0.002
225	27.5	+53.5	0.39	27.5	+52	0.20
226	8.5	+58.5	2.63	12.5	+53	1.06
227	357.5	+52.5	4.32	0.5	+56.5	2.06
228	45.5	+75	1.48	46.5	+65.5	0.53
229	77	+77.5	2.71	69	+65.5	1.11
230	350	+69	1.81	345.5	+67.5	0.86
231	60.5	+75	1.35	45	+61	0.67
232	20.5	+63.5	0.60	25	+61	0.28
STRAWBERRY POINT:						
pm 174	21	+46	0.02	26.5	+52.5	0.013
177	10	+43	0.04	36.5	+52.5	0.05
182	306.5	+59.5	0.09	289.5	+52.5	0.05
186	0	+57.5	0.05	34	+45.5	0.02
187	8	+54	0.24	8	+54	0.12
188	29.5	+57	0.33	25	+54	0.10
189	8.5	+57.5	0.30	9	+59	0.07
190	328	+58.5	2.87	327.5	+60.5	0.16
191	323.5	+48	1.83	326.5	+47.5	1.05
192	39	+63.5	4.29	34.5	+64.5	2.61
193	1	+74	0.97	8	+74.5	0.53
194	319	+72.5	0.09	303.5	+61	0.04
195	301.5	+74	0.09	279.5	+64.5	0.09
196	350	+72.5	0.12	339	+70.5	0.04
200	31.5	+74	0.28	38.5	+77.5	0.06
201	323.5	+79.5	0.12	7	+81	0.04
204	----	----	----	4	+72.5	0.25
206	6	+72.5	0.14	347.5	+76.5	0.18
208	44.5	+71.5	0.13	30.5	+56	0.05
211	359	+77.5	0.05	78.5	+79	0.03
213	324.5	+74	0.09	334	+76.5	0.07