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**COLUMBIA (PLEISTOCENE) SEDIMENTS
OF DELAWARE**



by
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ABSTRACT

The Columbia deposits of Delaware form a sheet of sand with a maximum thickness of approximately 150 feet which covers most of the Coastal Plain portion of the State. The dispersal pattern, deduced from foreset dip directions of cross-bedding, indicates that the sediment entered the study area from the northeast, i.e., from the direction of the valley of the Delaware River between Wilmington and Trenton, and spread south and southeast over Delaware.

The Columbia sediment is essentially medium sand but coarser admixtures are typical in the northern, and finer admixtures in the southern parts of the area. Median grain size, maximum particle size, and maximum grade size decrease, and sorting improves, in the down-paleocurrent direction. The sands are subarkosic; the coarser fraction consists mainly of vein quartz, sandstone and quartzite, and chert and the heavy mineral suite is dominated by zircon, epidote, amphibole, sillimanite, and staurolite. The sandstone pebbles and cobbles (some containing Paleozoic fossils) are evidence of the Younger Appalachian source of some of the Columbia sediment, and the heavy minerals of metamorphic origin are indicative of an additional source in the crystalline rocks of the Piedmont and Reading Prong.

The sands represent deposits of a major stream system; the distal portion of which has been reworked by a transgressing and regressing sea which at one time covered at least the southern half of Sussex County. The systematic variation of the properties studied suggests only a single cycle of deposition. It is postulated that the stream which deposited the Columbia sands derived its great volume of water (and sediment) in part from the meltwater of a continental glacier. Channel cutting and filling are attributed to the distributary portion of this stream system operating on the Coastal Plain during a time of glacial advance and lowered sea level. A later stand of the sea several tens of feet above the present level is required by the marine features of southern Delaware.

INTRODUCTION

Purpose and Scope

The subject of this investigation is the upper layer of sand which covers a large portion of the northern Atlantic Coastal Plain. Detailed study has been limited to the Coastal Plain of Delaware, and adjacent parts of Maryland, an area of some 2,000 square miles.

The coarse, unconsolidated sands are often referred to as the "Pleistocene sediments" or the "surficial sands and gravels." They rest unconformably upon the truncated edges of the older rocks of the Coastal Plain, and are lithologically distinct from these rocks so that they may be readily separated, except in the subsurface of the southernmost part of the study area. The Pleistocene age of the deposits is known for a few areas and horizons, and is probable for the rest, but fossils and radiometrically datable material, and therefore, definite proof of age, are lacking over most of the area.

Economic justification for the study of these sands in Delaware may be found in the fact that almost all of the ground water contained in deeper aquifers must first pass through the upper sands, and the sands themselves constitute potentially the most productive aquifer in the State. In addition, a vast majority of the commercial sand and gravel, as well as various grades of fill, come from these deposits.

To decipher the geologic conditions under which such vast amounts of coarse clastic matter can be derived, transported, and deposited is an interesting challenge which has stimulated the thought of many authors; however, all of the problems have not been resolved. That the sands are unconsolidated facilitates study of their texture and mineralogy and offers an opportunity for investigation of the areal variation of these properties within a dispersal system defined by the fairly abundant cross-bedding. The present study seeks to provide a more nearly adequate petrographic description of this mass of sediment and to consider its dispersion and environments of deposition as aspects of its genesis. Finally, some conclusions are drawn regarding the relationship of geologic events recorded in the Delaware Coastal Plain to Pleistocene history.

Acknowledgments

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STRATIGRAPHY AND GROSS LITHOLOGY

Many names have been used to designate the deposits which are the subject of this report. A review of the very early studies, none of which has had a lasting effect on the nomenclature, has been given by McGee (1888). Recently Jordan (1962) has reviewed later stratigraphic terminology.

The materials in question are far easier to distinguish in nature than they are in terms of stratigraphic nomenclature. They form a nearly continuous veneer over the Coastal Plain of Delaware. The lower contact is an angular unconformity on the beveled edges of the older Coastal Plain rocks, the youngest known of which are Miocene. The only superjacent beds are Recent alluvium. The possible stratigraphic interval for the age of the sediments studied is mid-Miocene to Recent. Pleistocene age may be demonstrated by fossils or by radiometric means at a few localities and the entire mass is generally conceded to be Pleistocene. Evidence of this is found in correlation of the rare datable horizons, postulated conditions of deposition, and state of preservation. Accurate stratigraphic delineation within the interval will require painstaking efforts of long duration for few "bench marks" are available and much indirect study must be applied.

Names which have been used for all or a part of the deposits in this area include Columbia (McGee, 1886, 1887, 1888), Sunderland, Wicomico, and Talbot (Shattuck, 1901, 1906), Parsonsburg, Pamlico, Walston, and Beaverdam (Rasmussen, et al., 1960), and Omar (Jordan, 1962). Also, although Delaware is north of their usual area of application, the names of the terraces of Cooke (1930, 1936, 1945) are sometimes used. Jordan (1962) suggested a return to McGee's use of Columbia as a possible clarification of the nomenclature. Columbia Formation would designate the undivided sequence in northern Delaware; this would be elevated to Columbia Group in the south where it includes at least the Beaverdam and Omar Formations.

The terms in most common usage seem to be "Pleistocene," "surficial," and "Columbia" used in conjunction with such words as "sediments" or "deposits," e.g., "Pleistocene sands and gravels." Objections may be raised to all of these terms. "Pleistocene . . ." is poor usage of stratigraphic nomenclature, and, more important, the age of the materials, although not seriously challenged, is often extremely difficult to *prove*. "Surficial . . .," besides being evasive, neglects the fact that older rocks may be exposed at the surface and that alluvium is often present; also, it hints at a two-dimensional aspect which is misleading. This study suggests that the material in Delaware may be divorced genetically, as well as geographically, from the type Columbia of Washington, D.C. and therein lies the writer's only objection to the term Columbia. In view of its introduction into Delaware by its author (McGee, 1887; 1888) it is retained here as representing the closest approximation to an accurate, convenient, inclusive, and legal term for the entire mass.

The surficial sands consist mostly of fine-, medium-, and coarse-grained quartz sand. Gravel beds, cobbles, and even boulders are conspicuous in northern Delaware and silt beds are found both north and south but are thicker and more common to the south. The volume of all of these materials is small in comparison to that of the sands. The deposits are essentially unconsolidated although locally there may be considerable differences in the degree of induration due to interstitial clay and/or iron oxides. Heavy bands of limonite-cemented conglomerate are common, especially toward the north. Colors range from white through yellow, tan, and brown to reddish-brown. Total thickness of the surficial sand is variable

in the north, ranging up to a maximum of about 100 feet in features interpreted as former fluvial channels. The thickness becomes more constant at an average of about 125 feet to the south in Sussex County. A great variety of grain size is usually present in a vertical section of any length; however, individual beds of sand are rather "clean." Beds cannot be traced from exposure to exposure in the north despite more extensive exposure there than to the south. Subsurface exploration in southeastern Sussex County (Jordan, 1962) showed a section consisting of a thick medium- to coarse-grained sand (Beaverdam Formation) overlain by alternating beds of silt and sand (Omar Formation) which could be traced for more than 13 miles. Subsurface control is not adequate to extend this section with certainty.

METHODS OF INVESTIGATION

General Statement

The basic approach of the present study has been to examine as many exposures of the Columbia deposits as possible and record at each locality the gross lithologic and structural features of the sediments, including, in particular, cross-bedding dip azimuths. Samples from each locality were collected for laboratory study of texture and mineralogy.

For the purpose of an areal investigation, the properties of a sediment may be classified as directional or scalar. The term vector may be substituted for directional if a magnitude is assigned or a unit magnitude acknowledged. The only directional feature occurring in sufficient quantity to be useful in this study is cross-bedding. Various scalar properties which have been measured include, for example, thickness of cross-beds, dip of foresets, maximum particle size, and mechanical and mineralogical composition as well as several parameters derivable from these measurements.

Field Procedures

The achievement of a reasonably equal geographic distribution of samples and the compilation of data are facilitated in an areal study by the use of a sampling grid. The grid utilized here is based on meridians and parallels spaced 5 minutes apart which yields individual quadrangles measuring about 4.5 miles by 6 miles. This is the same system used for the statewide well numbering system employed in Delaware and described by Marine and Rasmussen (1954, p. 18, plate 2). The distribution of the seventy-five 5-minute quadrangles systematically sampled in this study is shown in plate 1 as are geographic and political entities useful for location reference. The quadrangles are designated by an upper and a lower case letter. Two outcrops were studied within each quadrangle. This yields a sample density comparable to similar published investigations as indicated in the definitive work and summary of Potter and Pettijohn (1963). The localities are designated as I or II preceded by the quadrangle letters. Sample localities are also shown in plate 1 and are listed in Appendix I.

Selection of the two exposures to be studied within a quadrangle was based primarily upon the degree of development and exposure of cross-bedding, and secondarily upon the thickness of the section exposed. No locality was selected within a quarter mile radius of any other locality. Because of a general scarcity of exposures the necessity of choosing between localities was rarely encountered. The location and altitude of each locality was noted and a section measured and

recorded. In addition, note was made of all other exposures encountered in the search for adequate sample localities and of significant morphologic features.

Most of the exposures used were sand pits, active or abandoned. Other artificial exposures included railroad and road cuts, drainage ditches, foundation excavations, and the banks of the Chesapeake and Delaware Canal. Natural exposures displaying a satisfactory thickness of strata are rare but some, found along streams and bays, were useful.

Potter and Pettijohn (1963, p 69) define a cross-bed as a sedimentation unit containing internal bedding inclined to the principal surface of deposition. The inclined beds are called foreset beds.

Potter and Olson (1954) found that the variability in dip direction of the foreset dips is greater between cross-beds than within one cross-bed. This appears also to be true of the surficial sands of Delaware as indicated by preliminary study of some large exposures containing many cross-beds. As pointed out by Potter and Olson (1954) and Potter and Pettijohn (1963), a more representative sampling will be obtained in such cases by measuring the dip azimuths of many cross-beds than making more than one measurement within each cross-bed. In the present investigation one measurement was made in each cross-bed and up to five cross-beds were used at each exposure. In general not many more than five cross-beds are exposed at a given locality and in some cases a lesser number was used. At slightly less than half of the sample localities no measurable cross-bedding could be found.

Selection of the cross-beds to be measured was based mainly on their degree of development and accessibility. Wherever possible the measurements were taken in a vertical line and an effort was made to get vertical distribution. Measurements of dip azimuth and dip were taken by Brunton compass with the aid of a "dip direction indicator" patterned after the device described by Pryor (1958). Where possible the type of cross-bedding (tabular, trough), the thickness of the cross-bed, and the nature of the bottom contact of the foresets (tangential, non-tangential) were recorded.

Samples of the channel type were taken through the entire thickness exposed at each locality excepting only the soils. The actual intervals channeled were from 2.5 feet to 38 feet thick. After excavating and cleaning the exposure the channel was cut by shovel or pick keeping control of the width and depth of the channel so that equal representation might be given to each part of the section. Samples of about 0.5 to 25 kilograms were taken in this manner, the size being controlled by the height of the section and the texture of the material. Sample size was reduced in the field by means of a large sample splitter, heeding the suggestion of Wentworth (Krumbein and Pettijohn, 1938) that several particles of the largest grade present be retained. In this manner samples of from several hundred grams to almost 6 kilograms were selected for mechanical analysis. Particles larger than 64 mm in intermediate diameter were sized and weighed in the field.

Schlee (1957), Pelletier (1958), and Yeakel (1959), in particular have demonstrated the usefulness of the largest fragments found at each exposure in determining distribution patterns. In this study the largest particle found in place or at the foot of the exposure near the site of channel sample was selected to represent the maximum particle size present at each sample locality. The maximum, intermediate, and minimum diameters (Krumbein, 1941) of these particles as well as the lithology and mass were recorded in the field. In all cases it was attempted to avoid ice-rafted material as judged by the criteria of unusual size, angularity, and preservation of delicate surficial features. The objectivity of this procedure

may be questioned but it is rapid and seems safe when used with other criteria of transport direction.

Estimates of the composition of the coarse fraction of the sediments were obtained by classifying 100 pebbles according to lithologic type at several representative localities. Only pebbles with intermediate diameters of between 1 and 2 inches were used. These were selected by sieving large channel samples taken vertically through the exposures.

Laboratory Procedures

U.S. Standard sieves in increments of whole phi units from 4ϕ to -6ϕ (0.062 to 64 mm) were used for the mechanical analyses. After a preliminary disaggregation with mortar and rubber-tipped pestle the samples were mechanically shaken in the nest of sieves for 10 minutes. Results, in weight percentages, were plotted in phi units (Krumbein, 1934) on linear probability paper (Doeglas, 1946). Statistical measures were derived from the cumulative frequency curves according to certain of the formulae proposed by Inman (1952). It was occasionally necessary to extend the curves by small amounts in order to obtain values of ϕ 84. These values and the parameters derived from them are indicated as estimates in the tabulation of textural properties (Appendix III). Mechanical composition was also plotted as histograms. In three cases it was necessary to employ hydrometry in addition to sieving because of relatively large amounts of material smaller than 4ϕ . Hydrometry methods have been discussed by Krumbein and Pettijohn (1938, p. 172-174) and the method used here is essentially that specified by the American Association of State Highway Officials (1950, p. 215-224).

Systematic investigation of the mineralogy of the sediments was confined to the size range between 1ϕ and 4ϕ (500 to 62 μ) from one sample from each quadrangle. Material of this size was selected from the splits resulting from the mechanical analysis and reduced in size to about 20 grams by successive passes through a microsplit. "Light" and "heavy" fractions were separated in separatory funnels by flotation in tetrabromoethane after cleaning with hot dilute hydrochloric and nitric acids. The density of the tetrabromoethane was maintained within the limits 2.90-2.95. Samples were weighed before and after these separations so that the weight percentage of heavy minerals in the selected size fraction could be computed. The "heavy" fraction of each sample was split to appropriate size and mounted in Canada balsam for petrographic examination. At least 100 non-opaque grains were identified for each sample.

Quartz and feldspar dominate the "light" fractions and in order to facilitate rapid distinction between them a staining technique similar to that described by Bailey and Stevens (1960) was employed. In order to manipulate grains in this procedure (which was originally described for slabs and thin sections) the grains were mounted in a very thin layer of Canada balsam on a petrographic slide in such a way that most of the grain was exposed. Staining is accomplished by exposure to the fumes of hydrofluoric acid and treatment with solutions of sodium cobaltinitrite (stains potash feldspars yellow) and rhodozonic acid (stains plagioclase red).

DISPERSAL PATTERN

A formidable body of evidence relating to the usefulness of various structures as current-direction indicators has been amassed since Sorby urged the study of

them as early as 1859. Cross-bedding is the most generally abundant structure possessing directional significance. It is certainly the only feature of this nature to be found in quantity in the sediments studied here.

The importance of cross-bedding to paleocurrent study rests primarily on three factors: (1) foreset laminations are formed by local current action so that the dip direction is downcurrent; (2) the direction of flow of a given current, such as a stream, is adequately reflected in the dip directions of foresets within the cross-bedding resulting from that current; (3) meaningful areal dispersal patterns result from the measurement of sufficient randomly selected dip directions. The validity of these assumptions has been attested by numerous studies in which independent evidence was available, as well as in experimental situations. Potter and Pettijohn (1963) have summarized this work, and have shown that cross-bedding has major importance as a paleocurrent direction indicator.

Cross-bedding sufficiently well developed to measure was found at 86 of the 150 sample localities used in the present study. At other localities, especially those where only a small section is exposed, and in relatively homogeneous materials, cross-bedding was either absent or not apparent under field conditions. Individual foreset beds are usually discernible because of differences in texture, although differences in composition, such as heavy mineral concentrations, and iron and manganese oxide staining are sometimes helpful. In these unconsolidated rocks structures are only occasionally etched into relief by differential weathering. Sometimes features visible on a weathered surface cannot be seen when the surface is cleaned off; however, scraping the fresh surface with the shovel blade held normal to the surface tends to restore textural contrasts.

Probably as a result of recent increased interest in cross-bedding, the problem of classification of various types of cross-bedding has received special attention. Prominent among those who have studied this problem are McKee and Weir (1953), Pettijohn (1962), Allen (1963), and Potter and Pettijohn (1963). Observations in the surficial sediments of Delaware seem to confirm the concept of Pettijohn (1962) that there exist only two major types of cross-beds, referred to in that paper as "planar" and "festoon" and later (Potter and Pettijohn, 1963) termed "tabular" and "trough." As indicated by these authors, gradations do exist between the two principal types and in the present case all units adequately exposed to study could be assigned to a position between these two end members.

Tabular cross-bedding predominates in the Columbia sediments of Delaware. Although all of the cross-beds in which dip azimuths were measured could not be accurately classified because, in some cases, not enough of the unit could be seen, 310, or 87 percent, of the 356 cross-beds typed were judged to be tabular, or close to it, as opposed to the trough type. Trough type cross-beds occurred grouped in certain outcrops, or scattered singly. No areal or other pattern of occurrence was obvious though but few observations were made. It is possible that, if all of the cross-beds were well exposed, some of the seemingly tabular units might be seen to be large trough cross-beds.

Of 319 cases in which the bottom contact of the foreset beds could be clearly characterized, 83 percent were non-tangential. Tangential and non-tangential lower contacts of foreset beds occur in both tabular and trough type cross-beds, although on the basis of the meager number of observations made here it appears that trough type cross-beds tend to have tangential contacts.

A great variety of textures are present in individual cross-beds. As Potter and Pettijohn (1963, p. 35) have observed, cross-bedding may be obscure in very coarse sediment. It may be possible that there is an upper limit to the coarseness of sedi-

ments in which cross-bedding is formed. Four examples of the mechanical composition of cross-beds in Delaware are shown in figure 1. The sample from Cc-II is, by visual estimate, about as coarse textured as any cross-bed found in these deposits. Coarser material is available but does not appear to be incorporated into cross-beds. It is found in ordinary beds of gravel. The few analyses made of individual cross-beds seem to confirm field observation that the sediments of cross-beds may vary greatly in grain size, sorting, and skewness and may be uni- or bimodal.

The thickness, or scale, of the cross-beds examined was also measured. These measurements were taken at the places where the dip and dip directions were measured on each cross-bed and they generally represent the clearest development of the unit. In a few instances the lower contact of thick cross-beds could not be exposed. Visible thicknesses range from 10 feet to 0.2 feet. The mean for 396 measurements is 1.48 feet. A histogram showing the distribution of cross-bed thicknesses is presented as figure 2.

Dips of foreset beds have been determined by several authors and Potter and Pettijohn (1963, p. 79) conclude that the average dip is generally 18 to 25 degrees. The mean of 400 such measurements in the present study is 22.1 degrees. No values lower than 10 degrees were accepted and the few values between 35° and 40° (the maximum recorded) may be the result of deformation in view of the maximum angle of repose to be expected in sands. The distribution of cross-bed foreset dip angles is shown in figure 3.

An estimation of the trends of the paleocurrent system which dispersed the sediments may be determined from the dip directions of foreset beds within cross-beds. A total of 400 measurements of foreset dip direction was obtained and recorded as azimuths. Reiche (1938) and Krumbein (1939) recognized, as do later authors (Chayes, 1954; Tanner, 1955; Curray, 1956; Potter and Pettijohn, 1963), that measures of the central tendencies of circular distributions must differ from those of linear data because the measures for circular distributions should be independent of the starting point. The simple arithmetic mean of values spread over more than 180° may be misleading, as, for example, in the case of the mean of 359° and 1°. The vector mean of Reiche (1938) provides a measure of central tendency applicable to circular distributions. This vector mean was utilized for individual outcrops where the spread of the readings exceeded 180° and for the means of the moving average and the grand mean. It was computed trigonometrically using azimuths grouped into 30° intervals. Details of this method are presented by Curray (1956) and Potter and Pettijohn (1963, p. 264). The terminology of these authors has been followed and the formulae used in computations of the vector mean are presented in Appendix II.

The locations of exposures where dip directions were obtained is shown in plate 2. The number of measurements at each site is given and the mean, or vector mean, azimuth indicated. The map shows a general trend in paleocurrent directions which may be visually approximated as north to south. It is apparent that the deposits in the northern portion of the area are more abundantly cross-bedded than those in the south. Areas poor in information or showing anomalous directions may be contrasted with those in which data are plentiful and consistent.

The same dip azimuths obtained from the measurement of cross-bedding in the field are presented as a moving average diagram in plate 3. The moving average provides a smoothing effect and a means of interpolating between the actual field localities. Its use is recommended by Pettijohn (1962) and Pettijohn and Potter (1963) and it has been used in practice by Potter (1955), Schlee (1957), Pelletier

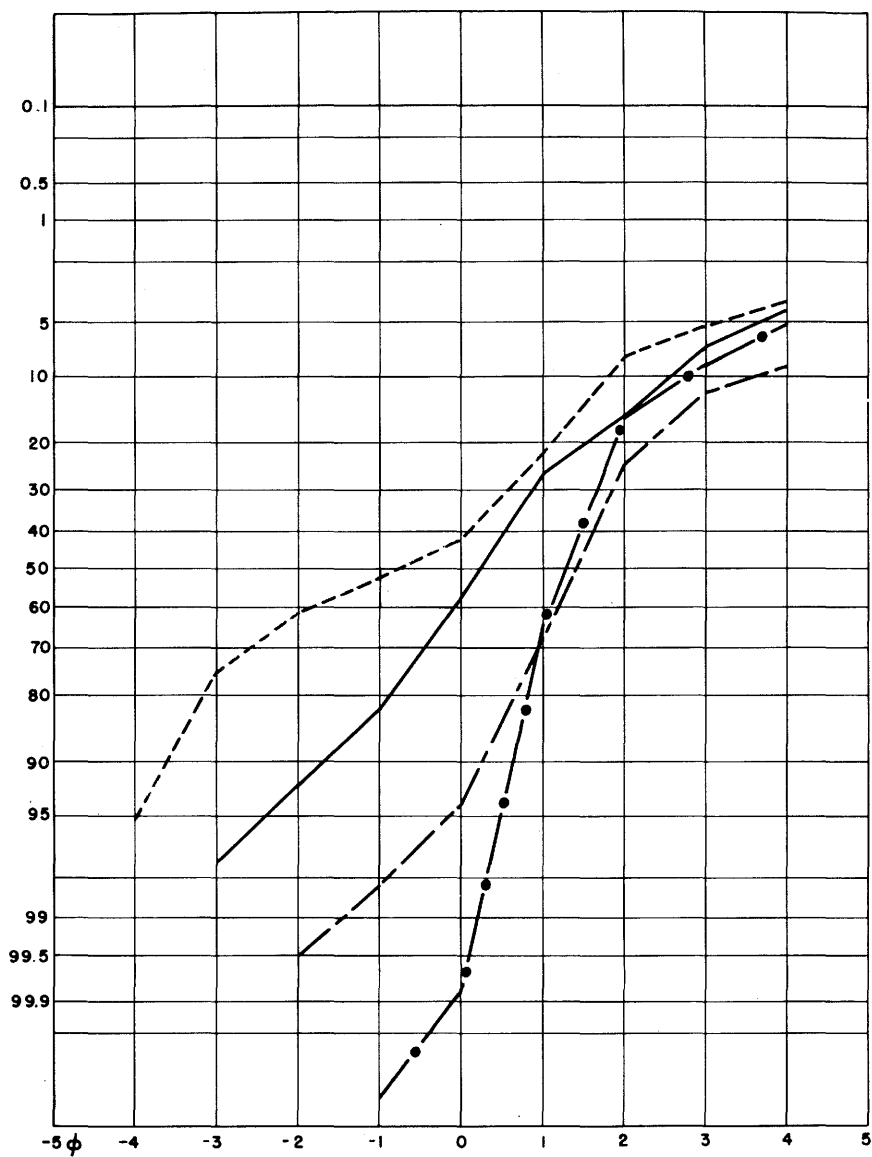


Figure 1. Representative cumulative curves of cross-beds.

Location Cc-II — — — —
 Cd-II ······
 Jd-I —————
 Mf-I —●—

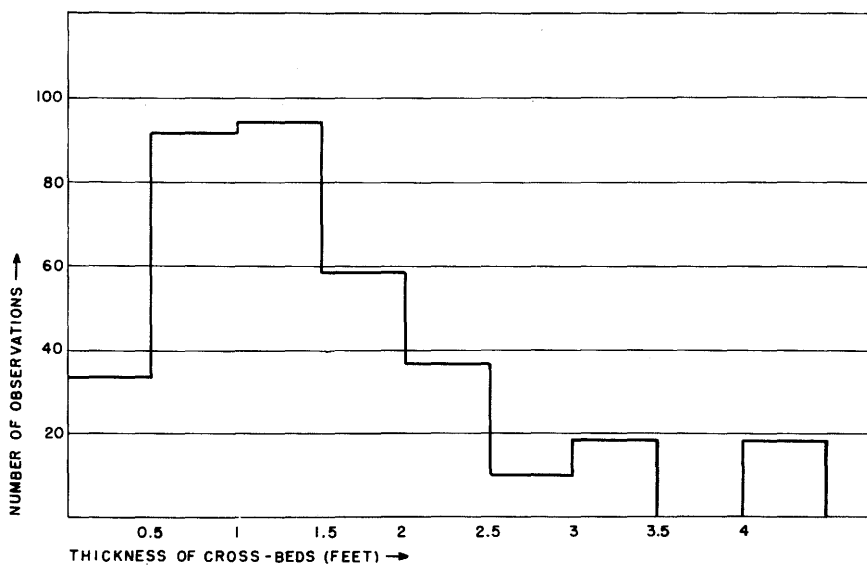


Figure 2. Distribution of thicknesses of cross-beds.

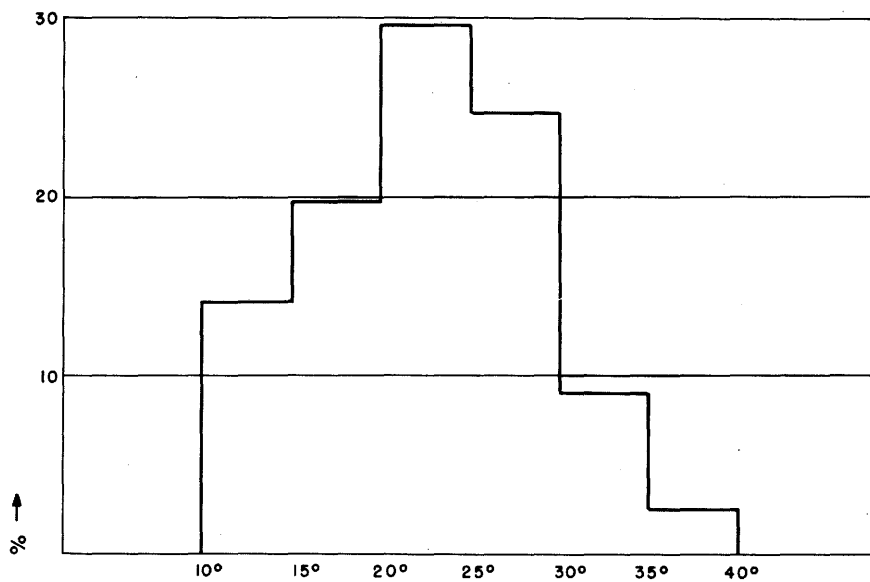


Figure 3. Distribution of foreset dip angles. All dips of less than 10° were rejected.

(1958), Yeakel (1962), and Burtner (1963). The moving average is constructed by computing the average of values, in this case dip azimuths, derived from all localities within the four grid squares whose adjacent corners are formed by the intersection of two grid lines. The average value is indicated at the intersection of the grid lines and the process is repeated for each such intersection. In this manner each measurement is used four times.

Plate 3 has been constructed by using the vector mean of the individual dip azimuth measurements. The reliability of each vector shown is a function of the consistency and the number of azimuths represented. The length of the resultant vector is a measure of the directional spread of the azimuths. This divided by the number of measurements yields the consistency ratio of Reiche (1938). It has become more common recently to express the consistency ratio in terms of percent (L of Curray, 1956) who has shown that L, the vector magnitude, may be related to the Rayleigh test of significance. The chart presented by Curray provides a convenient method of determining the level of significance which may be attached to any computation of vector means. All vector means for which the level of significance does not exceed a value of 0.05, that is, those in which there are more than 5 chances in 100 of its being due to chance, are so indicated in plate 3. In most cases significance levels of 10^{-3} to 10^{-5} are found for the values shown in the moving average.

A grand vector mean of 170° has been calculated from all of the individual measurements of dip azimuths. This is shown graphically with a circular histogram indicating the spread of the individual values. This grand vector mean has a vector magnitude of 51.5 percent which, for 400 measurements, gives a very high level of significance.

The pattern evolved from the moving average suggests that the streams which distributed the Columbia sediments of Delaware departed southwestward from the present path of the Delaware River in the Wilmington to Delaware City area. The paleocurrents appear to have swept farther to the west, and, when reconstructed, join the Delaware River as tangents to its present curve to the southeast into Delaware Bay. In the central portion of Delaware a strong southerly trend is present which fans out toward the south where a more easterly tendency may be noted. As a generalization it may be stated that the paleocurrent system flowed from north to south.

Schwartzacher (1953), Pelletier (1958), and Burtner (1963) found that the thickness of cross-beds decreases in the down-current direction in the units which they studied. This is also true of the cross-beds in the deposits investigated here. If the transport direction may be characterized as north to south, a crude representation of the change of thickness of cross-beds with distance down-current may be constructed by averaging the thickness of all of the cross-beds measured in each east-west row of quadrangles, roughly normal to the transport direction, and plotting these values against north-south direction. This is represented in the histogram of figure 4 from which it may be seen that, despite irregularities, there is a decrease in the average thickness of cross-beds toward the south (down-current). A similar plot for foreset dip (figure 5) appears to show no systematic variation with distance.

TEXTURE

The mechanical compositions of the 150 samples collected systematically have been determined. Twenty additional analyses were made on samples from indi-

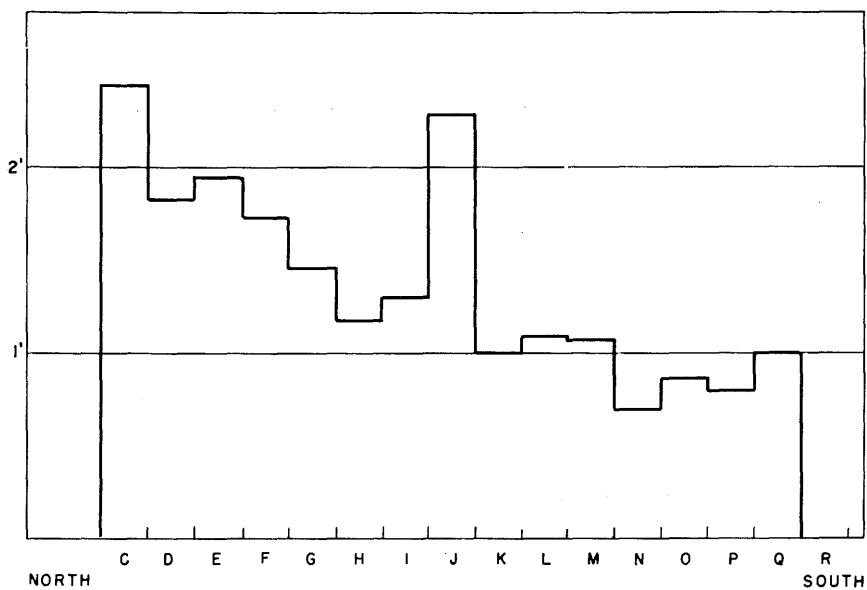


Figure 4. Average cross-bed thickness for each east-west row of quadrangles arranged to show decrease to the south.

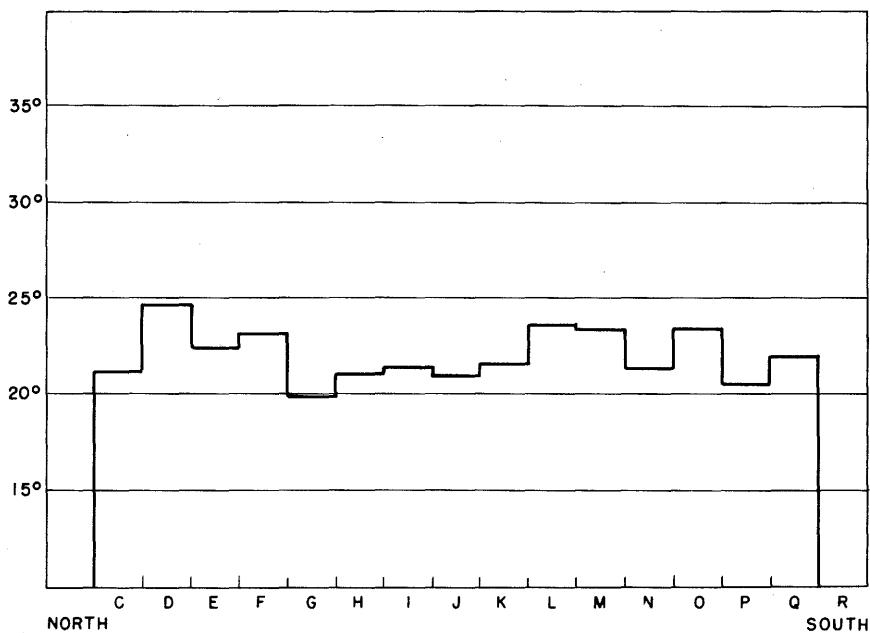


Figure 5. Average foreset dip angles for each east-west row of quadrangles showing little variation from north to south.

vidual beds or from exposures not included in the systematic sampling. In each case cumulative frequency curves and histograms were plotted. Statistical measures of the median grain size, sorting, and skewness were computed according to the methods proposed by Inman (1952). The formulae used in these computations are given in Appendix II and the parameters determined for each sample are tabulated in Appendix III. Advantages of Inman's procedures are that a large portion of the cumulative frequency curve is measured (84th to 16th percentiles, or one standard deviation either side of the mean), and that measures of sorting are directly comparable.

The measure of the central tendency of the size distribution is determined at the 50th percentile on the cumulative frequency curve and is the median diameter, which, when expressed in phi units, is designated $Md\phi$. The measures of sorting and skewness, also expressed in phi units are designated $\sigma\phi$ and $\alpha\phi$, respectively.

The median grain size may be used to characterize the sediment in terms of the Wentworth scale textural terminology. Most of the samples of the surficial sands of Delaware have median grain sizes between 1ϕ and 2ϕ and are therefore medium sands. Extreme coarse and fine values range from -2.40ϕ to 3.51ϕ , or from fine pebble gravel to very fine sand. The distribution of values of $Md\phi$ is shown in figure 6. It should be noted that almost all of the samples are sands, mostly medium sands, although they represent random sections which may include some fine and some coarse beds. The mass of sediment may be most accurately described as a medium sand. The mean of all values of $Md\phi$ is 1.30.

The areal distribution of the median grain size of the surficial sands is shown in plate 4. A moving average has been contoured in order that major trends may be identified. From the figure it may be seen that the median grain size of the sands decreases from north to south, i.e., in the downstream direction as indicated by the analysis of foreset dip azimuths. The coarsest textures occur in the northeastern corner, which is where the paleocurrents entered the area. This downstream decrease in grain size is also indicated, in simple fashion, in figure 7 where the means of the median diameters of all samples in each east-west row of grid quadrangles (roughly normal to the 170° grand vector mean current direction) are plotted against north-south distance.

The contours of the moving average of plate 4 indicate a decrease in $Md\phi$ from 0ϕ to 2ϕ in a distance of approximately 80 miles. This decrease, although fairly regular, does not seem to admit to expression as a mathematical statement, probably because more than one environment is represented and because the competency of the transporting media varied widely from place to place in the time interval represented by the samples. This decrease in median grain size with distance down stream is expected and is generally attributed to a decrease in the competency of the transporting medium. Kuenen (1959) has shown that the rounding of sand-size particles in transport is such a slow process that wear cannot be considered a major factor. Streams debouching upon the Coastal Plain from the Piedmont Upland suffer reductions in gradient and therefore competency after passing through the Fall Zone.

In addition to the cumulative curves, the mechanical composition of a rock may be represented by a histogram. The sediment may be classified according to the size interval in which the greatest weight of material occurs—the modal class. The modal class of 64 percent of the systematic samples is $1\phi - 2\phi$, which agrees well with the values of $Md\phi$ determined from the cumulative frequency curves. In almost all of the analyses in which the $1\phi - 2\phi$ interval does not contain the

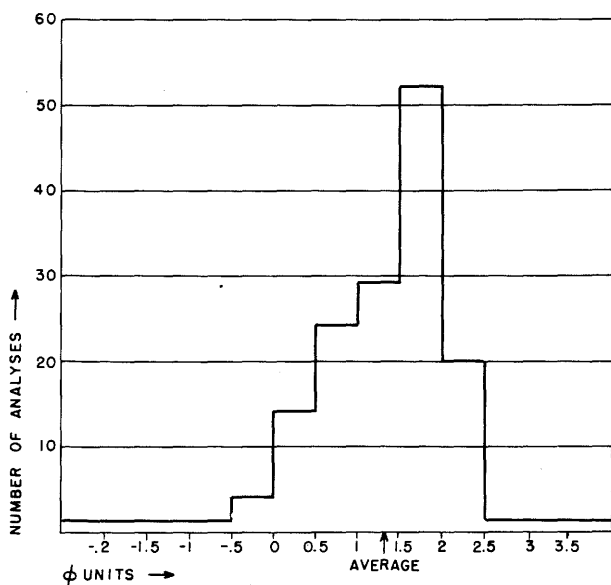


Figure 6. Distribution of median grain sizes ($Md \phi$).

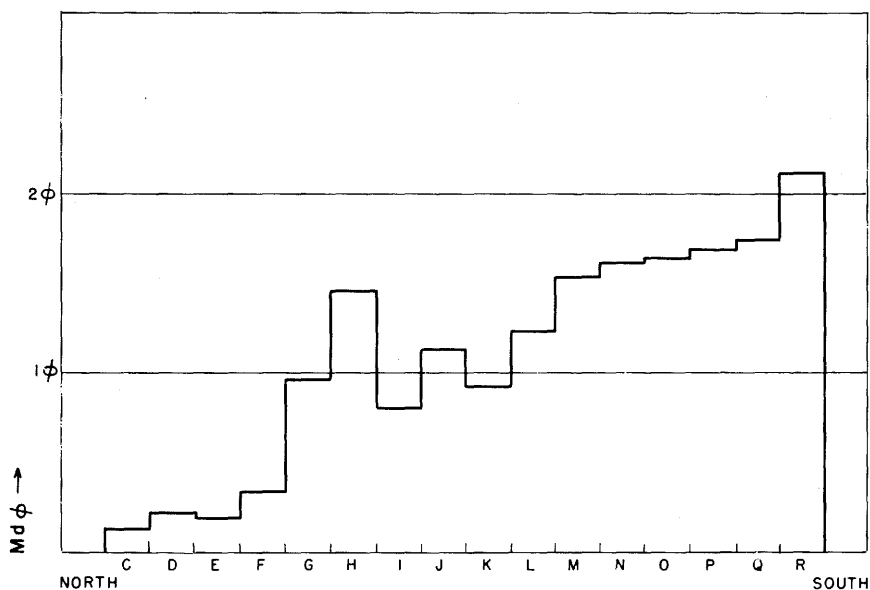


Figure 7. Average median grain size for each east-west row of quadrangles arranged to show decrease to the south.

greatest weight of sediment, the modal class occupies one of the immediately adjacent intervals.

Other maxima may occur in the particle size distribution of a sediment. Such polymodal distributions result from the addition of admixtures removed in size from the modal class. Of the 150 systematic samples, 71, or 47 percent, are bimodal and an additional 9, or 6 percent, have three modes. Some of the secondary modes are very weak. The polymodal samples are concentrated in the northern part of the study area and almost always reflect a coarse admixture in the form of gravel beds. Most of the secondary modes occur between -6ϕ and -3ϕ .

The mean of the sorting coefficients, $\sigma\phi$, for 150 analyses is 1.58. Although there are wide deviations from this mean value (extreme values of $\sigma\phi$ are 0.51 and 4.72), the mean may be loosely interpreted as indicating that the sands are moderately well sorted over-all. The distribution of values of $\sigma\phi$ is shown in figure 8. Individual beds are, of course, better sorted than the composite channel samples which may include many beds. Contours based on a moving average of the sorting coefficients are shown on plate 5. Considerable fluctuation in $\sigma\phi$ may be observed from station to station. This local variation appears to be mainly a function of the total thickness of exceptionally coarse, or fine, beds which the channel sample happened to intersect. There is however, a regional tendency for the sorting to improve from north to south (downcurrent). This may be judged from the contours of plate 5 or the graph of figure 9, notwithstanding the presence of some irregularities and anomalies. The tendency toward increased sorting downcurrent is of interest considering the general lack of documentation of such trends in the literature indicated by Pettijohn (1957, p. 542). The downcurrent increase in sorting may indicate that the decrease in abundance, grain size, and thickness of gravel beds is not compensated, in this area, by the increased deposition of silt and clay.

Skewness of the cumulative frequency curves has also been measured, using the value $\alpha\phi$. Skewness values range from strongly positive (0.678) to strongly negative (-0.848). The distribution of these values is shown in figure 10. Fifty-seven percent of the samples are negatively skewed. Negative skewness indicates an excess of coarse-grained material above a symmetrical distribution and positive values indicate an excess of fines. The areal distribution of $\alpha\phi$ values is complex and rather irregular. The significance of the skewness of size distributions has been discussed by several authors; prominent among them are Mason and Folk (1958) and Friedman (1961, 1962). The skewness of a size distribution may reflect the energy regimen of the depositing agent. If the energy of the agent of deposition rarely falls below a relatively high threshold, as with waves, material finer than a certain size will not be deposited and the sediment should display negative skewness because of the resulting excess of coarse detritus. Conversely, if the transporting ability of the depositing agent rarely exceeds a certain threshold, as with wind, the resulting deposit may be deficient in coarse grains and therefore have an excess of fine material and positive skewness. This thesis ignores complicating factors such as the availability of the various sizes and is probably only applicable to rather large values of either positive or negative skewness.

The relationships between positive and negative skewness, presence or absence of secondary maxima, and clay and silt content of the sands are illustrated in plate 6. Samples plotted as bimodal here are those which have distinct, as opposed to weak, secondary modes. Certain general observations may be made from plate 6: negatively skewed samples, bimodal distributions, and smaller admixtures of

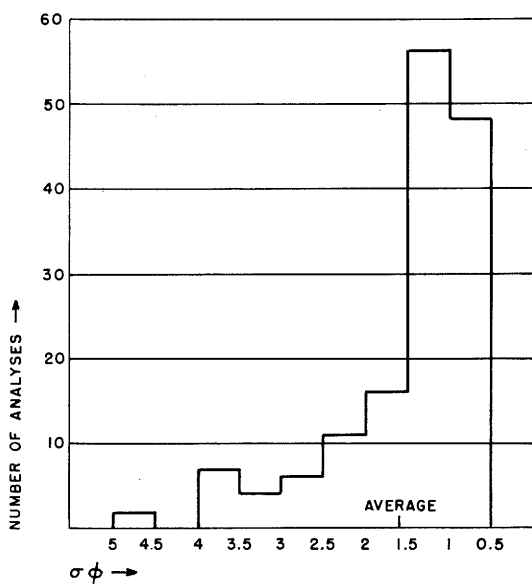


Figure 8. Distribution of sorting coefficients ($\sigma\phi$).

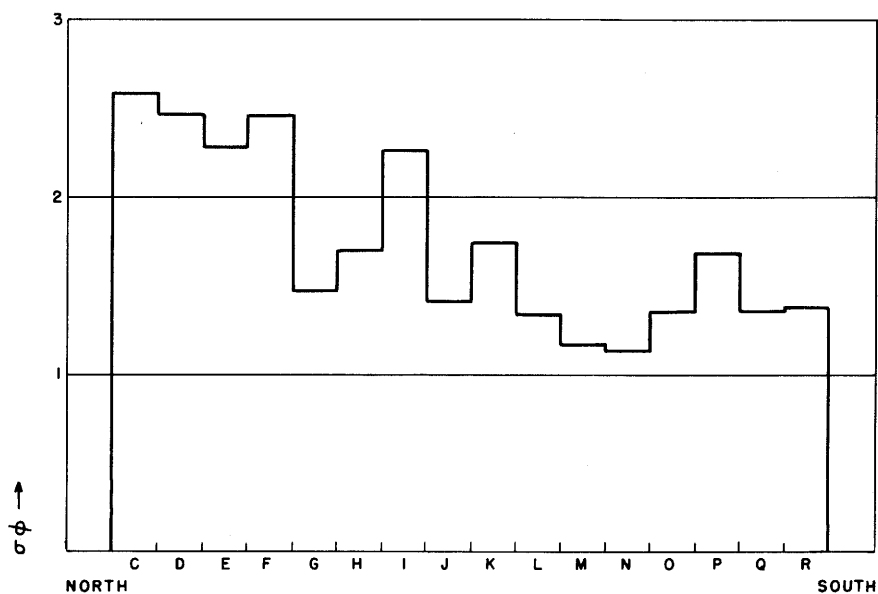


Figure 9. Average sorting coefficients ($\sigma\phi$) for each east-west row of quadrangles arranged to show better sorting toward the south.

silt and clay all tend to occur in the north; most of the positively skewed, unimodal samples occur in the south where the clay and silt content is above average; most negatively skewed samples are bimodal; most positively skewed samples are unimodal.

The dominantly negative skewness of the sands in northern Delaware reflects the presence of coarse beds as indicated by the bimodal size distributions of most of these same samples. Toward the south many of the negatively skewed samples are not also bimodal, which suggests that their negative skewness is due to a deficiency of fines rather than the addition of gravel. Large positive skewness values correlate with the presence of distinct silt beds. Smaller positive values may indicate a lack of coarse material in well-sorted sands or the presence of silt and clay mixed with sand in poorly sorted deposits.

The largest fragment present in a sedimentary deposit provides another measure of the areal variation of texture. A review of the technique and its applications is found in Potter and Pettijohn (1963, p. 202-205). In the present investigation the largest particle found in place or in talus at the base of the face where the channel samples were taken was selected for study. There are obvious possibilities for subjective error in the selection of the largest particle which can be reduced by averaging the measurements from a number of particles, however, in the materials studied it was usually not difficult to make a reasonable selection. The largest size class present in each of the mechanical analyses provides a figure based upon more objective procedure. Both the largest single particle and the largest size grade of the channel samples have been considered and may be compared.

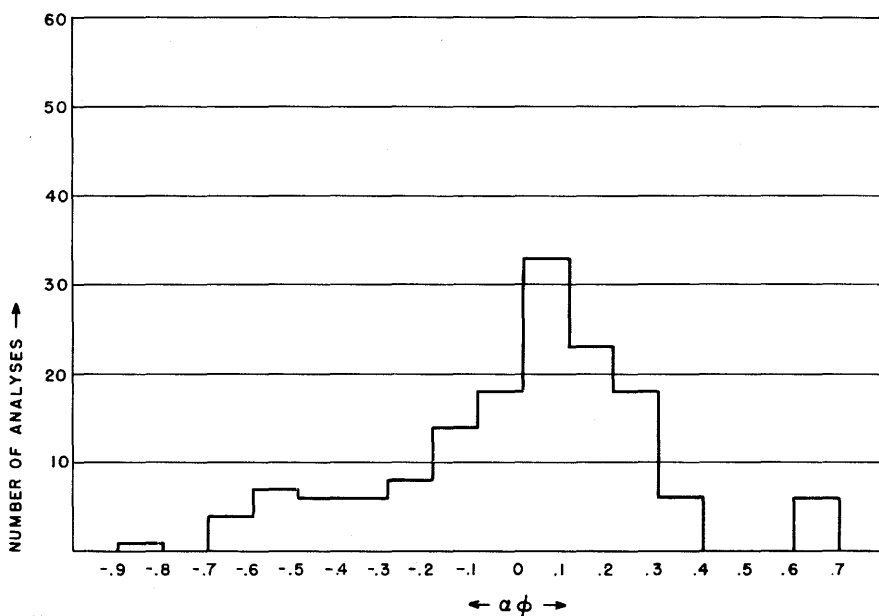


Figure 10. Distribution of skewness values ($\alpha\phi$).

Caution must be exercised when considering the larger particles of the Columbia sands because of the presence of ice-rafted blocks. As ice-rafted blocks have a different transportation history from normal fluvial detritus they have been avoided. Ice-rafted blocks are generally characterized by lack of wear commensurate with their size. This is indicated by the preservation of sharp corners and delicate bedding-plane structures such as ripple marks. The blocks may also be of abnormally large size and of less durable composition than the gravels in which they are found.

The results of measuring intermediate diameters and weights of individual particles selected as the largest at each station are shown in plate 7. This may be compared with plate 8 in which the areal distribution of maximum grade sizes is shown. Each measure shows the same essential feature: a north to south decrease in size.

The pattern displayed by the contours of the moving average of the maximum grade sizes is the most regular of those shown in plates 7 and 8. The largest grade class decreases from -6ϕ to -2ϕ in a distance of about 80 miles. Moreover, the contours are subequally spaced (especially those of -6 , -5 and -4ϕ) suggesting agreement with Sternberg's Law which calls for an exponential downstream decrease in pebble size. Discussions of Sternberg's Law are provided by Schlee (1957), Pettijohn (1962), and Potter and Pettijohn (1963) among others. The downstream decrease in size is somewhat more erratically expressed in the plots of the intermediate diameters and masses of single large particles.

The texture of a sediment is a result of a more or less complex history of erosion and transportation in which the final and most influential stage is deposition. Each statistical measure of texture reflects some aspect of this history and, as they are all related through a particular sample, should show some relationships to other parameters. These relationships between textural parameters may be as useful as the parameters themselves in determining the environment of deposition of the sediment (e.g., Friedman, 1961, 1962).

In figure 11 the median grain size ($Md\phi$) and sorting ($\sigma\phi$) characteristics have been plotted for the 150 samples. Although there is considerable spread, it may be seen that, as median grain size decreases, sorting tends to increase. The two samples in which $Md\phi$ is greater than 3, however, show relatively poor sorting. This distribution is in agreement with the findings of Inman (1949) which indicate the existence of an optimum size, fine sand, above and below which sorting decreases. Of the samples studied, few are as fine or finer than fine sand and so the plot shows clearly only the case of decreasing sorting above the critical size.

The coefficient of sorting ($\alpha\phi$), has been plotted against the coefficient of skewness ($\alpha\phi$), in figure 12. A tendency for the better-sorted samples to also be less skewed may be discerned. As the degree of sorting decreases the skewness tends to increase in either positive or negative values of $\alpha\phi$. The sorting values reflect the degree of fluctuation in competency of the depositing currents (Groot, 1955) during the periods of time represented by the channel samples. The skewness values, as suggested above, may indicate whether the range of fluctuation tended to exceed a high minimum competency, resulting in a paucity of finer material and negative skewness, or whether the range was below a certain low maximum, resulting in an excess of fines and positive skewness. As the range of fluctuation of current competency increases (i.e., sorting decreases) the fluctuations have greater tendencies to become asymmetrical (skewness increases, positive or negative).

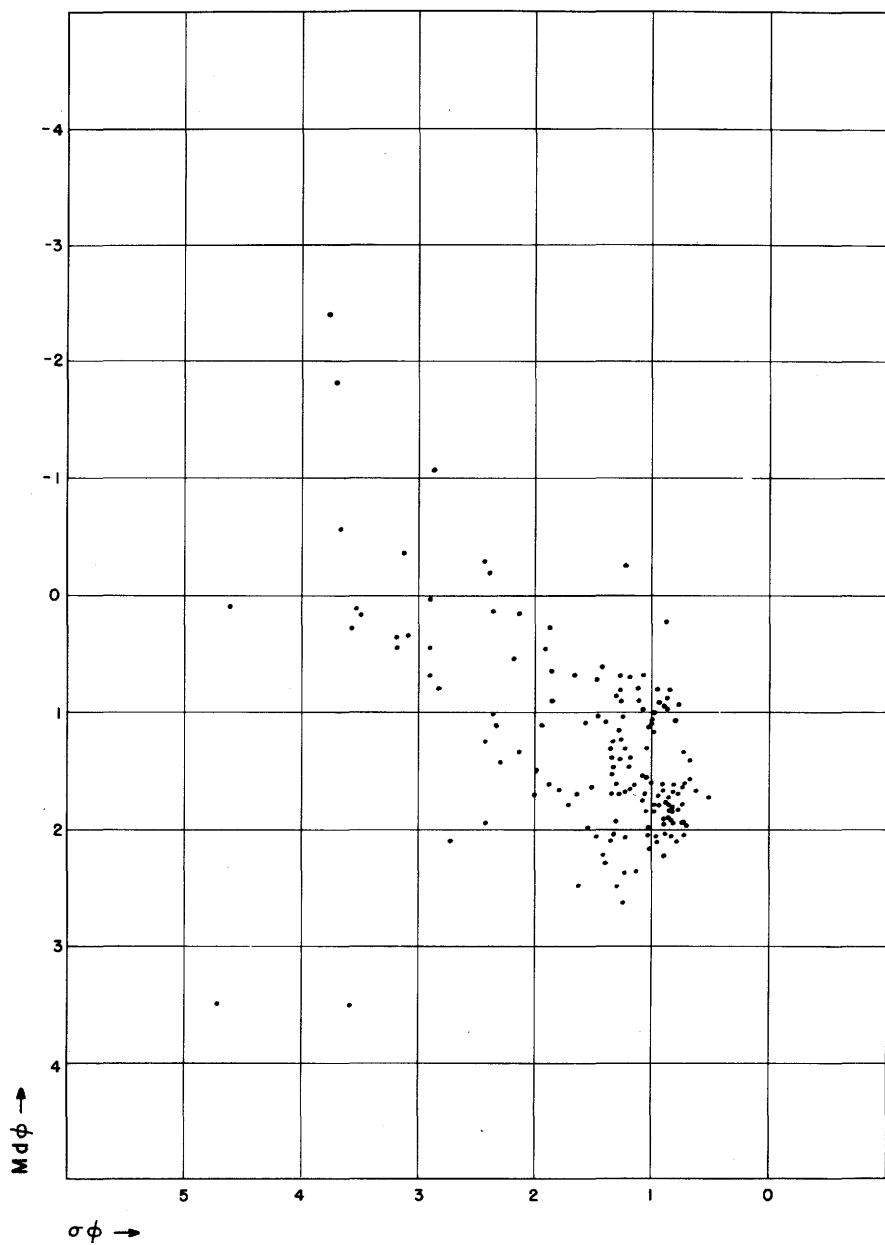


Figure 11. Relationship between median grain sizes ($Md \phi$) and sorting ($\sigma \phi$).

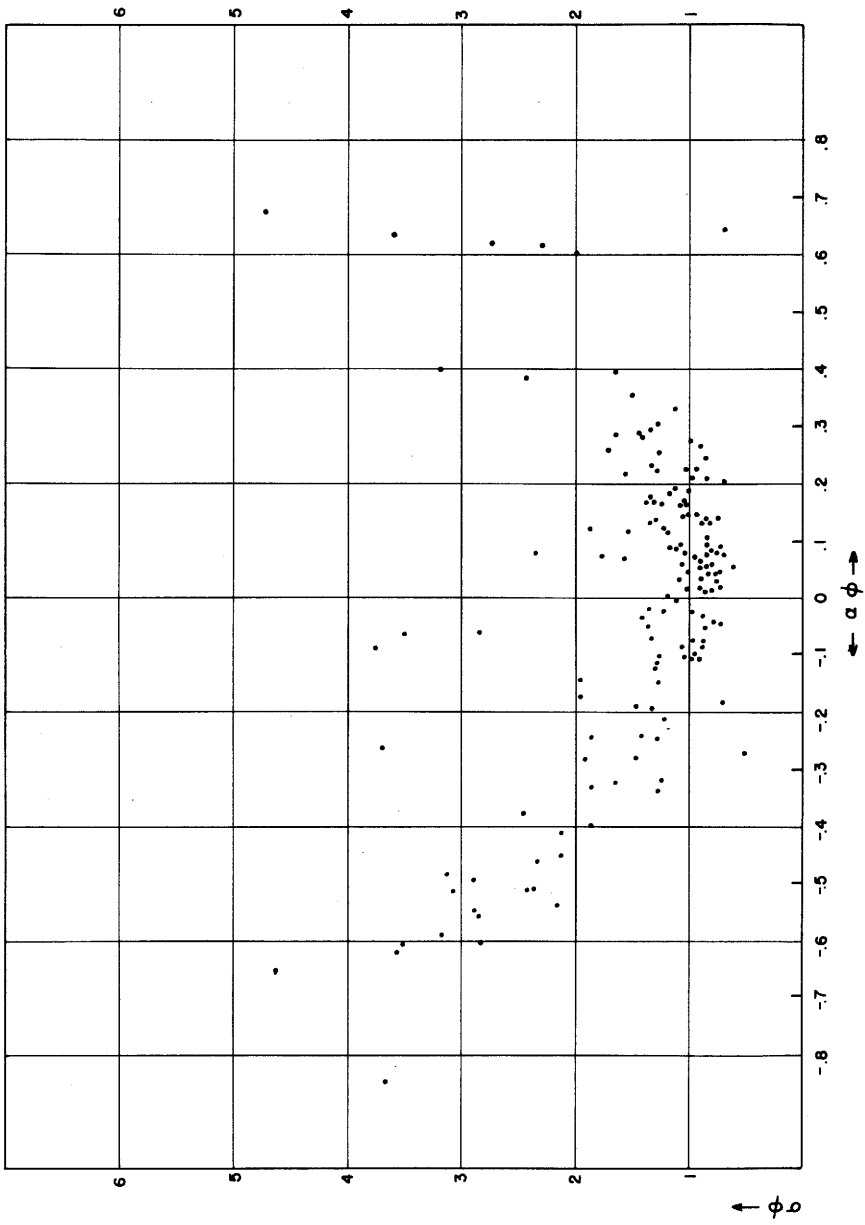


Figure 12. Relationship between sorting ($\sigma \phi$) and skewness ($\alpha \phi$).

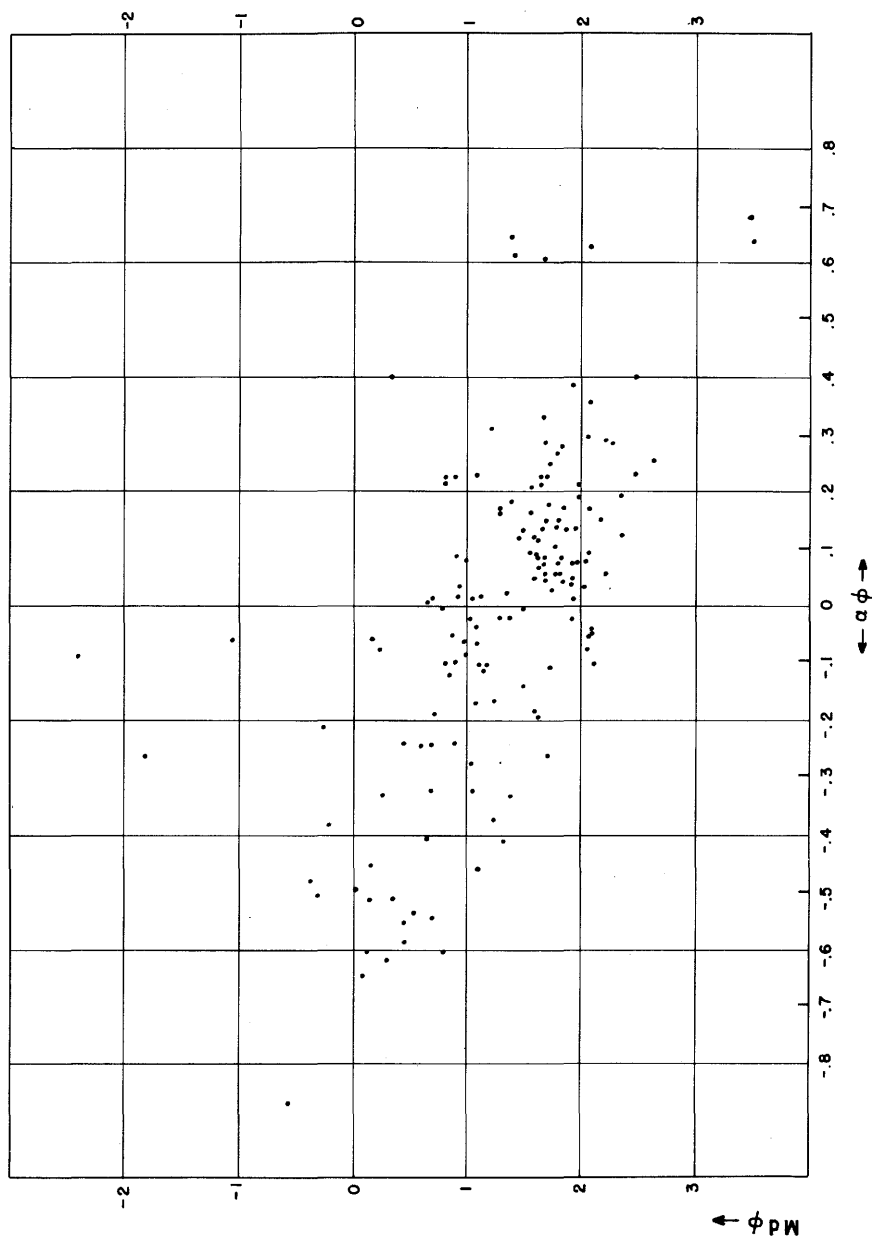


Figure 13. Relationship between median grain sizes ($Md \phi$) and skewness ($\alpha \phi$).

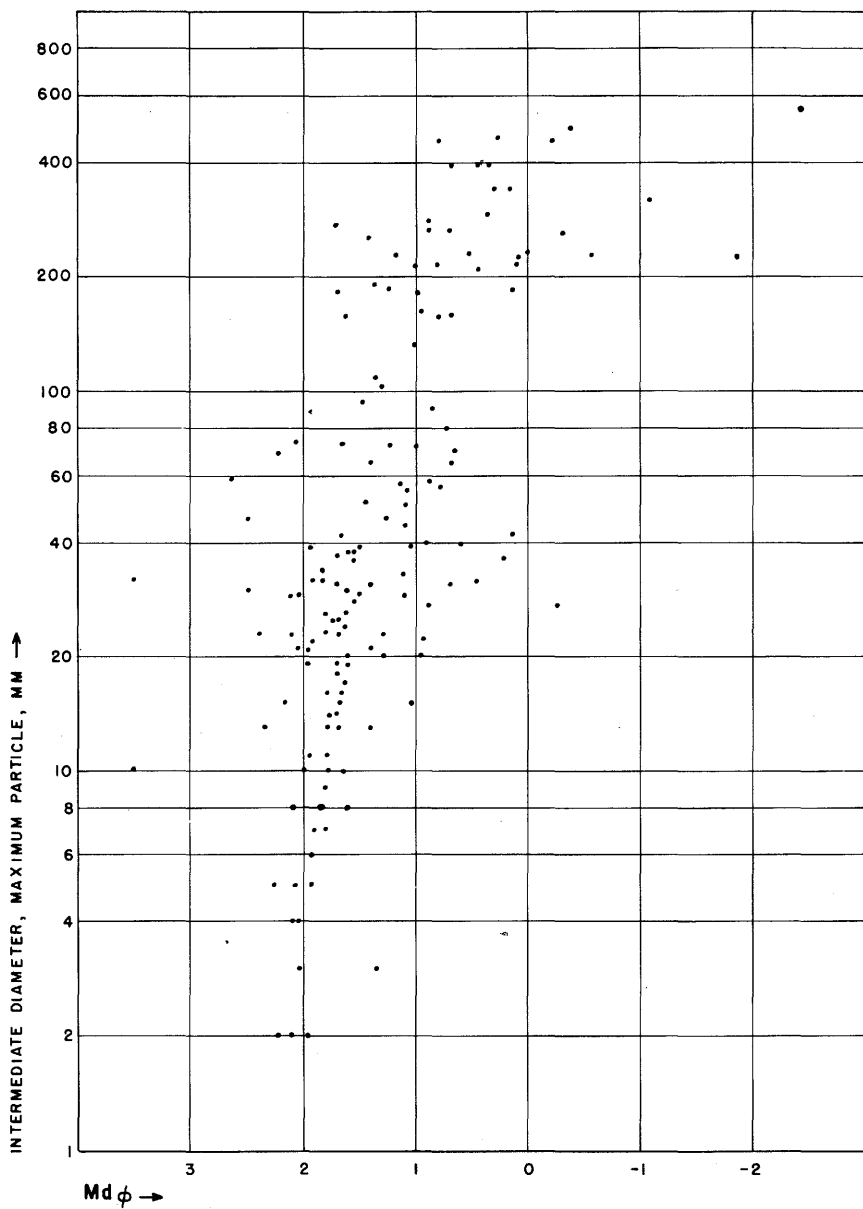


Figure 14. Relationship between intermediate diameters of largest particles and median grain sizes ($Md \phi$).

Median grain size and skewness have been considered together in figure 13. A correlation is present between increasing grain size and decreasing (and negatively increasing) skewness in the sands studied. The coarser-textured sands tend to contain gravel beds which result in bimodal grain size distributions with a coarse admixture causing negative skewness. Fine-grained sands commonly contain even finer admixtures which result in positive skewness.

Median grain size and maximum particle size have been used to indicate the degree of coarseness of the sands. Both have been found to indicate a down-current decrease in grain size and should, ideally, show a linear correlation. Median grain sizes in phi units have been plotted against the intermediate diameters of the largest particles in millimeters on a logarithmic scale in figure 14. As the median diameter decreases, the maximum particle size decreases rapidly. There is, however, much spread in the plot, especially in the coarser sizes. Either parameter provides a scalar measure of directional trends but the spread suggests that the use of a large number of samples is desirable.

COMPOSITION

Gross Mineralogy

The mineralogy of the sand fraction of one sample from each of the five-minute quadrangles was studied. Within each quadrangle the sample representing the thickest stratigraphic sequence was selected. The study of the mineralogy of the sands is therefore based on 75 samples rather evenly distributed over the study area.

Splits in the size range 62 to 500 microns were treated in order to stain the feldspars by the procedure of Bailey and Stevens (1960) which was modified for use with grains. Two hundred grains were identified petrographically in each sample. The staining procedure did not result in a perfect separation of quartz, potash feldspar, and plagioclase as was intended. Most of the difficulty may be attributed to grains already stained naturally and to clay coatings which caused the quartz grains to appear to be stained. Ultrasonic cleaning of the grains alleviates these problems. A tabulation of the results of the study of the mineral composition of the "light" fraction of the sands is presented as Appendix III. Quartz, potash feldspar, plagioclase, mica, and "others" were differentiated. The mica is muscovite, some of which is partly altered, and "others" includes rock fragments, chert, and unidentified aggregates.

The composition of the sands is strongly dominated by quartz, which averages 80.4 percent in the samples examined. The feldspar content of the sands is somewhat variable and averages 18.4 percent. The amount of feldspar is greater than was expected, but is fairly typical for Pleistocene and Recent sands when compared to those listed by Pettijohn (1957, p. 123). In general, potash feldspar is about 5 times as abundant as plagioclase.

Expressed as an overall average, the mica content of the sands is about 0.5 percent. The amount of mica present varies considerably between localities. This is indicated by field observation and confirmed by the grain counts. Considering the shape of mica particles, it is to be expected that they will sort and concentrate in seemingly erratic fashion in a quartz-feldspar population. Rock fragments, chert, and aggregates comprise about one percent of the 62 to 500 micron sand fraction and are somewhat more abundant in the larger grades.

As a mass, the sands are subarkosic in the usage of Pettijohn (1957).

The composition of the gravel and cobble portions of the deposits is significant because of the possibility of discovering fragments sufficiently distinctive to permit identification of their source ledges and thereby provide an indication of provenance.

The composition of the coarse grades is dominated by the most resistant common material, silica, in the form of sandstone (mostly quartzite), vein quartz, and chert. A striking feature is the variation of composition with size. In table 1 the lithologies of the largest particles selected at each sample station are listed accord-

Table 1—Size and lithology of largest particles.
Figures indicate number of particles.

Intermediate diameter (millimeters)	Sandstone- Quartzite	Vein Quartz	Chert	Crystalline rocks
1024				
	1			
512				
	16			
256				
	20	1		
128				
	7	4	1	1
64				
	10	15	3	1
32				
	4	17	17	
16				
		10	9	
8				
		5	3	
4				
		2	1	
2				
		2		
1				

ing to abundance in size grades. The table substantiates field observations that the largest particles are dominantly quartzose aggregates, i.e., various sandstones and conglomerates of the most durable types, whereas the intermediate and smaller sizes are dominated by vein quartz along with considerable amounts of chert. The variation of lithology with the size grade of gravel has been studied by Davis (1958) who concluded that comparisons between samples must be restricted to specified size grades. The composition of the coarse fraction of the deposits, based on samples of 100 pebbles 1 to 2 inches in diameter from each of 6 localities, is presented in table 2. Vein quartz, sandstone and quartzite, and chert, in order of decreasing abundance, are predominant in the gravel. Igneous and meta-

Table 2—Lithologic composition of gravel (in percent).

Location	Vein Quartz	Sandstone	Chert	Crystalline rocks	Shale
Dc-I	53	40	6	1	—
Dc-II	50	34	11	3	2
Hc-I	56	33	7	3	1
Jd-I	43	39	14	4	—
Mf-I	30	46	18	2	3
Qc-I	45	24	27	1	3
Average	46	36	14	2	2

morphic rocks and shales are far less common. There is a general similarity between the lithologic composition of the Columbia gravel and that of the upland gravel of Southern Maryland as reported by Schlee (1957).

Heavy Minerals

The investigation of the heavy mineral content of the sands is based upon the same 75 samples which were used to estimate the gross mineral composition.

The percentage of heavy minerals in the 62 to 500 micron range was determined by weighing the "light" and "heavy" fractions after separation. This percentage was recalculated as an estimate of the heavy mineral content of the entire sample. Heavy minerals occur in sizes other than the 62 to 500 micron range; however, the amount of silt and clay in the samples is generally small and in the sizes larger than medium sand heavy grains are very rare. The method outlined provides a minimum percentage which is probably a reasonable estimate of the actual heavy mineral content of the sediments. This average, computed for 75 samples, is 0.83 percent. Of this, an average of 77.6 percent of the grains are opaque. These data, together with the percentages of non-opaque mineral species, are listed in Appendix IV for each sample. A summary of the accessory mineralogy of the sands is given in table 3. In all cases non-opaque mineral abundances are indicated as rounded percentages of the total non-opaque fraction.

The most common non-opaque heavy minerals, listed in order of decreasing abundance, are: zircon, epidote, amphibole, sillimanite, tourmaline, and staurolite. Other minerals present, also in order of decreasing abundance, are kyanite, altered grains, garnet, chloritoid, pyroxene, andalusite, apatite, monazite, sphene, and spinel. The last four are very rare. Characteristics of the more abundant heavy minerals are listed below:

Amphibole: For convenience in this reconnaissance all amphiboles have been grouped together. Hornblende greatly predominates with small amounts of actinolite and tremolite often present in addition. The hornblende varies in both color and the intensity of color; most is green and there are lesser amounts of brown and blue-green. The amphiboles retain their characteristic cleavage-controlled prismatic shape but terminations vary from ragged to relatively rounded.

Epidote: Epidote tends to occur in irregular but roughly equidimensional grains or, much more rarely, in crude prisms. Color varies from very pale green, almost colorless, through yellow-greens to rather

intense, bright, green grains. The characteristic pleochroism is present to a degree approximately proportional to the intensity of coloration.

Sillimanite: The sillimanite examined occurs in colorless, usually prismatic grains. Less common are more or less tabular grains flattened parallel to 001 which could easily be misidentified except that they yield acute bisectrix interference figures showing the distinctive small 2V of sillimanite. The fibrous variety, fibrolite, is included here and comprises about 10 percent of the total sillimanite.

Tourmaline: The tourmaline, although usually prismatic, is found here not infrequently in "flakes" controlled by the 0001 parting. The 0001 grains show little pleochroism and give centered interference figures. The degree of rounding varies greatly. Yellow-to-brown pleochroism is by far the most common; however, pink-black is persistent in small amounts and blue-black has also been observed.

Zircon: Zircon is the most common of the heavy minerals. It varies in appearance from nearly clear euhedra to rounded, cloudy, fractured grains. The greatest number group midway between these extremes, being colorless, subangular and subrounded, and usually retaining some evidence of their original prismatic habit. Most appear to be broken rather than worn. An average of 1 to 2 percent of the zircon grains are pink; a very few are tan.

Table 3—Summary of heavy mineral content.

	Minimum Percentage	Maximum Percentage	Average Percentage	Percentage of samples in which present
% H.M. by weight 500–62 μ	0.05	6.48	1.69	
% H.M. by weight of total sample	0.04	2.19	0.83	
% opaque	23.5	96.7	77.6	100
Amphibole	0	69	13.8	97
Andalusite	0	2	0.4	35
Chloritoid	0	4	1.0	65
Epidote	3	45	17.6	100
Garnet	0	11	1.6	76
Kyanite	0	7	2.1	88
Pyroxene	0	6	0.5	33
Rutile	0	8	3.7	95
Sillimanite	1	30	12.7	100
Staurolite	0	23	4.3	99
Tourmaline	0	25	5.8	99
Zircon	3	62	33.6	100
Altered	0	7	1.9	77
Unidentified	0	4	1.5	79
Apatite, Sphene, Monazite, Spinel			Trace	

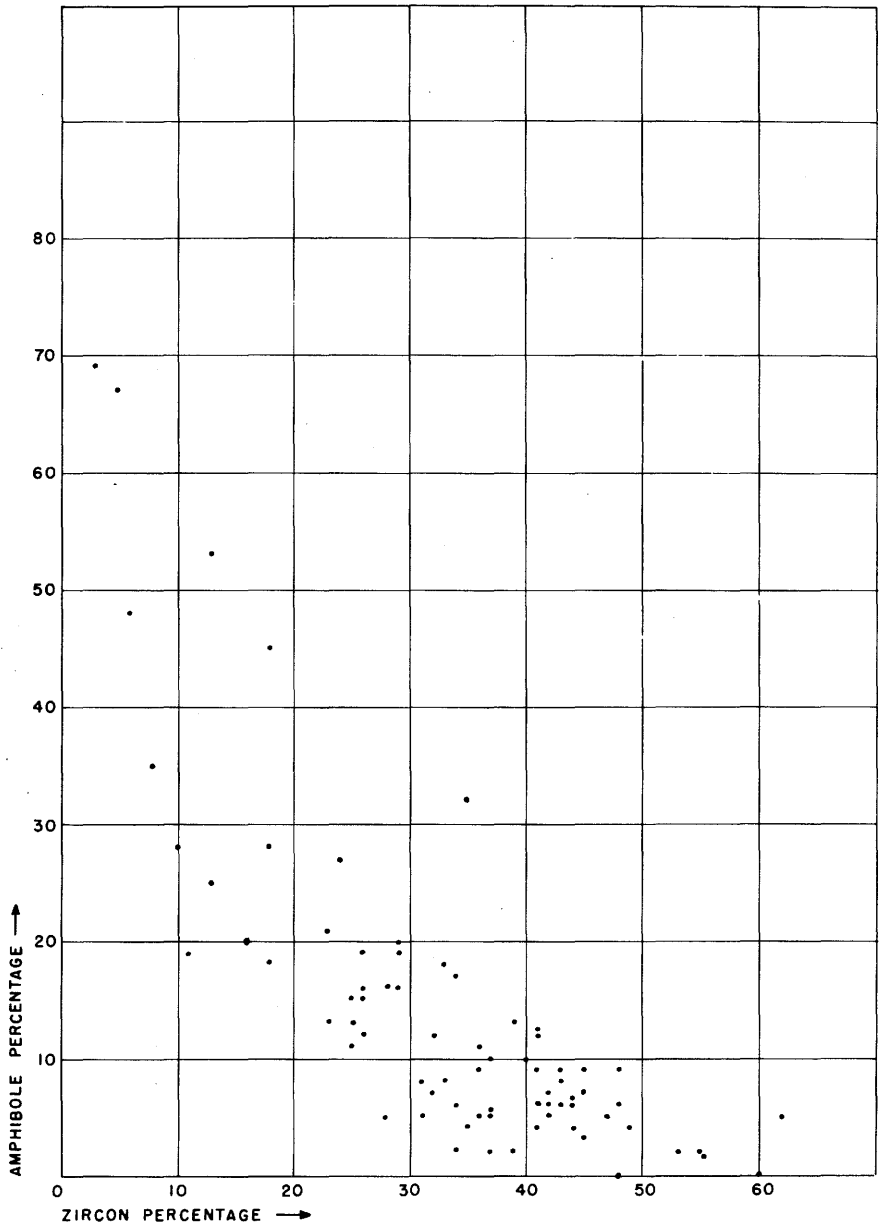


Figure 15. Relationship between relative amounts of amphibole and zircon among non-opaque heavy minerals.

Notable features of other species may be summarized: about half of the garnets are colorless, half pink; a small amount of the kyanite is rounded; the pyroxene is mostly hypersthene; rutile, although never very abundant, is remarkably persistent. Altered grains are those which have weathered to such a degree that the optical properties of the original mineral are destroyed. The "unidentified" category includes aggregates and grains whose identification is in doubt because of surface coatings, unusual orientations, or lack of distinctive features.

The heavy mineral content of the Columbia sands is rather consistent over-all, despite great differences between the individual samples representing extremes of fluctuation of the suite. The heavy minerals, as now known, may best be considered as comprising a single suite with occasional major variations. The suite is dominated and characterized by zircon, epidote, amphibole, and sillimanite. The major variations concern the amounts of zircon and amphibole contained in individual samples. The average percentages of zircon and amphibole are 34 and 14, respectively, and the extreme ranges are from 3 to 62 percent and 0 to 69 percent. A suggestion of a reciprocal relationship between amounts of zircon and amphibole may be seen in figure 15 where these values have been plotted as a scatter diagram. The most notable exceptions to the trend, and to the suite in general, are a few samples in which the amphibole content is extreme. These samples are not taken to represent a separate suite because they are widely scattered geographically (Cd-II, Kd-I, Ld-II, Nh-I, Oh-II), share no other distinctive feature, and are in all other ways intimately mixed with the other sands. The unusual concentrations of amphibole might be due to variation in supply but seems more likely to be related to selective sorting. That this might occur is suggested by the work of Bullard (1942) who found that hornblende discharged into the Gulf coastal currents by Texas rivers tends to outrun associated minerals because of its shape and density.

The heavy mineral suite described above generally resembles that indicated by 25 samples from Delaware Pleistocene sands published by Groot (1955). The present study differs mainly in the presence of more zircon and less amphibole and greater consistency of the suite.

ORIGIN OF THE COLUMBIA DEPOSITS OF DELAWARE

General Statement

The conspicuous surficial deposits of the Atlantic Coastal Plain have, in general, drawn the attention of many authors and it is fortunate that several fine summaries of the literature are available. Especially useful are the reviews of the very early studies by McGee (1888) and Shattuck (1906) and the later summaries of Flint (1940) and Hack (1955). Except for the early investigations by Booth (1841) and Chester (1884, 1885) work in the area by the present study has been largely incidental to that in adjacent areas or to investigations of primarily economic emphasis.

The continuity of the post-Miocene veneer of the Atlantic Coastal Plain implies that the history of a portion of the area will bear upon that of other localities. Workers in adjacent areas, and even within the same area, have, however, held differing views regarding the origin of these deposits. The summary of Flint (1940, p. 770) showed: "... that the hypothesis of dominantly fluvial deposition in New Jersey has been unchallenged; that serious objections to a hypothesis of chiefly marine deposition in Maryland exists, that there is evidence of both marine

and fluvial deposition in Virginia, and finally, that widespread evidence (chiefly morphologic) or marine deposition has been reported from Georgia, South Carolina, and Florida." According to the literature, the mode of origin of the sediments in Delaware would seem to be something between probably mostly fluvial as postulated for New Jersey, chiefly on the work of Salisbury and Knapp (1917), Campbell and Bascom (1933), and MacClintock and Richards (1936), and possibly marine as espoused for Maryland by Shattuck (1906) and Cooke (1930) but challenged by Campbell (1931), Flint (1940), Hack (1955), and Schlee (1957). Specifically within Delaware, the authors of the two folios which cover much of the State (Miller, 1906; Bascom and Miller, 1920) subscribe to the concept of marine deposition and terrace formation stated by Shattuck (1906). Later, Marine and Rasmussen (1955), Ward and Groot (1957), and Rasmussen *et al.* (1960) mention melt-water flooded streams and lowered sea levels as major influences during the deposition of Pleistocene sediments. Clearly, the hypotheses concerning the origin of the surficial sands of the Delaware area are varied and the matter is unresolved. It is an objective of this study to attempt to formulate a framework which will account for the observed characteristics of the sediments. Consideration will be given to the provenance, environments of deposition, sedimentary framework, and possible history of the deposits.

Provenance

Coarse fragments. Within the sediments, the larger particles provide the most readily available indication of provenance as they are pieces of the source rocks, not mere mineral grains. Not all source rocks have the same chance of preservation due to variations in block-forming ability, and resistance to chemical and mechanical attack and, therefore, it is not to be expected that all of the source rocks will be represented among the gravels. The composition of the gravels of the surficial sands of Delaware has been summarized in tables 1 and 2. It is dominated by such resistant materials as sandstone, vein quartz, and chert; shales, igneous, and metamorphic rocks are much rarer.

The sandstones, including conglomerates and quartzites, closely resemble rocks exposed in the Folded Appalachians. Medium to coarse-grained ortho-quartzite sandstone and conglomerate fragments are common in the coarser grades. Occasionally these fragments contain ripple marks or cross-bedding. They are highly suggestive of the ridge-forming quartzites of the Appalachians such as the Shawangunk, Oriskany, Pocono, and Pottsville. Some poorly preserved casts of brachiopods have been found in cobbles and boulders.

Metaquartzites found in the Piedmont (Chickies, Antietam, Setters) could have yielded the less distinctive cobbles of fine-grained welded quartz. The rather rare fragments of less pure sandstones, commonly more deeply weathered, could come from many sources in the Appalachians or the Triassic Basins.

Vein quartz, although probably the most abundant material in the coarse sediment, provides little evidence of provenance, though within the Appalachian System it is more likely to have been derived from crystalline than from sedimentary rocks. It is singularly lacking in distinctive features.

Chert, such as is present in the sediments, is found in abundance in some of the Paleozoic limestones of the Appalachians. This source is rendered more probable by occasional occurrences of crinoid stem fragments and corals in the chert pebbles.

Most of the few particles of shale and slate observed are red. The source of

these particles may lie in the continental deposits of Paleozoic or Triassic age found to the north and west.

Crystalline rock fragments are scarce, and poorly preserved. The predominant mica schists suggest derivation from the Piedmont, although the other large mass of crystalline rock nearby, the Reading Prong, may also be represented.

Heavy minerals. The general composition of the heavy mineral suite (table 3) reflects varied terrane in the source area but with a marked metamorphic rock influence shown by the relatively large amounts of sillimanite and epidote. The smaller amounts of staurolite, kyanite, and andalusite also ultimately require metamorphic sources. The obvious source of such minerals is the Piedmont Province where rocks of varied metamorphic grade occur. In general, these rocks contain the distinctive metamorphic minerals as well as hornblende and zircon (Dryden and Dryden, 1964) which are so prominent in the sands.

Certain varieties of heavy minerals may be sufficiently distinctive to allow correlation with a specific source rock. Dryden and Dryden (1964) have catalogued the occurrences of such minerals in the rocks of the northern Piedmont, Triassic Basins, Reading Prong, and Folded Appalachians. This enables a few of the heavy minerals found in the Columbia sands to be traced more or less precisely to source rocks. The occurrence of a distinctive type of tourmaline with reddish-to-black pleochroism has been recorded. This type of tourmaline is yielded by the Wissahickon Schist and has no other known source in the region. The pink zircons which constitute a few percent of all zircons in the sands come from the Baltimore Gneiss and the ancient gneisses of the Reading Prong.

Summary. Actual fragments of rock, some containing Paleozoic fossils, are present in the gravels of the Columbia of Delaware which are very probably derived from the sedimentary rocks of the Appalachian System. The evidence derived from the strongly metamorphic suite of heavy minerals indicates sources in the crystalline rocks of the Piedmont and Reading Prong. It appears that both the crystalline and sedimentary rocks lying to the north and west of Delaware served as sources of the sands. How much came from which area cannot be judged. It is not surprising that these sources are indicated by different aspects of the composition of the Columbia because some of the older sediments are durable enough to be preserved as large fragments but contain severely impoverished heavy mineral suites (Dryden and Dryden, 1964) whereas the crystalline rocks yield durable and distinctive heavy minerals (Dryden and Dryden, 1964) but weather readily as aggregates.

It is likely that the older sediments of the Coastal Plain constitute a third source of material. The evidence here, however, is less direct. The erosional contact between the Columbia and older rocks is suggestive as is the occurrence of andalusite which is found also in the Columbia sands but which does not have known source rocks among those now exposed in the Piedmont. Chloritoid has a very limited source in the low-grade facies of the Wissahickon Schist (Dryden and Dryden, 1964) and additional amounts may have been derived from some of the older rocks of the Coastal Plain in which it is plentiful (Groot, 1955). Glauconite is found in the sands immediately above contacts with Cretaceous and Eocene greensands.

Additional evidence of the nearby crystalline rocks as sources of detritus is provided by the feldspar content of the sands. Neither the sedimentary rocks of the Coastal Plain nor those of the Appalachians are feldspar-rich.

As Dryden and Dryden (1964) have pointed out, a rock cannot certainly be said to be the source of a given sedimentary fragment until all other possible sources have been examined and excluded. In application this is difficult, if not impossible, to achieve. The derivation of some of the Columbia sediment from sources other than those considered and which are now either destroyed or seemingly too far removed, remains a possibility.

Environments of Deposition

General statement. Evidence regarding the environment of deposition of a sediment is commonly sought in its organic content and physical properties. In the Columbia sediments the interpretation relies heavily on physical characteristics because of the scarcity of fossils. Characteristics such as morphology, structures, textures, and impurities may identify environments of deposition with varying degrees of certainty. Because this study deals with a large mass of sediment spread over a considerable area and is based on observations made at a limited number of surficial exposures, delineation of environments of deposition in great detail has not been achieved. However, sufficient information is at hand to decipher the gross depositional system, at least for the materials exposed at the surface.

The interpretation presented here recognizes three major environmental facies: fluvial, shoreline complex, and, possibly, bay. Not all of the sections visited can be fitted unequivocally into one of these three categories. The over-all depositional scheme calls for a major stream system discharging into the sea with the shoreline interface between the two changing position with time because of changes in relative sea level. The areas occupied by materials representing the various environments of deposition are indicated on plate 9. It should be noted that the facies discussed are not necessarily limited to the areas indicated; remnants of one type may be found in an area dominated by another or erosion of a surficial unit may reveal a different older facies below.

Fluvial facies. The fluvial origin of the deposits found over much of the northern part of Delaware has been accepted by most modern authors. Certainly these deposits may be more confidently relegated to a specific environment of deposition than the others dealt with here. A division might be made, as has been on plate 9, between a major channel facies and a minor channel or perhaps inter-channel, facies. The features of the former are present in the latter but at reduced scale.

The variable thickness of the Columbia Formation over much of the northern two-thirds of Delaware has been noted. This is a reflection of the occurrence of the sediments as fillings of former stream valleys, hence the "channel" deposits. Several of these channels may actually be observed in the banks of the Chesapeake and Delaware Canal. The occurrence of some channels there has been indicated by Groot, Organist, and Richards (1954). At the Canal the channels range up to more than 0.5 miles in apparent width and have been cut into various Cretaceous formations, commonly to depths below sea level, from general land surface elevations of 40 to 70 feet. Exactly how many channels cross the Chesapeake and Delaware Canal is not clear because of destruction of much of the exposure by engineering projects; however, probably about 6 channels cross the Canal between Chesapeake City, Maryland and Reedy Point, Delaware, a distance of about 13 miles in a line roughly normal to the paleocurrent direction. Channels, or at least restricted areas of unusually thick surficial sands, have been located by drilling at many localities north of the Chesapeake and Delaware Canal and at isolated

sites, such as near Smyrna and Milford, far to the south. Rasmussen *et al.* (1957) have described the channels north of the Canal and provide a structure contour map on the base of the "Pleistocene." An interpretation of these channels as valleys cut at four different elevations during glacial stages and filled in interglacials is also offered. The form of the valley system indicated by the morphology strongly suggests erosion by streams. The detritus filling the channels is also interpreted as being of fluvial origin based on its sedimentary structures and textures.

A striking feature of the sediment of the fluvial facies is its distinct bedding. The degree of sorting may vary widely over-all but individual beds large and small are discrete units which tend to retain their distinctive textural characteristics, often through the entire length of an exposure. Pebbles are segregated into beds of gravel and thin, but persistent, beds of silt may be present. This reflects the rapidly changing current regimen common to many streams. Cross-bedding is generally well developed and of the tabular type, and some individual cross-beds persist over distances of hundreds of feet. The magnitudes of the vector means of dip azimuths tend to be large at individual outcrops, indicating strong unidirectional current trends. Cut and fill structure is sometimes present and a few slump features were observed.

The coarseness of the sediments of this facies varies considerably from bed to bed and, to a lesser degree from outcrop to outcrop. It does, however, show a somewhat systematic decrease in the downcurrent direction (plate 4; figure 7). The sorting is also variable but possibly greater contrasts exist between than within outcrops and it does generally improve downcurrent (plate 5; figure 8) mostly due to attrition within the larger size grades. Krumbein and Sloss (1951, p. 200) have commented on the abruptness of textural changes between beds within a systematically changing framework as an attribute of alluvial sediments. The grain size distributions are usually bimodal, a characteristic which Schlee (1957) found to be indicative of the fluvial environment. The presence of bimodal distributions with the secondary mode representing a coarse admixture is reflected in the predominantly negative skewness values for these materials (plate 6). Conspicuously large fragments are often present. Some of these are of the angular type attributed to ice-rafting. Finally, of all the deposits examined, these show the strongest coloration, usually tan, brown, or reddish-brown. This and the common occurrence of limonite ledges suggests an oxidizing environment, probably subaerial.

The only evidence from fossils which may be pertinent to the problem of the origin of this facies is the report by Hyyppä (*in* Flint, 1940) of Recent freshwater diatoms from poorly sorted gravel near New Castle, Delaware.

The points offered as indications of the fluvial emplacement of this facies are not all of equal weight, however, the total evidence, plus the lack of contradictory evidence, is thought to be very strong.

Shoreline complex. The environments of a shoreline complex are those associated with, but not limited to, the beach and include the lagoonal, dune, and inner sublittoral as well as the beach proper. This is recognized as a transitional environment, some parts of which retain characteristics of both continental and marine conditions thereby making the differentiation of some beach and stream sediments especially difficult. Lagoonal, dune, and wave-deposited sediments are present in southern Delaware; the association of these entities strengthens the interpretation of the shoreline complex beyond the evidence provided by any one element.

Over much of Sussex County south of a west-southwest axis running roughly from Lewes through Seaford, some of the sands suggest reworking and final deposition by waves. Actual beaches may be present near that axis in both eastern and western Delaware. The beaches are not related to distinct topographic features, except, perhaps, for the presence of hummocky topography probably more expressive of beach-associated dunes. The sand is medium- or coarse-grained, well sorted, and often negatively skewed but not bimodal. These textural properties are indicative of a high-energy level which is fairly consistent and stays above a minimum threshold. The unimodal nature of the sands indicates that the negative skewness is probably a result of selective removal of fine grains as opposed to the admixture of coarse detritus. The former situation is thought to be a characteristic tendency of beaches as explained by Mason and Folk (1958), Friedman (1961), and Moss (1963); the addition of coarse material is more common in streams due to the mixing of bed and suspended loads (Schlee, 1957). The bedding of these sands is well developed, although not as prominent as in the fluvial facies for the sorting is better and the degree of textural contrast is reduced. Beds are also thinner (usually less than 1 foot), less continuous, and more complex. Cross-bedding is common but is smaller in scale and less continuous than that found in the fluvial sands. Small-scale cut-and-fill structures and abrupt thinning and thickening of individual beds are often present. Close examination reveals laminae within the main beds which appear to be similar to those described by Thompson (1937) and McKee (1957) as prominent features of beach deposits. The sedimentary structures found in this facies of the Columbia deposits resemble some of those illustrated by Thompson (1937) or his description of the features of the difficult-to-observe foreshore. Broken and worn shell fragments have been found at a few localities but never in place. The fragments are unidentifiable and highly suspect because of the long standing practice of fertilizing fields in the area with shells.

A secondary feature observed at many localities where "clean" sands are found is a tan coloration, with thin brown bands, of the upper four to eight feet of the otherwise light gray to white sand. At the top of such exposures is found a soil horizon a foot or so thick to which the brown bands are texturally similar as they are marked by interstitial concentrations of silt and clay. The disposition of the thin brown bands is controlled by the main bedding. The transition from the tan and brown zone to the unweathered sand beneath is sharp. These features are thought to be a normal and possibly characteristic form of soil development. The feldspar content of the otherwise clay-free sands is probably sufficient to provide, upon leaching, the necessary fines which are concentrated between sand grains at places where the greatest changes in permeability occur.

Other sands occur in Sussex County which are sorted as well or better than those attributed to the beach environment. These sands are sometimes associated with hummocky topography suggestive of stabilized dune-fields. Structures are somewhat vague because of a rather low textural contrast between beds. Soil development is usually of the type described above and serves to emphasize the main bedding. Although detailed structure is present it is very difficult to discern unless the surface is perfectly fresh and fresh surfaces are difficult to obtain because the well-sorted sand will not hold a steep face. Samples Pg-I, Rc-I, Re-I, Rf-I, and Rf-II are representative of the facies. All are very well sorted and positively skewed. The positive skewness results more from the omission of coarse material than the admixture of fines; the largest particles at these localities are relatively small. Excellent sorting and positive skewness have been found to be

attributes of dune sands (Mason and Folk, 1957; Friedman, 1961). The best example of this facies is not one of the systematic samples but is in the Pc quadrangle 2 miles southeast of Blades, near Seaford, Delaware, south of the Nanticoke River. A low rolling ridge parallels the Nanticoke for several miles and in a pit near its center is exposed a sand with $Md\phi = 1.71$, $\sigma\phi = 0.565$, $\alpha\phi = 0.0442$. For comparison with this and the other mechanical analyses 3 samples of the active dunes and 3 of the beach between Lewes and Fenwick Island, Delaware were taken. The statistical parameters derived from these samples agree with the findings of, for example, Friedman (1961) in that the coarseness and sorting of the beach and dune sands may be similar, however, beyond a certain small overlap the dunes are positively skewed and the beaches negatively skewed. Some of the postulated dune sands are associated directly with a shoreline development as is the one near Seaford, others may be wave-reworked dunes or dunes formed from the sands exposed after the retreat of the sea.

The lagoonal facies is the only one for which evidence is supplied mostly by fossils. Localities starred on plate 9 indicate the known fossil localities. These are in addition to the dredged material from eastern Sussex County described by Richards (1936). The fauna at Pepper Creek Ditch and U. S. Route 113 between Dagsboro and Frankford includes:

Mollusca

Crassostrea virginica (Gmelin)
Odostomia (Chrysallida) dianthophila H. W. Wells
O. (Chrysallida) sp.
O. (Menestho) impressa Say?
O. (M.) trifida Totten?
O. (M.) sp.
Cerithiopsis greeni C. B. Adams
Cumingia tellinoides Conrad
Noetia ponderosa (Say)

Foraminifera

Elphidium clavatum Cushman
E. florentinae Shupack
Rotalia beccarii tepeda Cushman

indicating shallow, possibly brackish water conditions. The faunas of the other localities, although not so well preserved, appear to be similar. The fossiliferous sediment in each case is dark silt or silty, very fine sand and is found in beds one to two feet thick which are interspersed among coarser, better-sorted sand beds. In the subsurface of southeastern Sussex County the alternating sands and silts and silty sands have been described as the Omar Formation (Jordan, 1962). Unfossiliferous Omar may also be seen at the surface at localities Qh-I and Qh-II. The shell beds seem to be of rather limited extent and most of the unit is unfossiliferous, except that diatoms have been found in cores of the Omar.

Many exposures in association with the facies already described as belonging to the shoreline complex in southern Sussex County reveal light tan or gray, fine-, medium-, or coarse-grained, unimodal, moderately well-sorted, positively skewed sands. As a rule bedding is indistinct and cross-bedding is not visible. It is possible that more structure and lithological variation would be apparent if the exposures were better, but they are poor and the sands appear rather homogenous. Potter and Pettijohn (1963, p. 63) noted that cross-bedding is more common in fluvial and eolian than littoral or marine sands. Elongate ridges are present within the area of the shoreline complex as indicated on plate 9 and most are associated

with the rather nondescript sands mentioned. The ridges are from 0.5 to 3 miles long and rise 10 to 20 feet above the surrounding areas. They may be narrow or broad and some are slightly sinuous but are not to be confused with the circular or elliptical rims of "bays and basins." Some of the longer and broader of the ridges may be related to beach ridges, however, the majority are discrete elements widely scattered in position and elevation. These features are considered to be wave and current generated bars. This assignment is tentative pending investigation of their internal characteristics and relationships to adjacent materials. MacClintock (1943) has described similar features in the vicinity of Cape May, New Jersey and interprets them as bars built near the shoreline. The general northeast-southwest trend of the New Jersey bars is in agreement with trend of the features found in Sussex County.

The interpretation of the sands of southern Delaware as wave-washed and possibly of sublittoral origin, or perhaps related to the retreat of the sea is based on the circumstantial evidence of association with more clearly defined facies and the bar-like features.

Estuarine (?) facies. The final distinctive type of lithology noted is best developed in southwestern Kent County and northwestern Sussex County in the general area north of Greenwood, Delaware. Medium-grained and occasionally fine and coarse sands are found there which are distinguished by their irregular and indistinct bedding and abrupt lateral and vertical color changes. Mottling is common and applies in some cases both to lithology and the tan, brown, yellow, and gray colors. Some of the mottling resembles the work of bottom-dwelling organisms; however, no fossils have been found. The origin of these materials is a matter of conjecture and it is only on the resemblance of the mottling and poor bedding to similar features in some of the Recent sediments of estuaries off the Delaware Bay that the estuarine origin is tentatively suggested.

Synthesis

The sedimentary framework in which the Columbia deposits of Delaware were formed may be described as a continental shelf system in which the detritus was derived from the seaward-facing slopes of the continent and transported to and, in part, across the coastal plain into the area of influence of the sea. There are, of course, ordinarily many variables in such a system and the effects of the multiple Pleistocene glaciation may be expected to have had additional profound effects upon it. Some of the previous work has stressed the cyclical nature of erosion and deposition, of great and small stream discharge, and of high and low sea level based on the assumption that each advance or withdrawal of the glaciers would produce similar effects in the area of deposition. Furthermore, some have relied upon each successive interglacial stand of sea level being lower than that preceding as is required to preserve terraces from destruction by the rising seas. Changes in sea level are considered to be eustatic and the Coastal Plain tectonically very stable. These concepts lead to the expectation of very widespread units representing, in alternation, continental and marine deposition or erosion and deposition and arranged one above the other at different elevations or, in places, superposed upon each other.

If several cycles of sedimentation are represented by the Columbia sediments of Delaware they are not obvious in the properties studied. The distribution of cross-bedding directions has a single mode and the moving average (plate 3) shows but a single large anomaly which is located in west-central Kent County.

The downstream decreases in cross-bed thickness, median grain size, maximum particle size, maximum grade size, and increase in sorting do not seem to be interrupted by the effects of any other depositional cycle. The Coastal Plain of Delaware reaches a maximum altitude of about 100 feet near the Fall Zone and nearly 80 feet in south-central Sussex County, and therefore might be expected to show several of the lower terrace features. If they are present in anything like the manner in which they have been mapped by Shattuck (1906) they are not apparent to the writer as either physiographic features or as sedimentary entities. Fluvial features may be found in the northern two-thirds of the State at all elevations and the bottoms of the channels may be as much as 100 feet or more below present sea level (based on Rasmussen *et al.*, 1957, and other well records) whereas features interpreted as marine may occur at elevations as great as 50 to 60 feet in the south. Neither grain size nor mineralogy appear to be primarily controlled by elevation. In figure 16 the median grain sizes of samples have been plotted against the altitude of the top of the sampled section. No systematic relationship between grain size and altitude appears to exist.

The oldest materials should be the most highly weathered. This might be indicated by changes in the heavy mineral suite such as Sindowski (1949) found in the Pleistocene terraces of the Rhine and Neiheisel (1962) found in Pleistocene terraces of the Georgia Coastal Plain where the older (higher) terraces are depleted in the less stable heavy minerals. The proportion of a heavy mineral suite which consists of the most resistant minerals may be regarded as a measure of the relative amount of weathering suffered. The ZTR maturity index (Hubert, 1958, *in* Hooper, 1961) which is the percentage of the non-opaque heavy minerals represented by zircon, tourmaline, and rutile together is such a measure. Here the ZTR index has been calculated as the ratio of the combined percentages of zircon, tourmaline, and rutile to all other non-opaque minerals. This figure is recorded as a "stable mineral index" in Appendix V. Fluctuations, which largely reflect changes in the relative amounts of zircon and amphibole, are considerable but cannot be correlated with other features of the samples. In figure 17 the "stable mineral index" has been plotted against the altitudes of the tops of the respective channel samples. Also in figure 17, the percentages of feldspar in the same samples have been plotted against altitude. The generally greater durability of quartz than feldspar in the sedimentary environment suggests that the proportions of these minerals in the sands might be controlled, in part, by the ages of the deposits. No relationship is present between the feldspar contents and the heavy mineral suites of the samples from the Columbia. As shown in figure 17, the vertical distributions of both heavy minerals and feldspars lack marked trends. It also seems that, on the basis of the present sampling, no pronounced trends in the geographic distribution of these compositional elements are present. This is in agreement with statements of Shattuck (1906), Miller (1906), and Bascom and Miller (1920) that the "terraces" are not lithologically identifiable.

The Columbia in Delaware appears to be, in terms of its dispersion and lithology, essentially a continuum. If true, this indicates that evidence of older cycles of deposition has been destroyed by succeeding cycles, that the cycles are so similar that the superposition of one on another is not evident, or that only a single cycle is present, perhaps due to the distributary system shifting laterally from an area occupied by the sedimentary products of a previous cycle. The choice between, or perhaps, the analysis of the combination of, these possibilities is rendered difficult in the present investigation by the lack of stratigraphic control within the sands.

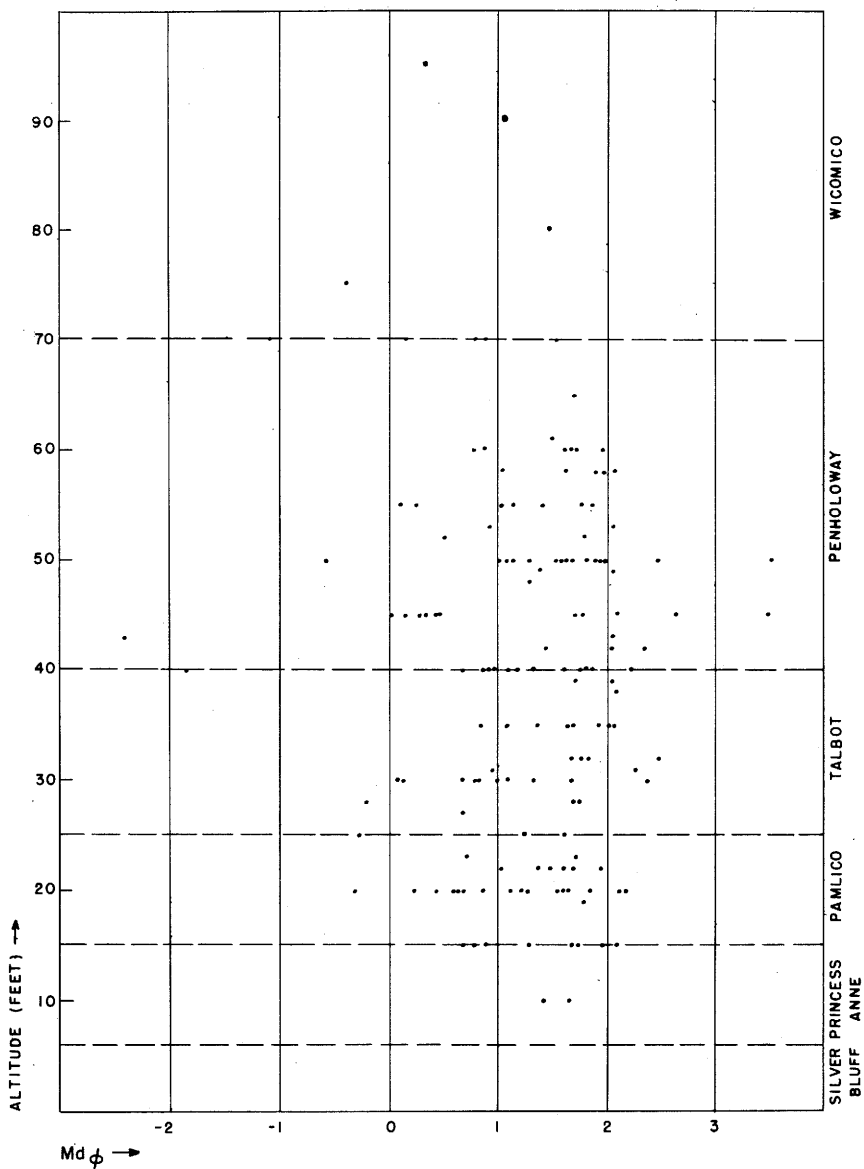


Figure 16. Relationship of median grain sizes to altitudes of tops of channel samples. Terrace designations from Cooke (1930).

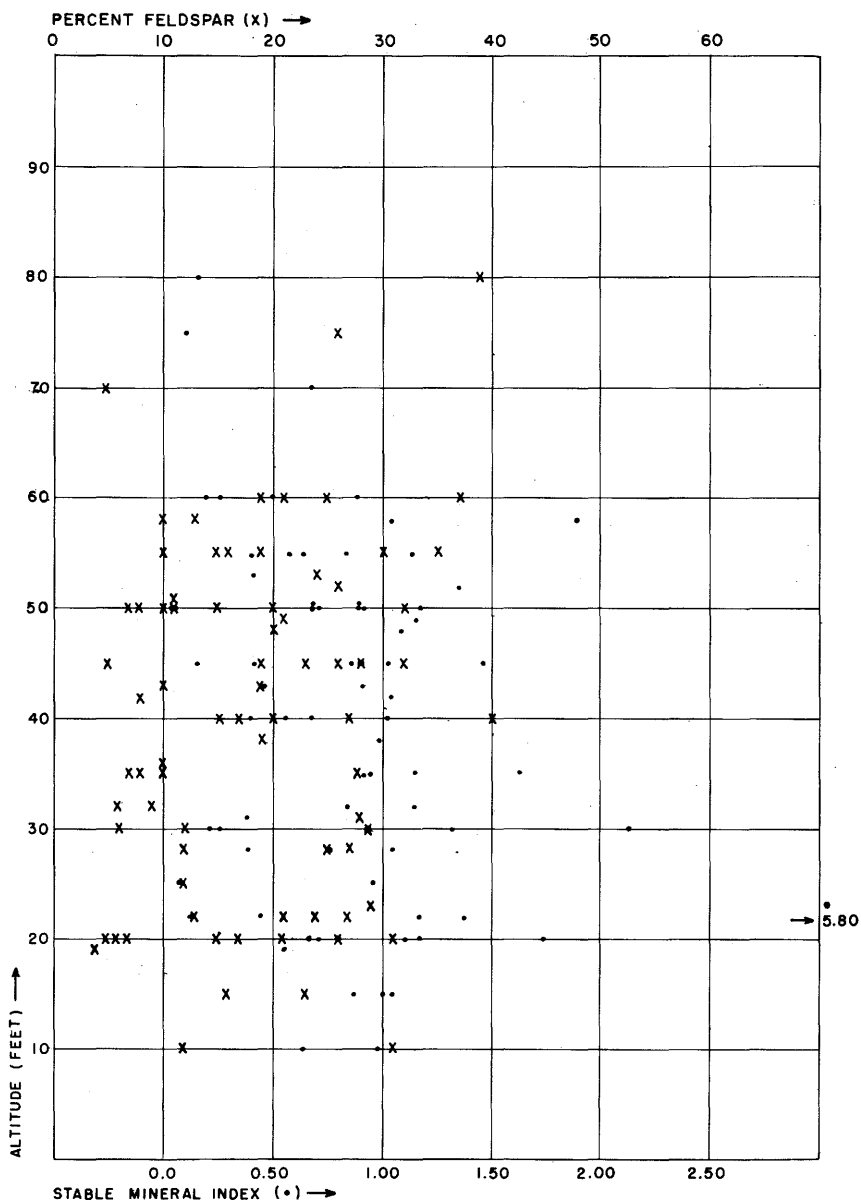


Figure 17. Relationships of "stable mineral index" (•) and feldspar content (x) of channel samples.

It seems inescapable that during Columbia time both deposition and erosion occurred on the Coastal Plain and that the net effect was the deposition of a large mass of coarse clastic sediment. It is desirable but difficult to relate the chronology of the Coastal Plain events to that of the glaciated area; recent interpretations of these relationships for the Eastern Shore of Maryland are provided by Rasmussen and Slaughter (1955, 1957). If a marine environment of deposition is postulated for the Columbia, sea level must have been high and the glacial ice at a minimum; on the other hand, if streams flowed below present sea level in the process of depositing the Columbia, then sea level was lower still and the ice far advanced. Chronology is severely hampered by the lack of fossils and even the rare occurrences of fossils provide only indirect evidence of age through environment and climate: a warm interglacial might be indicated but not which one of the interglacials. Materials suitable for radiometric dating are as scarce as fossils but offer more hope for precision in correlation.

Some of the events of Columbia time in Delaware may be detected and sequentially arranged on the basis of physical properties and interrelationships. At its simplest the system of deposition calls for a stream system cut to below sea level and then filled with coarse fluvial detritus; the rising sea then reworked the distal portion of the fluvial deposits to relative elevations above present sea level. It is possible that this was repeated several times; however, it is suggested that only one episode is well preserved and so it alone will be discussed.

Erosion of the channels of the Coastal Plain cannot be strictly relegated to glacial periods and deposition reserved for interglacials; nor can the reverse situation be adhered to rigorously. If transportation and deposition of the coarse detritus was accomplished by glacial meltwater floods then the channels would have had to be cut below sea level in the previous interglacial, during a presumed high stand of sea level. The alternative is equally untenable because if erosion occurred when the ice advance was maximal and sea level low, and then deposition occurred during an interglacial when conditions approximated those of today, the channel fillings would be expected to resemble the present fine-grained detritus of the Delaware rather than the coarse materials actually present.

The first event recorded is the cutting of the stream channel system found in New Castle and Kent Counties. This process may have flushed older deposits from the valleys. At the time of deepest channel cutting sea level was lower than at present because the bottoms of the valleys are now well below sea level. The lowered sea level must have persisted through at least the time necessary to fill the portions of the channels now below sea level with fluvial detritus. During this phase the channel system of the Delaware Estuary and Bay could not have been developed as it is at present or the streams would have flowed in it rather than on what is now a divide area. Within this framework it is apparent that the regimen of the streams was quite different from that of the present Delaware River. The older streams carried large amounts of coarse detritus; the Delaware and its major downstream tributaries today carry mostly silt and clay (U. S. Geological Survey, 1960; Jordan and Groot, 1962). In order to transport vast amounts of coarse detritus (the sedimentary mass is estimated at nearly 40 cubic miles in Delaware alone and the average size is medium sand) the streams' gradients, or volumes, or both must have been greater than at present. The area which primarily influenced the streams is removed from the study area but some idea of its nature may be derived from the examination of the products of the streams. The position of sea level, although it provides essentially the lower limit for the activity of a stream system, in this instance cannot alone account for major changes in stream gradi-

ents for within the bounds of the continental shelf the slopes of the emergent and submerged portions of the surface are essentially the same. It has been noted that both channel cutting and filling took place at times of lowered sea level. Factors which could have had major effects on the source areas and their drainage include tectonic activity, climatic changes, meltwater additions, and stream capture. That uplift of a source area can result in coarser detritus reaching the basin of deposition is an established principle; however, the writer knows of no independent evidence of uplift of the Appalachian system during this time. Isostatic adjustment to the weight of the ice may have affected the stream gradients. If a hinge line was present between the ice front and the area of deposition on the Coastal Plain stream gradients should have decreased. Certainly major climatic changes occurred between interglacial and glacial times but precisely how this affected stream flow is not clear; perhaps the major influence would be the promotion of generation and preservation of clastic detritus in glacial time. The most obvious source of a large increase in stream volume is meltwater which would be available as long as the ice was in the drainage basin, during the advance or retreat. Major changes in the drainage pattern of the source area such as proposed by Campbell and Bascom (1933) would also influence the regimen downstream. It is possible that each of these factors exerted some influence on the deposits in Delaware.

The features in southern Delaware which reflect marine influence are thought to be generally younger than the stream deposits to the north. The higher features of the shoreline complex could not have been contemporaneous with streams flowing below present sea level and do not seem to be older because they would block the access of the streams to the sea. Actual superposition in a single exposure has not been observed but shoreline features are found at higher elevations than nearby fluvial deposits in the transitional zone and the fluvial trends to the southeast are truncated by the east-northeast strike of the zone of shoreline development discussed above. The transgression resulted in the burial or reworking of the fluvial facies south of this zone. A higher stand of sea level is required by the shell beds found in southern Sussex County at elevations from about 10 to 20 feet. The faunas indicate shallow water and, depending on the relief of the bottom and the rate of change of sea level, might be essentially contemporaneous with the formation of the bar-like features and the transitional zone of the shoreline along the Nanticoke River. An oyster shell from Pepper Creek Ditch near Frankford, Sussex County has been dated by radio-carbon at 34000 ± 2000 years. As this is generally agreed to be about the middle of the Wisconsin Glaciation it would be a remarkable time for relative sea level to be at least 20 feet higher in southern Delaware than it is at present. Wood from altitude -2 feet from the type well of the Omar Formation near Omar in southeastern Sussex County has been dated at approximately 32000 years. The Omar is interpreted as a product of transgression and regression containing lagoonal deposits. The dates may be spurious and misleading; however, if they are valid, it may be necessary to turn to movement of the land to account for the position of sea level at the indicated times. Possible instability of the Coastal Plain must be considered, although the data of the present study shed no additional light on the problem. The continental shelf may be considered generally unstable in the sense that there must be a net subsidence to account for the accumulation of the great volume of sediment of which it is composed. Major transgressions and unconformities are recorded in the older rocks of the Coastal Plain and a major basement structure, the Salisbury Embayment (Richards, 1948) extends into southern Delaware and is reflected in the older Coastal Plain sediments (Jordan, 1963). MacClintock and Richards

(1936) felt that the possibility of diastrophic movement should not be ignored. Hack (1955, p. 39) found that in southern Maryland "... there appears to be no reliable evidence ..." relating to crustal stability but argues that deformation has not been disproven and must be recognized as a possibility. The case for Delaware is similar.

Any history of the Columbia deposits of Delaware must include explanations of (1) channel cutting and filling below present sea level; (2) the transportation and deposition of a large volume of coarse sediment by streams; (3) a higher stand of sea level, the effects of which seem to be restricted to the southern part of the area by some mechanism other than altitude alone; and (4) the incision of the Delaware Bay channel system to below present sea level. All or some of these events may have occurred more than once. To the best of our knowledge they happened during Pleistocene time and certainly the complex history of the Pleistocene provides ideal opportunities for a geologically rapid sequence of major events. Relative sea level, climate, stream volume, stream gradient, availability of sediment, distance of transportation, and possibly other factors all varied, and magnitudes, rates, and times are imperfectly known. The presence of many poorly defined variables is a disadvantage in that their complex interrelationships are generally imponderable; however, it provides one sufficiently imprudent with many possible mechanisms by which to explain the observed phenomena.

The geologic events on the Coastal Plain recorded by the fluvial Columbia sediments are thought to correlate with glacial, as opposed to interglacial, time if the former is considered to include the transition from interglacial conditions to maximum glacial development and back to interglacial. It is during this glacial phase of the cycle that sea level would be expected to be relatively low, the detritus most available, and the streams at maximum volume. Negative evidence is provided by the situation observed today in which what are commonly assumed to be essentially interglacial conditions do not produce fluvial sedimentation of the type displayed by the Columbia Formation. The writer's hypothesis assumes the existence of an interglacial situation similar to that presently existing. A set of channels, cut below sea level, is present as an inheritance from some previous phase although it may be more or less filled by fine-grained sediment. With the onset of glaciation sea level falls and the ancestral stream adjusts but does not necessarily cut deep channels because the surface of the shelf has a low gradient. As glacial ice advances into the drainage basin of the stream system, an increment in stream flow is derived from the meltwater and it is possibly the resulting increase of stream competency which flushes and extends the channels on the Coastal Plain. The initial phases of retreat of the ice would bring an additional increment in stream volume from meltwater and also the debris derived from the glacier would become available for transportation. At this time sediment-choked streams of high volume, perhaps aided by increasing gradients from the crustal adjustment to unloading as a result of glacial thinning and retreat, could transport the glacial debris and rework outwash which the earlier stream regimen could not move. As these streams debouch on the Coastal Plain the reduction in gradient, and therefore competency, triggers deposition and filling of the channels. After this initial clearing of upstream channels and filling of those downstream, but before the ice retreats from the drainage basin or sea level regains its high stand, the downstream (Coastal Plain) deposits are fluvially reworked and channeled by readjusting streams which must maintain grade to the sea. These channels are inherited by the succeeding interglacial during which marine reworking and deposition occur above the benchmark of present sea level. During such a phase

the features in southern Delaware attributed to the marine environment were formed. The fluvial channels need not be filled during this phase even as the Delaware Bay channels are only partly filled with fine-grained sediment today except at the mouth of the Bay where wave-transported sand has built a threshold.

If all this activity is attributed to the most recent cycle of glaciation and deglaciation (and this is not implicit in the data) it may be necessary to appeal to some crustal movement to accomplish the return to the present sea level.

Ward (1938) unhesitatingly attributed the coarse-grained terrace deposits of the Delaware River between Delaware Water Gap and a point below Easton, Pennsylvania to conditions which existed when the glacial ice occupied a portion of that reach at a time approximating maximum ice advance. The terraces of the Susquehanna River are described by Peltier (1949); as the Susquehanna is similar to the Delaware in that only the headwaters were invaded by ice and that downstream it becomes a Coastal Plain estuary, some of his findings are probably equally applicable to the latter stream. Peltier (1949, p. 4-5) describes the situation upstream:

During each glacial stage large quantities of debris of all sizes were carried by the streams which flowed from the ice edge. The existing terraces are graded not to the most advanced position of each ice sheet, but to positions of the retreat. Debris was deposited during each ice advance, but it cannot be distinguished from that carried by the meltwaters of the retreating ice. The debris of the advance must have been either relatively small in volume, redeposited, or else buried by the outwash laid down during the glacial recession. A sufficient quantity of outwash was carried downstream to choke the valley for 150 to 200 miles from the ice front.

Peltier (p. 139) concludes that: "The stream channels were washed free of these deposits early in the interglacial or interstadial periods and the remnants of the preceding alluviation were left as terraces along the valley walls." If Peltier's "early . . . interglacial" may be construed to mean an early stage in the recession of the glacier, it is at this point that the fluvial deposits of Delaware correlate with the history of the glaciated area. Ewing *et al.* (1963) attribute the now submerged "apron" of the Hudson River to subaerial deposition (i.e., presumably glacial time with the ice front advanced and sea level lowered). It may be assumed that this feature is now being impressed with marine features and it may therefore be analogous to the Columbia deposits of southern Delaware.

Much of the foregoing discussion of possible history of deposition of the Columbia is highly speculative but it provides a working hypothesis based on an investigation of the sediment itself. It also serves to focus attention on deficiencies in our knowledge of these deposits. More detailed description over a wider area is desirable. Dated positions of shorelines are necessary and some means of correlation of channels and channel fill with shorelines on one hand and river terraces and moraines on the other must be found. The problems are challenging and it is to be hoped that a response may be forthcoming worthy of the work of McGee, Shattuck, Salisbury, Cooke, Flint, and the others who have provided the basis for our continuing efforts.

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APPENDIX I

Location of Samples

Approximate locations are given in reference to landmarks which may be found on 7½ minute quadrangle topographic maps and to highways and roads as numbered on the Delaware State Highway Department (D.S.H.D.) maintenance map, 1961 edition.

- Cb-I. Abandoned pit east of Rt. 356, 0.3 miles north of Rt. 273 at Ogletown, Del.
- Cb-II. Railroad cut north of Baltimore and Ohio Railroad, 0.1 mile east of Rt. 355, near Harmony, Del.
- Cc-I. Road cut south of Delaware Turnpike, 0.1 mile west of Rt. 339.
- Cc-II. Greggo and Ferrara pit north of Rt. 273, 0.2 miles west of U. S. Rt. 13 and 40 at Hares Corner, Del.
- Cd-I. Gravel pit 0.4 miles west of U. S. Rt. 13 at Minquadale, Del.
- Cd-II. Abandoned pit 0.2 miles east of U. S. Rt. 13 and 40 at Delaware Memorial Bridge approach.
- Da-I. Drainage ditch 0.1 mile north of U. S. Rt. 40, 0.7 miles west of Delaware-Maryland line at Thompson Estates, Maryland.
- Da-II. Abandoned pit 0.6 miles north of U. S. Rt. 40, and 0.7 miles west of Rt. 896, near Glasgow, Del.
- Db-I. Abandoned pit north of Delaware Turnpike, 0.2 miles west of Rt. 356.
- Db-II. Whittington's Sand and Gravel Co., 0.4 miles east of Rt. 346, 0.7 miles north of U. S. Rt. 40.
- Dc-I. Whittington's Sand and Gravel Co., south of Rt. 71, 0.2 miles east of Rt. 7 at Red Lion, Del.
- Dc-II. Wilson Contracting Co. pit, 0.1 mile east of U. S. Rt. 13 and 40 at Hares Corner, Del.
- Ea-I. South bank of Chesapeake and Delaware Canal, 0.5 miles west of Bethel, Md.
- Ea-II. South bank of Chesapeake and Delaware Canal, 0.8 miles east of Bethel, Md.
- Eb-I. Kirkwood Sand and Gravel Co. pit, east of Rt. 71, 0.4 miles north of Pennsylvania Railroad near Kirkwood, Del.
- Eb-II. Small pit west of Rt. 413 at Scott Run.
- Ec-I. D.S.H.D. pit west of U. S. Rt. 13, 0.1 mile south of Scott Run.
- Ec-II. South bank of Chesapeake and Delaware Canal, 0.6 miles east of St. Georges, Del.
- Fb-I. D.S.H.D. pit 0.1 mile east of Rt. 896 south of Deep Creek near Middletown, Del.
- Fb-II. Pit east of Rt. 896 north of Deep Creek near Middletown, Del.
- Fc-I. Abandoned pit west of U. S. Rt. 13, 0.3 miles north of Drawyer Creek.
- Fc-II. Farm pit 0.2 miles east of U. S. Rt. 13, 0.4 miles north of Drawyer Creek.
- Gb-I. Abandoned pit west of Rt. 446, 0.3 miles north of Rt. 25 near Townsend, Del.
- Gb-II. Ditch north side of Rt. 463, 0.2 miles west of Pennsylvania Railroad near Forest, Del.
- Gc-I. Pit south of Rt. 456 at Beaver Branch.
- Gc-II. Abandoned pit 0.2 miles south of intersection of Rts. 51 and 465.
- Hb-I. Abandoned pit east of Rt. 483, 0.6 miles north of Rt. 40.

- Hb-II. Abandoned pit east of Rt. 131 at Blackiston Church, Del.
- Hc-I. D.S.H.D. pit north of Rt. 134, 0.8 miles north of Rt. 6 at Clayton, Del.
- Hc-II. Abandoned pit south of Rt. 487, 0.3 miles west of U. S. Rt. 13.
- Hd-I. Roadcut east side of Rt. 325, 0.5 miles north of Rt. 12.
- Hd-II. Abandoned pit 0.2 miles north of Rt. 82, 0.3 miles east of Rt. 317.
- Ib-I. Pit 0.3 miles north of Rt. 300 at its intersection with Rt. 44.
- Ib-II. Pit 0.3 miles east of Rt. 11, 0.9 miles south of Rt. 300.
- Ic-I. Pit west of Rt. 91 north of Leipsic River.
- Ic-II. Pit 0.3 miles south of Rt. 149, 0.4 miles west of U. S. Rt. 13.
- Id-I. Dover, Del., dump 0.5 miles west of U. S. Rt. 13, 0.5 miles north of State College.
- Id-II. Abandoned pit 0.6 miles north of Rt. 331, 2.0 miles east of U. S. Rt. 13.
- Jb-I. Abandoned pit just west of Delaware-Maryland line 1.3 miles south of Maryland.
- Jb-II. Abandoned pit 0.2 miles northeast of Rt. 221 at Tappahanna Ditch.
- Jc-I. Abandoned pit 0.25 miles north of Rt. 52 at Rt. 230.
- Jc-II. Abandoned pit 0.4 miles east of intersection of Rts. 232 and 227.
- Jd-I. Pit 0.8 miles north of Rt. 365, 0.7 miles east of U. S. Rt. 113A.
- Jd-II. Pit 0.2 miles east of U. S. Rt. 13, 0.1 mile south of Isaac Branch.
- Je-I. St. Jones River Sand and Gravel Co. pit, west of U. S. Rt. 113, 0.8 miles south of Rt. 357.
- Je-II. Abandoned pit 0.2 miles west of Rt. 363 north of Cypress Branch.
- Kb-I. D.S.H.D. pit 0.7 miles north of Rt. 10 at Sandtown, Del.
- Kb-II. Clough and Caulk Sand and Gravel Co. pit, 0.3 miles north of Rt. 10 at Meredith Branch.
- Kc-I. Abandoned pit 0.1 mile west of intersection of Rts. 249 and 251.
- Kc-II. Roadcut south side of Rt. 232, 0.7 miles south of Rt. 10.
- Kd-I. Abandoned pit west of Rt. 232, 0.1 mile north of Rt. 54, near Woodside, Del.
- Kd-II. Roadcut on east side Rt. 380 north of Pratt Branch.
- Ke-I. Abandoned pit 1.2 miles east of U. S. Rt. 113 at Murderkill River.
- Ke-II. Pit, now abandoned, 0.5 miles south of Rt. 371, 0.5 miles west of U. S. Rt. 113.
- Lb-I. Abandoned pit north of Rt. 267, 0.3 miles east of Rt. 268.
- Lb-II. Abandoned pit 0.1 mile east of Rt. 291 at Bullock Prong.
- Lc-I. Drainage ditch 0.5 miles north of Rt. 59 at Marshy Hope Ditch.
- Lc-II. Roadcut east of Rt. 284, 0.15 miles south of Black Swamp Creek.
- Ld-I. Pit 0.1 mile east of Rt. 384 at Browns Branch.
- Ld-II. Pit 0.3 miles east of Rt. 384 north of Murderkill River.
- Le-I. Pit 0.1 mile west of Rt. 391, 0.2 miles south of Rt. 390.
- Le-II. Pit 0.5 miles northwest of Rt. 391, 0.8 miles south of Rt. 390.
- Lf-I. Abandoned pit 0.1 mile south of Rt. 410, 0.1 mile east of Rt. 124.
- Lf-II. Abandoned pit at Bowen Landing, 0.2 miles east of Rt. 409 on Mispillion River.
- Mb-I. Abandoned pit 0.2 miles east of Rt. 112, 0.8 miles south of Rt. 113.
- Mb-II. Abandoned pit 0.1 mile west of Rt. 301, 0.5 miles south of Rt. 14.
- Mc-I. Drainage ditch 0.1 mile south of Rt. 62, 0.6 miles east of Rt. 309.
- Mc-II. Roadcut and ditch east of Rt. 455, 0.2 miles north of Vernon, Del.
- Md-I. Nanticoke Watershed East Ditch 0.5 miles north of Rt. 439.
- Md-II. Prong of Nanticoke Watershed Ditch 0.2 miles southwest of Staytonville west of Rt. 36.

- Me-I. Abandoned pit 0.3 miles south of Rt. 634 on west side of Johnson Branch.
- Me-II. Abandoned pit (dump) east of Rt. 225, 0.4 miles south of Rt. 38 near Lincoln, Del.
- Mf-I. D.S.H.D. pit 0.2 miles north of Rt. 209, 0.5 miles east of Rt. 36, near Milford, Del.
- Mf-II. Abandoned pit north of Rt. 224, 0.3 miles east of Rt. 212.
- Mg-I. Roadcut west side of Rt. 222, 0.1 mile north of Rt. 38.
- Mg-II. Roadcut and ditch north of Rt. 219, 0.2 miles east of Rt. 220.
- Nb-I. Abandoned pit 0.1 mile east of intersection of Rts. 569 and 578 north of Woodenhawk, Del.
- Nb-II. Cut south of Rt. 404 at west end of Woodenhawk Bridge at Marshy Hope Creek.
- Nc-I. Abandoned pit 0.2 miles north of intersection of Rts. 16 and 390.
- Nc-II. Drainage ditch east side of Rt. 562, 0.6 miles north of Rt. 31.
- Nd-I. Abandoned pit 0.1 mile south of Rt. 565, 0.1 mile east of Rt. 591.
- Nd-II. Abandoned pit 0.1 mile east of Rt. 611, 0.2 miles north of Rt. 597.
- Ne-I. Abandoned pit southeast of Rt. 42 between Rts. 596 and 638.
- Ne-II. Drainage ditch north side of Rt. 16 at Oakley, Del.
- Nf-I. Abandoned pit south of Rt. 16 at intersection with Rt. 226.
- Nf-II. Abandoned pit south of Rt. 238, 0.8 miles west of Rt. 16.
- Ng-I. Abandoned pit 0.1 mile south of Rt. 38, 0.4 miles west of Rt. 14.
- Ng-II. Abandoned pit east of Rt. 14, 0.1 mile south of Primehook Creek.
- Nh-I. Pit south of Pennsylvania Railroad, 0.7 miles east of Nassau, Del.
- Nh-II. Abandoned pit 0.1 mile east of junction of Rts. 258 and 264.
- Ob-I. Roadcut 100 yds. east of Md. Rt. 306, 0.1 mile north of Houston Branch.
- Ob-II. Ditch in southeast corner of intersection of Rts. 18 and 558 west of Atlanta, Del.
- Oc-I. Abandoned pit west of Rt. 564 north bank of Bridgeville Branch.
- Oc-II. Ditch west side of Rt. 560, 0.1 mile north of Rt. 30.
- Od-I. Pit west of Rt. 525, 0.2 miles north of Rt. 526.
- Od-II. Abandoned pit 0.1 mile south of Rt. 526, 0.6 miles east of Rt. 527.
- Oe-I. Ditch south of Rt. 18, west side of Deep Creek.
- Oe-II. Roadcut east of Rt. 527, 2.0 miles north of Rt. 18.
- Of-I. Pit 0.1 mile south of Rt. 321, 0.2 miles west of Rt. 318.
- Of-II. Abandoned pit north of Pennsylvania Railroad 0.15 miles east of Rt. 309 near Georgetown, Del.
- Og-I. Pit south of Rt. 259, 0.3 miles west of Rt. 258 near Hunters Millpond.
- Og-II. D.S.H.D. pit 0.1 mile south of Rt. 18 at Gravel Hill, Del.
- Oh-I. Pit 0.3 miles west of Rt. 283, 0.5 miles south of Rt. 275.
- Oh-II. Roadcut east side of Rt. 277, 0.15 miles south of Rt. 24 at Angola Grange.
- Oi-I. Pit 0.3 miles north of Rt. 14 at Midway, Del.
- Oi-II. Pit south of Pennsylvania Railroad 0.2 miles southeast of Rt. 270.
- Pb-I. D.S.H.D. pit, 0.2 miles south of Rt. 20, 1.6 miles east of Reliance, Del.
- Pb-II. Abandoned pit 0.1 mile north of Rt. 79, 0.8 miles southwest of Woodland, Del.
- Pc-I. Figgs' Pit, east of Rt. 556, 0.6 miles south of Rt. 20.
- Pc-II. Abandoned pit west side of Rt. 478A, 0.3 miles south of Rt. 78.
- Pd-I. Abandoned pit east of Rt. 485, 0.1 mile south of Tubbs Branch, south of Concord, Del.
- Pd-II. North bank of Dukes and Jobs Ditch just west of Rt. 446.
- Pe-I. Abandoned pit 0.5 miles southeast of intersection of Rts. 20 and 442.

- Pe-II. Ditch east side of Rt. 444, 100 yds. north of Rt. 28.
- Pf-I. Abandoned pit 0.1 mile west of Rt. 435, 0.7 miles north of Rt. 20.
- Pf-II. Melvin Joseph Co. pit, east of Rt. 326, 0.1 mile south of Stockley Branch.
- Pg-I. Abandoned pit 0.2 miles southwest of Rt. 302A at Simpler Branch.
- Pg-II. Abandoned pit east of intersection of Rts. 410 and 328A west of Millsboro, Del.
- Ph-I. Abandoned pit west of Rt. 297, 0.1 mile north of Rt. 24 at Harmon School, Del.
- Ph-II. Drainage ditch north of Rt. 302, 0.9 miles east of Rt. 5.
- Qb-I. Abandoned pit north of Rt. 506, 0.3 miles west of Rt. 498.
- Qb-II. Howard Sand and Gravel Co. pit, 0.3 miles south of Md. Rt. 313, 0.5 miles west of Sharptown, Md.
- Qc-I. Pit, north bank of Broad Creek, 0.7 miles east of Bethel, Del.
- Qc-II. Pit, east of Rt. 493, 0.7 miles south of Portsville, Del.
- Qd-I. Abandoned pit east of Rt. 467, 0.2 miles north of Rt. 466.
- Qd-II. Abandoned pit 0.2 miles south of Rt. 20A, 0.2 miles west of Pennsylvania Railroad at Laurel, Del.
- Qe-I. Drainage ditch at intersection of Rts. 62 and 447.
- Qe-II. Abandoned pit north of Rt. 451, 0.1 mile southeast of Rt. 464.
- Qf-I. Abandoned pit east of Rt. 426, 0.1 mile south of junction with Rt. 424.
- Qf-II. Abandoned pit 0.1 mile north of Rt. 410, 0.2 miles south of intersection with Rt. 409.
- Qg-I. Atkin Brothers Sand and Gravel Co. pit, 0.3 miles west of Rt. 113 at Iron Branch.
- Qg-II. Roadcut east side of Rt. 83 south of Iron Branch.
- Qh-I. Pit 0.2 miles west of Rt. 343, 0.1 mile south of Rt. 26.
- Qh-II. Ditch west side of Rt. 356 (Honolulu Ave.), 0.1 mile north of Rt. 54 in Frankford, Del.
- Qi-I. Pit 0.2 miles north of Rt. 346, 0.15 miles west of junction with Rt. 347.
- Qi-II. Abandoned pit east of Rt. 347, 0.4 miles north of Rt. 26.
- Rc-I. Abandoned pit east of Rt. 509, 0.6 miles north of Rt. 508.
- Rc-II. Ditch west of Rt. 510, 1.0 mile north of Rt. 76.
- Rd-I. Abandoned pit west of U. S. Rt. 13, 1.0 mile south of Delaware-Maryland line.
- Rd-II. Abandoned pit 0.3 miles east of U. S. Rt. 13, 1.25 miles south of Delaware-Maryland line.
- Re-I. Abandoned pit in Maryland 0.8 miles south of Whitesville, Del.
- Re-II. Ditch (dump) south of Rt. 451, 0.6 miles east of Rt. 66.
- Rf-I. Pit 0.2 miles southeast of Rt. 417, 0.2 miles east of junction with Rt. 413, near Gumboro, Del.
- Rf-II. Abandoned pit 0.2 miles north of Rt. 413, 0.3 miles west of junction with Rt. 419.
- Rg-I. Drainage ditch in Maryland west side of extension of Delaware Rt. 418 at Careytown Branch.
- Rg-II. Drainage ditch 0.5 miles south of Delaware-Maryland line 1.8 miles east of Del. Rt. 418.
- Rh-I. Abandoned pit east of U. S. Rt. 113, 0.1 mile south of Md. Rt. 367.
- Rh-II. Drainage ditch, west of Rt. 380, 0.75 miles south of U. S. Rt. 113.
- Ri-I. Pit 0.35 miles north of Rt. 382, 0.3 miles west of intersection with Rt. 58.
- Ri-II. Pit 0.1 mile north of Rt. 396, 0.5 miles west of Rt. 396A.

APPENDIX II

Formulae Used in Computation

1. Mechanical Analysis: (after Inman, 1952)

$$\begin{aligned}\sigma\phi &= \frac{1}{2}(\phi 84 - \phi 16) \\ M\phi &= \frac{1}{2}(\phi 16 + \phi 84) \\ \alpha\phi &= \frac{M\phi - Md\phi}{\sigma\phi}\end{aligned}$$

Where:

$\sigma\phi$ = coefficient of sorting
 $\phi 84$ = size, in ϕ units, at 84th percentile
 $\phi 16$ = size, in ϕ units, at 16th percentile
 $M\phi$ = mean diameter
 $Md\phi$ = median diameter
 $\alpha\phi$ = skewness value

2. Cross-bedding Analysis: Computation of vector mean (after Potter and Pettijohn, 1963; cf. Curray, 1956)

$$V = \sum_{i=1}^n Ni \cos Xi$$

$$W = \sum_{i=1}^n Ni \sin Xi$$

$$\begin{aligned}\bar{x} &= \arctan W/V \\ R &= (V^2 + W^2)^{1/2} \\ L &= (R/n) 100\end{aligned}$$

Where:

xi = mid-point azimuth of i th class interval
 \bar{x} = azimuth of resultant vector
 ni = number of observations in each class
 n = total number of observations
 R = magnitude of resultant vector
 L = magnitude of resultant vector in percent

APPENDIX III

Mechanical Composition: Statistical Parameters

Sample Number	$\phi 16$	$\phi 84$	$Md\phi$	$M\phi$	$\sigma\phi$	$\alpha\phi$
Cb-I	0.34	3.00	1.49	1.670	1.330	0.1353
II	-4.10	1.62	-1.06	-1.240	2.860	-0.0629
Cc-I	-4.82	1.56	0.36	1.630	3.190	0.3981
II	-0.52	2.05	0.90	0.765	1.285	-0.1050
Cd-I	-6.50*	0.86	-0.56	-3.680*	3.680*	-0.8478*
II	-4.97	1.25	-0.36	-1.860	3.110	-0.4823
Da-I	-0.38	2.78	1.09	1.200	1.58	0.0696
II	-2.94	1.33	0.17	-0.805	2.135	-0.4567
Db-I	0.12	1.87	0.81	0.995	0.875	0.2114
II	-3.78	2.01	0.70	-0.885	2.890	-0.5484
Dc-I	-6.50*	0.90	-1.82	-2.800*	3.700*	-0.2648*
II	-5.50*	1.65	0.30	-1.925*	3.575*	-0.6223*
Ea-I	-1.60	1.71	0.69	0.055	1.655	-0.3233
II	-6.50*	1.02	-2.40	-2.740*	3.760*	-0.0904*
Eb-I	-5.50*	1.51	0.12	-1.995*	3.505*	-0.6034*
II	-3.73	1.93	0.80	-0.900	2.830	-0.6007
Ec-I	0.19	1.74	0.94	0.965	0.775	0.0323
II	-0.18	1.97	0.99	0.895	1.075	-0.0883
Fb-I	-2.20	1.52	0.28	-0.340	1.860	-0.3333
II	-2.80	1.54	0.54	-0.630	2.170	-0.5391
Fc-I	-4.04	1.73	0.45	-1.155	2.885	-0.5563
II	-4.29	1.50	0.03	-1.395	2.895	-0.4922
Gb-I	0.01	1.99	1.02	1.000	0.990	-0.0202
II	0.70	3.28	1.70	1.990	1.290	0.2248
Gc-I	-3.95	0.88	-0.30	-1.535	2.415	-0.5113
II	0.43	2.80	1.40	1.615	1.185	0.1814
Hb-I	-0.30	2.28	1.14	0.990	1.290	-0.1162
II	0.22	2.88	1.51	1.500	1.330	-0.0075
Hc-I	-1.40	2.30	0.90	0.450	1.850	-0.2432
II	-1.65	2.58	1.34	0.465	2.115	-0.4137
Hd-I	1.08	2.80	1.73	1.940	0.860	0.2441
II	1.07	6.50*	2.10	3.785*	2.715*	0.6206*
Ib-I	0.05	1.86	0.94	0.955	0.905	0.0165
II	0.17	2.05	0.90	1.110	0.940	0.2234
Ic-I	-5.50*	1.50	0.17	-2.000*	3.500*	-0.0620*
Ic-II	-7.50*	1.71	0.10	-2.895*	4.605*	-0.6513*
Id-I	-0.30	2.34	1.24	1.020	1.320	-0.1666
II	-2.30	2.28	1.42	-0.010	2.290	-0.6157
Jb-I	1.02	2.74	1.79	1.880	0.860	0.1046
II	1.00	2.88	1.80	1.940	0.940	0.1489
Jc-I	0.64	2.81	1.55	1.725	1.085	0.1612
II	-0.60	1.91	1.06	0.655	1.255	-0.3227

Appendix III (Cont.)

Sample Number	$\phi 16$	$\phi 84$	$Md\phi$	$M\phi$	$\sigma\phi$	$\alpha\phi$
Jd-I	-3.57	1.22	-0.20	-1.175	2.395	-0.3839
II	-0.50	1.89	0.69	0.695	1.195	0.0041
Je-I	-2.04	1.68	0.67	-0.180	1.860	-0.4032
II	0.68	2.56	1.72	1.620	0.940	-0.1063
Kb-I	-4.32	1.85	0.35	-1.235	3.085	-0.5137
II	-4.61	1.78	0.47	-1.415	3.195	-0.5899
Kc-I	0.67	3.04	1.64	1.855	1.185	0.2067
II	0.53	3.79	1.70	2.160	1.630	0.2822
Kd-I	0.14	2.19	1.15	1.165	1.025	0.0146
II	-0.90	1.67	0.70	0.385	1.285	-0.2451
Ke-I	-1.16	1.68	0.61	0.26	1.420	-0.2465
II	-0.31	1.84	0.70	0.765	1.075	0.0605
Lb-I	0.78	2.91	1.69	1.845	1.065	0.1455
II	0.63	2.66	1.60	1.645	1.015	0.0443
Lc-I	1.12	4.10*	2.08	2.610*	1.490*	0.3557*
II	1.13	3.10	1.84	2.115	0.985	0.2791
Ld-I	0.08	2.07	1.18	1.075	0.995	-0.1055
II	-0.60	1.95	0.81	0.675	1.275	-0.1058
Le-I	-0.46	2.47	1.03	1.005	1.47	-0.2789
II	-1.03	1.91	0.72	0.440	1.470	-0.1904
Lf-I	-3.43	1.29	0.14	-1.070	2.360	-0.5127
II	0.35	2.89	1.23	1.620	1.270	0.3070
Mb-I	0.94	2.60	1.66	1.770	0.830	0.1325
II	1.16	3.17	1.98	2.165	1.005	0.1840
Mc-I	0.86	3.31	1.68	2.085	1.225	0.3306
II	1.16	2.85	1.94	2.005	0.845	0.0769
Md-I	0.76	2.15	1.41	1.855	0.695	0.6402
II	0.26	1.88	1.06	1.070	0.810	0.0123
Me-I	0.32	2.33	1.10	1.325	1.005	0.2238
II	1.02	2.40	1.57	1.710	0.690	0.2028
Mf-I	-0.60	2.00	0.86	0.700	1.300	-0.1230
II	-0.04	3.71	1.61	1.835	1.875	0.1200
Mg-I	0.59	3.90	1.80	2.245	1.705	0.2609
II	0.91	4.90*	1.70	2.905*	1.995*	0.6040*
Nb-I	0.94	2.58	1.69	1.760	0.820	0.0853
II	1.12	2.33	1.69	1.725	0.605	0.0578
Nc-I	0.29	3.33	1.63	1.810	1.520	0.1184
II	0.76	3.89	1.99	2.325	1.565	0.2140
Nd-I	1.08	3.16	2.04	2.120	1.040	0.0769
II	0.66	3.40	2.10	2.030	1.370	-0.0510
Ne-I	0.04	2.50	1.30	1.270	1.230	-0.0243
II	1.48	4.10*	2.49	2.790*	1.310*	0.2290*
Nf-I	-0.36	2.45	1.10	1.045	1.405	-0.0391

Appendix III (Cont.)

Sample Number	$\phi 16$	$\phi 84$	$Md\phi$	$M\phi$	$\sigma\phi$	$\alpha\phi$
II	0.40	2.80	1.46	1.600	1.200	0.1166
Ng-I	-0.04	1.72	0.89	0.840	0.880	-0.0568
II	-0.32	1.91	0.80	0.795	1.115	-0.0044
Nh-I	0.76	2.19	1.61	1.475	0.715	-0.1888
II	0.98	3.10	1.86	2.040	1.060	0.1698
Ob-I	0.54	2.91	1.62	1.725	1.185	0.0886
II	1.04	3.53	2.08	2.285	1.245	0.1646
Oc-I	0.60	3.30	1.71	1.950	1.350	0.1777
II	0.89	2.83	1.79	1.860	0.970	0.0721
Od-I	0.04	1.99	0.81	1.015	0.975	0.2102
II	0.96	2.70	1.78	1.830	0.870	0.0574
Oe-I	0.99	2.80	1.69	1.895	0.905	0.2265
II	1.29	2.76	2.06	2.025	0.735	-0.0476
Of-I	1.97	11.40*	3.49	6.685*	4.715*	0.6776*
II	0.16	2.90	1.30	1.530	1.370	0.1678
Og-I	1.03	2.72	1.84	1.875	0.845	0.0414
II	-2.31	2.37	1.11	0.030	2.340	-0.4615
Oh-I	-0.72	1.04	0.23	0.160	0.880	-0.0795
II	0.77	2.61	1.63	1.690	0.920	0.0652
Oi-I	-0.30	2.24	1.40	0.970	1.270	-0.3385
Oi-II	0.05	2.68	1.62	1.365	1.315	-0.1939
Pb-I	-1.20	2.71	1.10	0.755	1.955	-0.1764
II	1.04	2.98	2.11	2.010	0.970	-0.1030
Pc-I	-0.76	3.20	1.50	1.220	1.980	-0.1414
II	0.94	2.53	1.70	1.735	0.795	0.0440
Pd-I	1.08	2.09	1.72	1.585	0.505	-0.2673
II	-0.04	2.06	1.12	1.010	1.050	-0.1047
Pe-I	1.02	2.70	1.81	1.860	0.840	0.0595
II	0.80	3.40	1.92	2.100	1.300	0.1384
Pf-I	1.02	2.96	2.06	1.990	0.970	-0.0721
II	0.01	1.78	0.97	0.895	0.885	-0.0847
Pg-I	1.14	2.65	1.79	1.895	0.755	0.1390
II	-2.10	2.77	1.26	0.335	2.435	-0.3798
Ph-I	1.32	3.36	2.19	2.340	1.020	0.1470
II	1.30	3.76	2.38	2.530	1.230	0.1219
Qb-I	0.62	2.11	1.35	1.365	0.745	0.0201
II	-1.74	0.70	-0.26	-0.520	1.220	-0.2131
Qc-I	-1.16	3.55	1.10	1.195	2.355	0.0785
II	-1.92	1.90	0.45	-0.010	1.910	-0.2408
Qd-I	1.50	4.75*	2.48	3.125*	1.625*	0.3969*
II	-0.10	2.11	0.91	1.005	1.105	0.0859
Qe-I	1.12	3.80	2.07	2.460	1.340	0.2910
II	1.21	4.03*	2.22	2.620*	1.410*	0.2836*

Appendix III (Cont.)

Sample Number	$\phi 16$	$\phi 84$	$Md\phi$	$M\phi$	$\sigma\phi$	$\alpha\phi$
Qf-I	1.01	2.80	1.93	1.905	0.895	-0.0279
II	0.00	2.70	1.38	1.350	1.350	-0.0222
Qg-I	0.68	2.87	1.74	1.775	1.095	0.0319
II	0.59	2.73	1.56	1.660	1.070	0.0934
Qh-I	0.42	2.5	1.30	1.470	1.050	0.1619
II	1.28	4.10*	2.29	2.690*	1.410*	0.2836*
Qi-I	0.67	6.54	1.94	4.105	2.435	0.3833
II	1.20	2.96	1.96	2.080	0.880	0.1363
Rc-I	0.96	2.41	1.62	1.685	0.725	0.0896
II	2.21	9.40*	3.51	5.805*	3.595*	0.6383*
Rd-I	1.70	4.20*	2.63	2.950*	1.250*	0.2560
II	1.43	3.70	2.35	2.565	1.135	0.1894
Re-I	1.07	2.77	1.91	1.920	0.850	0.0117
II	1.14	2.88	1.89	2.010	0.870	0.1379
Rf-I	1.32	2.73	1.97	2.025	0.705	0.0780
II	1.09	2.68	1.82	1.885	0.795	0.0818
Rg-I	1.20	3.00	2.03	2.100	0.900	0.0333
II	1.30	3.00	2.07	2.150	0.850	0.0941
Rh-I	1.28	2.85	2.10	2.065	0.785	-0.0446
II	1.22	2.73	1.94	1.975	0.755	0.0464
Ri-I	0.02	3.58	1.67	1.800	1.780	0.0730
II	1.37	3.17	2.22	2.270	0.900	0.0555

*Indicates value obtained by estimation from extended cumulative curve or calculated from such an estimation.

APPENDIX IV

Gross Mineralogy of Sand Fraction

Percentages of minerals are given to the nearest whole number. *T* = trace.

Sample					Sample				
Number	Quartz	Feldspar	Mica	Other	Number	Quartz	Feldspar	Mica	Other
Cb-I	55	39	4	3	Nd-I	93	7		
Cc-II	57	37	3	5	Ne-I	80	20		1
Cd-II	66	26	3	6	Nf-I	92	8		1
Da-II	94	5		2	Ng-I	94	6		
Db-II	81	15	<i>T</i>	4	Nh-I	88	12		<i>T</i>
Dc-II	68	26	3	4	Ob-II	78	21	<i>T</i>	1
Ea-II	90	10			Oc-I	68	28	4	<i>T</i>
Eb-II	73	25	<i>T</i>	3	Od-I	94	6		<i>T</i>
Ec-II	71	28	<i>T</i>	2	Oe-I	93	6		3
Fb-I	66	30	1	4	Of-I	77	23		
Fc-I	68	32	<i>T</i>		Og-I	90	9		2
Gb-I	90	10		<i>T</i>	Oh-II	83	15	1	1
Gc-I	76	21	<i>T</i>	3	Oi-I	87	13		1
Hb-I	89	10		1	Pb-II	92	7		1
Hc-I	67	27	6	2	Pc-I	79	21		
Hd-II	76	23		1	Pd-I	87	12		1
Ib-I	75	24	<i>T</i>	2	Pe-I	74	26	<i>T</i>	1
Ic-I	77	19	2	3	Pf-II	80	20	<i>T</i>	<i>T</i>
Id-II	88	12		1	Pg-II	95	5		<i>T</i>
Jb-I	80	19		1	Ph-II	88	12		
Jc-II	65	35		<i>T</i>	Qb-I	83	17		<i>T</i>
Jd-I	72	27		2	Qc-I	71	29		1
Je-II	70	29		2	Qd-II	83	17		<i>T</i>
Kb-I	96	5			Qe-I	81	19	<i>T</i>	
Kc-I	79	21		<i>T</i>	Qf-I	86	15		
Kd-I	68	32		<i>T</i>	Qg-I	76	25		
Ke-I	68	31	1	<i>T</i>	Qh-I	83	16		2
Lb-I	80	20	<i>T</i>	<i>T</i>	Qi-I	71	27		3
Lc-I	87	13		<i>T</i>	Rc-I	94	7		
Ld-II	68	30		3	Rd-II	91	8		2
Le-I	74	24	2	1	Re-II	84	16		<i>T</i>
Lf-II	73	26		2	Rf-I	89	11		
Mb-I	90	10			Rg-I	72	28		<i>T</i>
Mc-I	75	19	3	4	Rh-I	81	19		
Md-I	82	15	1	2	Ri-I	68	31		1
Me-II	89	11		1					
Mf-I	91	10							
Mg-I	96	4		<i>T</i>					
Nb-I	91	8	<i>T</i>	<i>T</i>					
Nc-I	87	10	2	2					

APPENDIX V

Heavy Minerals

Percentages of heavy minerals are given to the nearest whole number.
Abbreviations used for the trace minerals reported as "other" are:
Ap - apatite; Mon - monazite; Sph - sphene; Sp - spinel.

Sample Number	% Heavy Minerals (500-62 μ)	% Heavy Minerals (total)	% Opaque	Amphibole	Andalusite	Chloritoid	Epidote	Garnet	Kyanite	Pyroxene	Rutile	Sillimanite	Staurolite	Tourmaline	Zircon	Altered	Unidentified	Other	Stable Mineral Index
Cb-I	1.77	1.14	77.2	28	1		41	1	1		2	11	3	1	10	1	1	1 Ap	0.15
Cc-II	1.23	0.52	78.5	35		2	28	1	4		3	9	2	5	8	3	1	1 Ap	0.19
Cd-II	4.38	0.74	70.9	67			7	3		6	2	3	2	2	5	3	1		0.10
Da-II	6.48	1.70	83.6	19			9		2			11	20	11	29				0.66
Db-II	5.46	2.19	86.5	18	1	3	12	2	4		4	12	6	5	33	1	1		0.67
Dc-II	1.44	0.44	85.5	17			7	3	5		4	10	4	9	34	6		1 Sph	0.89
Ea-II	1.28	0.59	82.2	18	1	2	12	1	1		6	9	23	8	18	1	1		0.46
Eb-II	1.62	0.68	73.5	25			26	6	3	2	2	10	6	6	13	2	1		0.26
Ec-II	2.89	1.33	70.8	28	1	1	21	2	3	2	2	12	4	8	18				0.38
Fb-I	0.79	0.20	85.2	11		1	24	3	1		6	10	10	8	25		2		0.63
Fc-I	1.72	0.58	79.4	16		1	12	2	4		4	11	4	16	26	1	3		0.85
Gb-I	1.05	0.47	82.7	13		1	22			1	4	9	1	4	39	2	4		0.89
Gc-I	1.86	0.25	86.3	11			14		4	4	4	12	5	5	36	3	4		0.79
Hb-I	2.50	1.32	83.8	9	1		12	2	2	1	2	13	3	4	48	3	1	1 Ap	1.13
Hc-I	4.00	1.67	90.0	16	1		22	2			5	16	4	2	29	3	1		0.55

Hd-II	0.79	0.44	70.0	12	2	2	18	1	1	1	6	8	3	4	41	1	2	1.00
Ib-I	1.49	0.65	80.0	19		2	30	1	3			11	4	3	26	1		0.41
Ic-I	3.47	1.00	70.4	19	2		45	1	3		2	6	5		11	3	4	0.15
Id-II	0.52	0.35	85.6	8	1	1	15	2	3		1	11	9	16	33	1		0.98
Jb-I	0.65	0.52	70.3	16		1	18	2	5	1	2	17	2	7	28		2	1 Ap
Jc-II	1.25	0.60	71.1	20		1	45	1	2		4	7	1	1	16	1	3	0.26
Jd-I	5.34	0.83	96.7	7		1	16		1		4	14	3	5	42	5	1	1 Sp
Je-II	0.51	0.40	75.5	2		1	3				7	1	4	25	55	3	1	5.80
Kb-I	2.73	0.83	87.5	4		1	19	1	2		5	6	6	6	49	2		1.46
Kc-I	1.38	0.89	72.2	15		1	27	3	2		2	11	5	5	26	2	3	0.48
Kd-I	1.47	0.78	61.7	69			19	2				3	1	1	3	3		0.04
Ke-I	4.25	1.37	86.2	6	1	1	13		2	1	2	11	5	5	48	4	3	1.17
Lb-I	0.80	0.56	75.0	6			25	1	3		2	21	4	7	32			0.68
Lc-I	0.89	0.63	78.5	3		1	21	4	2		3	12	6	4	45		1	1.04
Ld-II	2.15	0.89	74.8	53			15	5			3	3	1	2	13	5	1	0.22
Le-I	2.48	1.14	88.2	2	1	1	15	4	3	1?	5	10	4	1	53		2	1.37
Lf-II	1.17	0.55	85.0	6		3	13		2	1	7	11	10	12	34		2	1.10
Mb-I	1.09	0.84	78.5	4	1	1	19		4		6	15	6	8	35	1	1	0.94
Mc-I	0.84	0.55	74.3	10			12	2	3	1	7	15	2	8	37	1	3	0.88
Md-I	0.55	0.38	74.9	12		1	35		4		2	11	4	1	26	2	3	0.40
Me-II	1.01	0.79	82.6	8		1	25	2	4	1	4	10	1	1	43		1	0.91
Mf-I	1.46	0.57	84.0	5			19	1	2		6	18	4	6	37		1	1 Ap
Mg-I	0.99	0.60	75.6	13	1	3	15	2	2	2	5	18	9	8	23	1		0.55
Nb-I	0.71	0.56	77.9	6	1	1	11	1	3	1	7	15	4	4	42		3	1.15
Nc-I	1.40	0.85	80.9			2	14	1	2	1	3	8	3	3	60		4	1.89
Nd-I	1.16	0.95	77.3	2		1	23		1	1	6	16	2	5	38	4	3	0.92
Ne-I	0.35	0.20	83.7	5	1	1	22	1	2		4	12	2	2	47	1	1	1.08
Nf-I	1.15	0.52	79.2	12		3	27	1	3		3	6	3	4	34	3	1	1 Ap
Ng-I	1.42	0.63	81.5	2			7	2	4		2	12	4	6	55	1	2	2 Ap
Nh-I	0.05	0.04	23.5	45	1	1	8	3	3	1		4	7	3	18	5	1	0.07

Appendix V (Cont.)

Sample Number	% Heavy Minerals (500-62 μ)	% Heavy Minerals (total)	% Opaque	Amphibole	Andalusite	Chloritoid	Epidote	Garnet	Kyanite	Pyroxene	Rutile	Sillimanite	Staurolite	Tourmaline	Zircon	Altered	Unidentified	Other	Stable Mineral Index
Qb-II	2.20	1.61	79.2	6		1	12	4	2		7	15	4	3	44	1	2		1.15
Oo-I	1.27	0.80	72.1	15	1	2	35	2	1	1	2	5		3	25	7	4		0.41
Od-I	1.94	0.69	82.8	5			8		4			8	3	4	62	2	2		2.13
Oe-I	1.48	1.16	80.5	6			12	3	2		7	14	4	4	43	4		2 Ap	1.15
Of-I	2.89	1.46	79.6	9		2	19	1	2		3	9	4	5	43	1	2	1 Mon	1.02
Og-I	1.54	1.16	74.1	9		2	24	2	2		2	7	4	3	41	1	3	1 Ap	0.84
Oh-II	2.02	1.51	77.0	32			8	3	2		3	9	4	3	35	3			0.67
Oi-I	0.41	0.26	39.4	48	2	3	9	11	1	3	1	6	4	4	6	2	1		0.12
Pb-II	1.70	1.37	78.3	6	1	1	8		1	1	5	14	3	14	41	3	3	1 Ap	1.43
Pc-I	2.33	1.17	82.8	6	2		16		1	1	5	16	1	5	44		3	{1 Ap 1 Mon}	1.17
Pd-I	0.65	0.55	72.1	8		1	10	1	1	1	6	24	6	7	31	2	2		0.76
Pe-I	1.50	1.22	80.9	7	1		15	1	1		8	13	3	5	45	1	1		1.35
Pf-II	0.68	0.29	76.3	13	2	1	32	1			3	13	7	3	25	1			0.44
Pg-II	2.07	1.02	80.1	9	1	24	2	3		3	9	5	3	36	1	3		2 Mon	0.71
Ph-II	1.76	1.34	83.6	9		2	7	2	1		4	15	3	9	45	3	2		1.32
Qb-I	0.93	0.63	82.9	4			12	1	7		5	18	5	4	41	2			1.02
Qc-I	2.91	1.93	80.0	12	1		18		1		2	10	1	6	41	5	3	1 Ap	0.94
Qd-II	1.52	0.66	80.6	5		2	19	1		2	5	14	3	5	42	3	1		1.04
Qe-I	1.41	1.03	79.0	5	1		8		2		3	28	3	8	37	3	3		0.91
Qf-I	1.41	1.15	77.1	4		1	16	2	1		6	18	2	4	44	1	1		1.17
Qg-I	1.52	1.05	69.2	27		1	23	1	2		1	6	3	3	24	7	1		0.39

Qh-I	2.44	1.39	79.6	10	2	13	1	1	5	21	2	2	40	3	1	0.87
Qi-I	0.71	0.37	74.4	21	1	23	3	3	2	17	3	6	23	2	2	0.44
Rc-I	0.66	0.53	74.6	5	2	14	1	1	5	23	5	10	31	3	1 Ap	0.84
Rd-II	1.05	0.84	80.1	2		8	4	4	5	30	5	9	37			1.04
Re-II	1.27	0.52	78.3	5		10	2	3	3	27	4	6	36	2	1	0.83
Rf-I	0.78	0.73	77.9	5		13	2	1	1	30	2	8	28	4	1	0.70
Rg-I	1.46	1.24	79.7		1	9	2	2	3	20	2	11	48	1	1 Ap	1.63
Rh-I	1.02	0.91	76.0	2	1	1	10	3	2	21	6	13	34	2	1	0.98
Ri-I	1.43	0.78	72.6	20	4	19			5	15	2	5	29	1	1	0.64

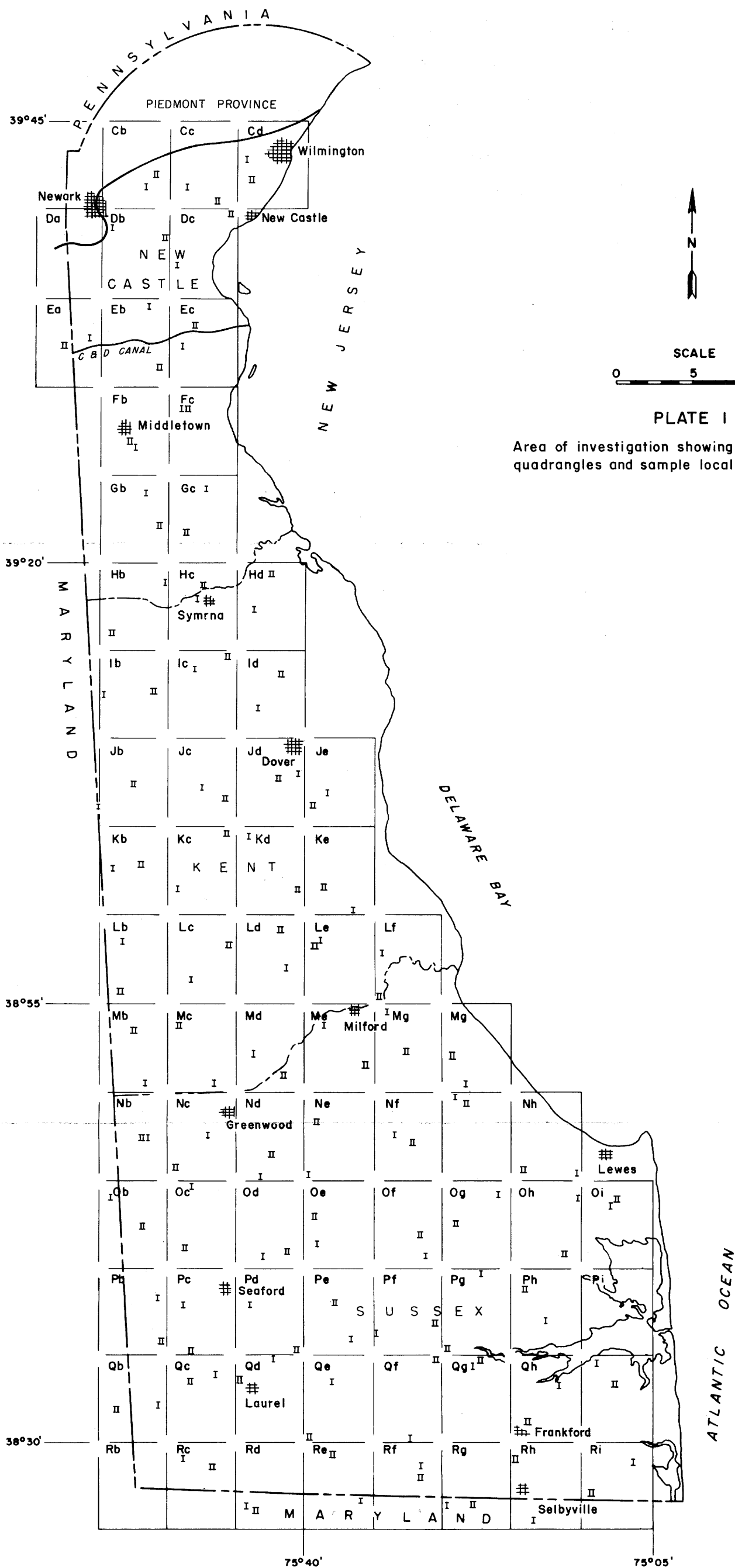
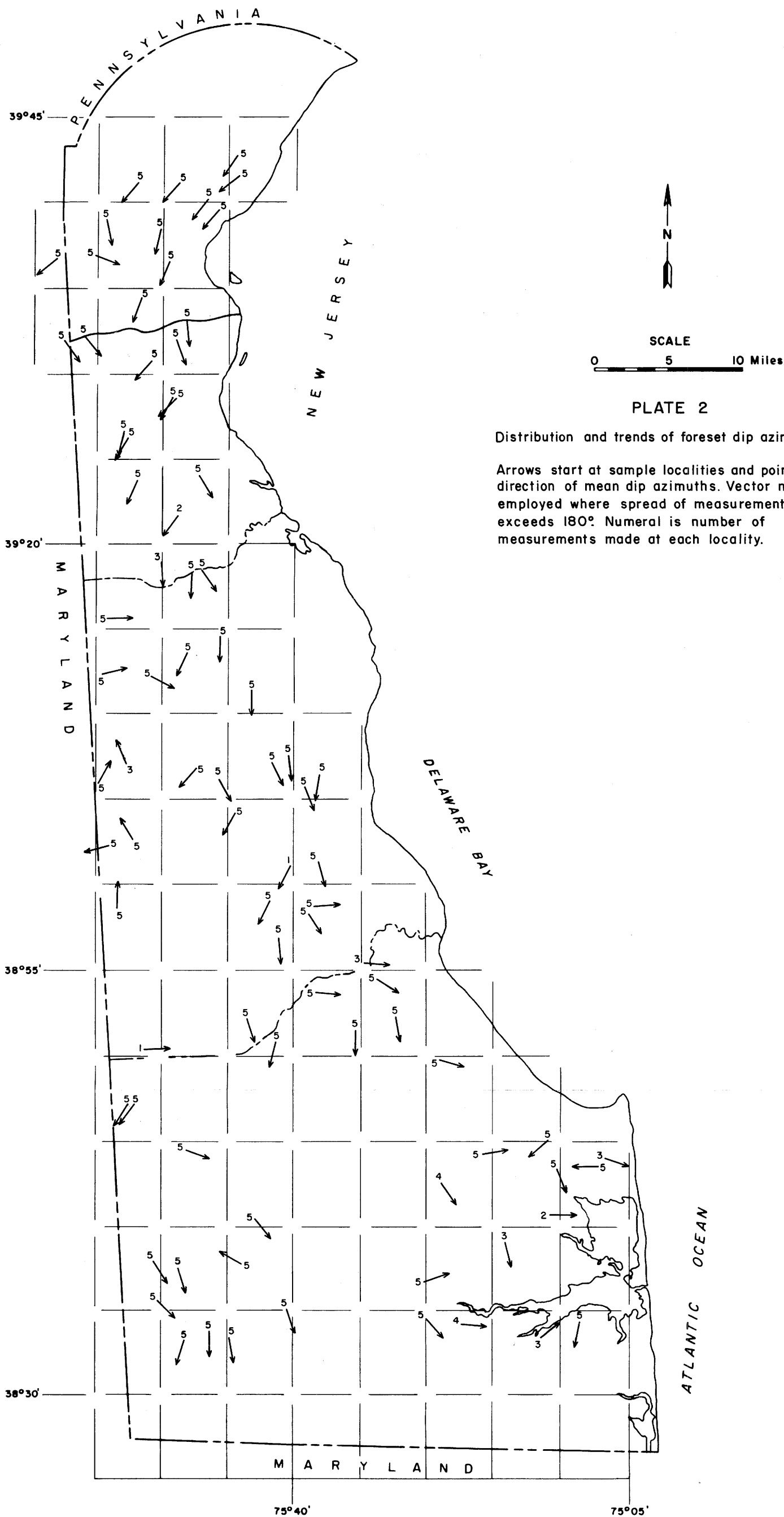
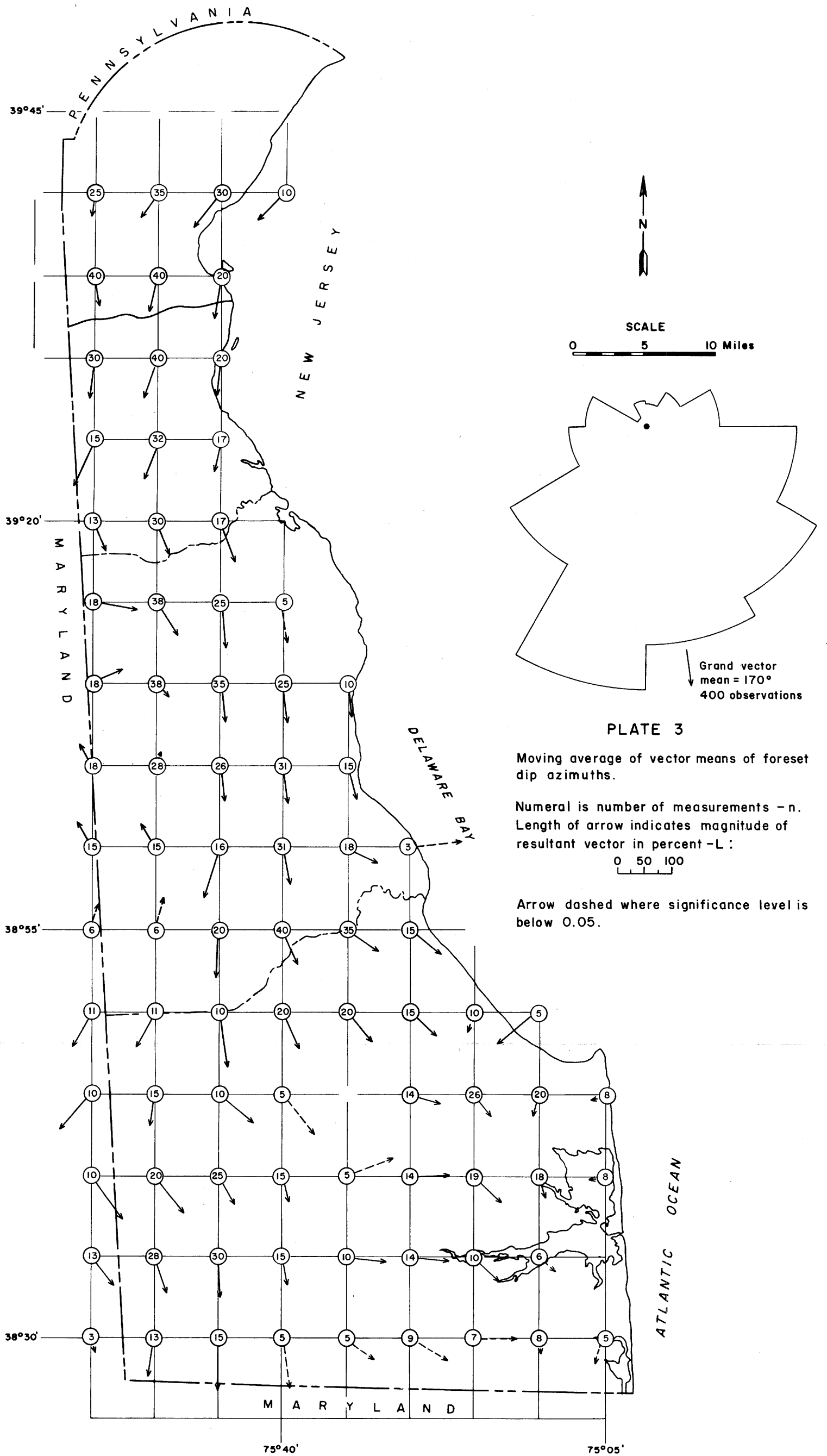


PLATE I

Area of investigation showing location of 5-minute quadrangles and sample localities.





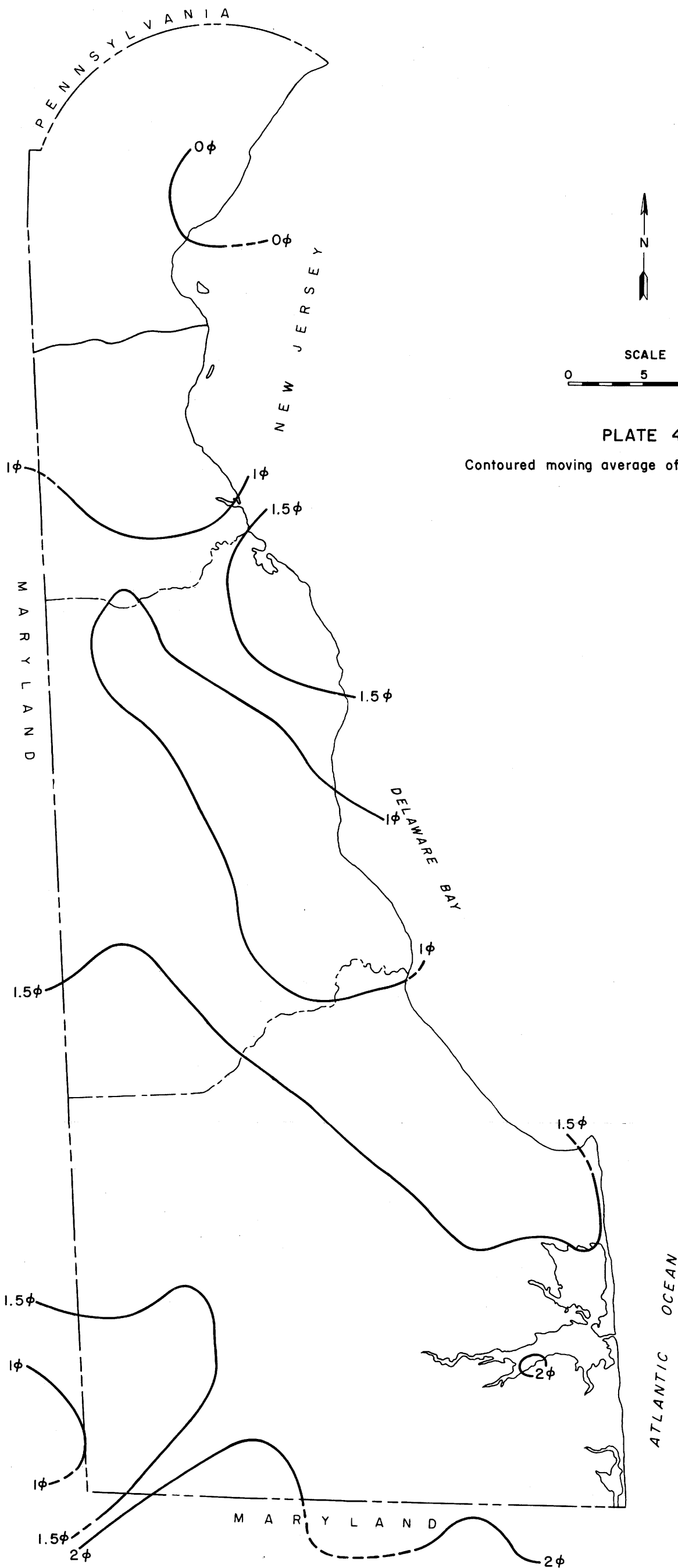


PLATE 4
Contoured moving average of median grain sizes (Mdφ).

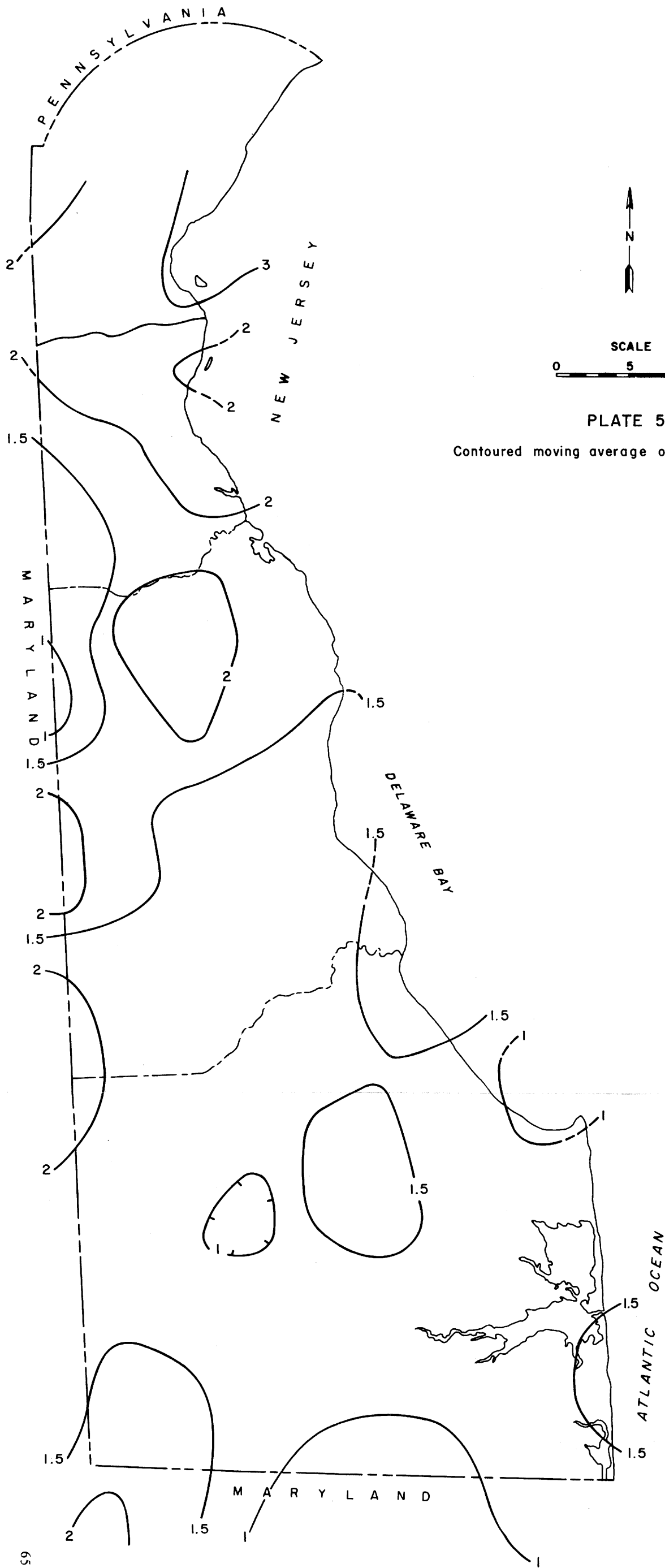


PLATE 5

Contoured moving average of sorting coefficients ($\sigma\phi$).

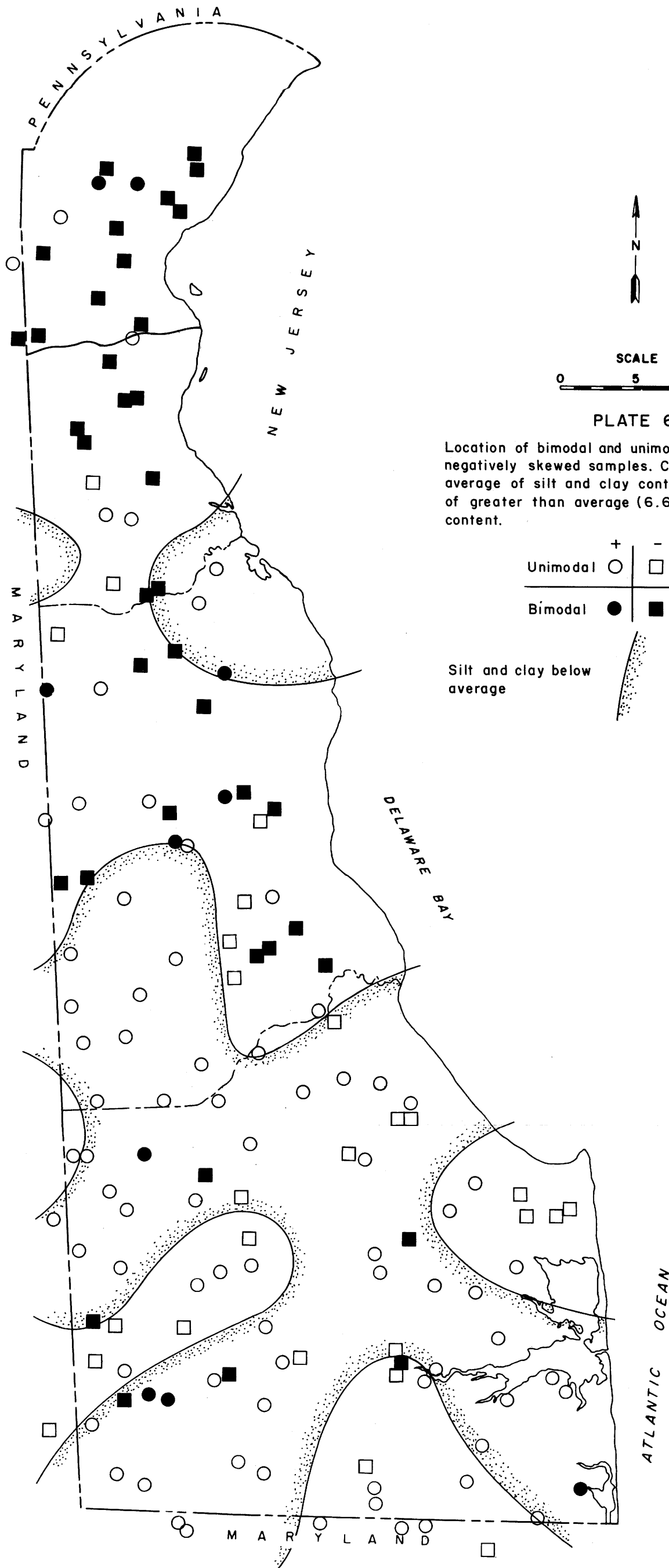


PLATE 6

Location of bimodal and unimodal, positively and negatively skewed samples. Contours of moving average of silt and clay content delineate areas of greater than average (6.6 %) silt and clay content.

	+	-	Skewness
Unimodal	○	□	
Bimodal	●	■	

Silt and clay below average

Silt and clay above average.

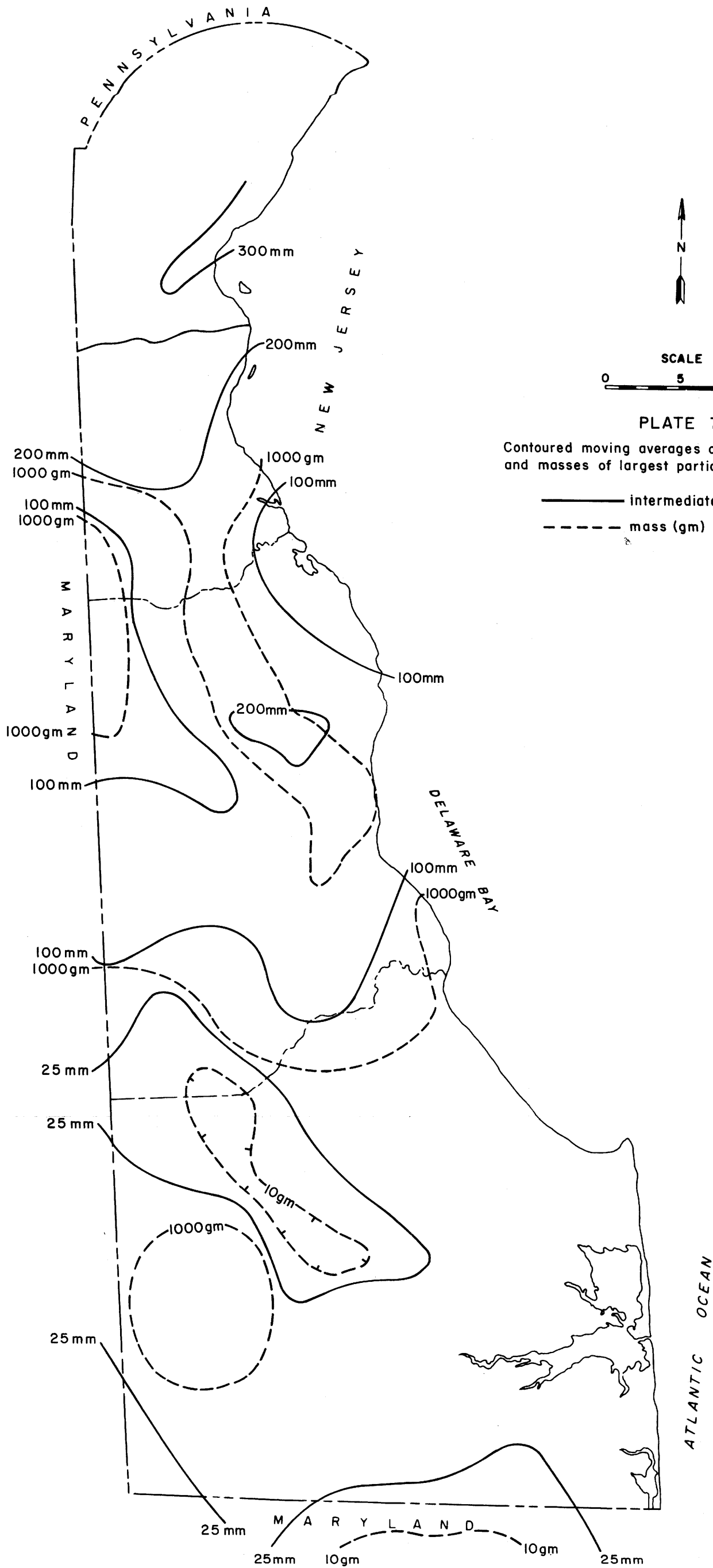


PLATE 7

Contoured moving averages of intermediate diameters and masses of largest particles.

———— intermediate diameter (mm)
 - - - - - mass (gm)

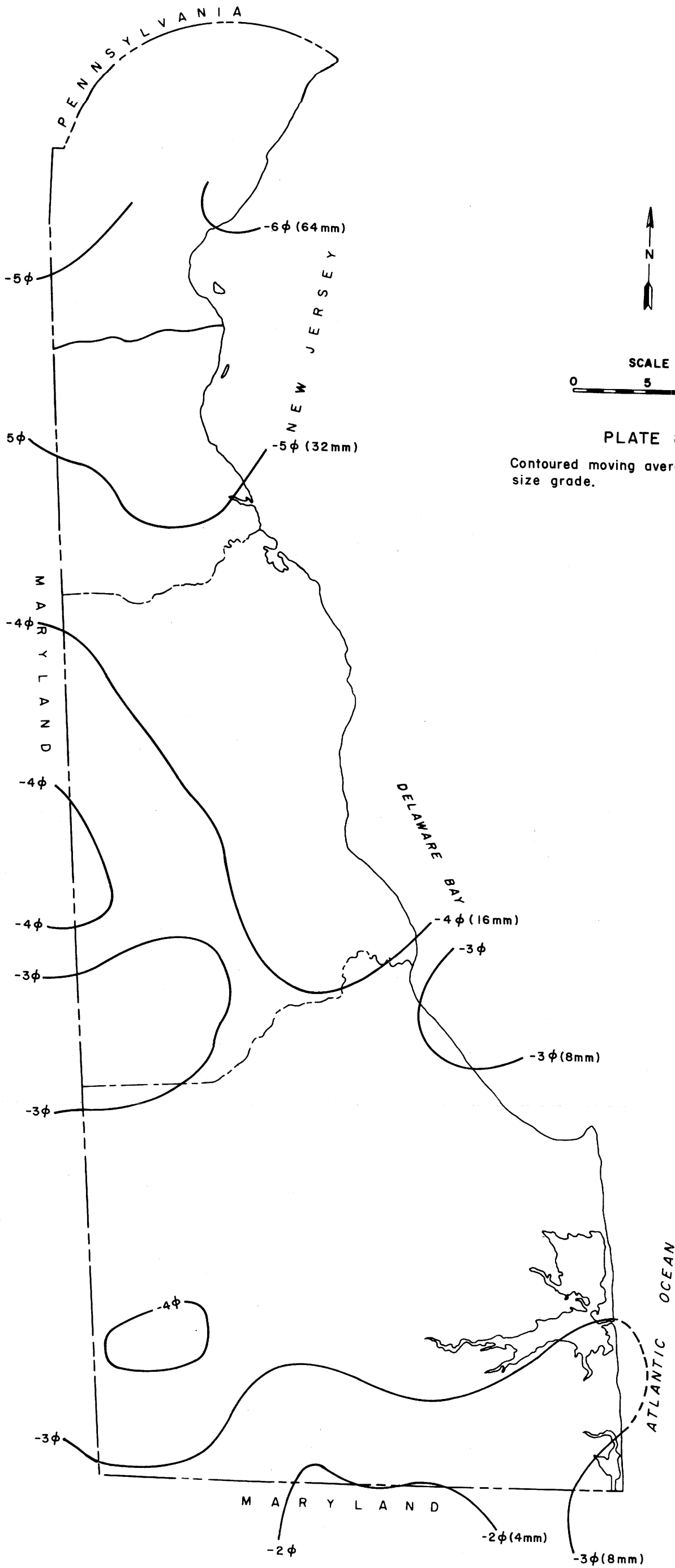


PLATE 8

Contoured moving average of maximum size grade.

