

**Institute of Polar Studies**

**Report No. 64**

# **Microparticles, Ice Sheets and Climate**

**by**

**Lonnie G. Thompson**

Institute of Polar Studies  
and  
Department of Geology and  
Mineralogy

**April 1977**



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## ABSTRACT

The Earth undergoes what appear to be major periodic changes in climate for which there must be a cause or combination of causes. The only readily available clues to past climatic change are those recorded on and in the land surfaces, the ocean floors and particularly in the polar ice sheets. The ice sheets of the world provide ideal locations for the investigation of changes with time in such climatic parameters as temperature and variations in concentration, size distribution and composition of dust. On the ice sheets annual snow accumulation averages from a few centimeters to over 100 centimeters, providing a much more expanded climatic record than is available in deep sea or lake bottom cores. The recent development of ice core drilling techniques has made it possible to recover ice cores to bedrock from Camp Century, Greenland (1966, 1387 meters) and Byrd Station, Antarctica (1968, 2164 meters).

Two initial studies of microparticle variation in the Byrd Station and the Camp Century deep ice cores have been conducted to clarify the relationship between atmospheric turbidity and climate by presenting the particle concentration and size distribution from sections of these two ice cores. These measurements have been compared with stable oxygen isotope values for ice from the same depths. This study has provided a comparison of particle concentrations and size distributions in an ice core from the Northern Hemisphere with one from the Southern Hemisphere.

In both cores the highest concentrations of particles occur where  $^{18}\text{O}/^{16}\text{O}$  ratios exhibit the greatest negative values. During the Wisconsin Glacial Stage ( $> 10,000$  years B.P.) the concentration of small diameter particles ( $0.65$  to  $0.82\ \mu\text{m}$ ) was as much as 100 times greater than the mean Holocene ( $< 10,000$  years B.P.) concentrations in the Camp Century core and more than 4 times greater than mean Holocene values for the Byrd core. Elemental composition and morphology of the microparticles suggest that most of the particles of Wisconsin age in the Byrd core are of volcanic origin and in the Camp Century core are of continental (eolian) origin.

One of the important results of this work has been the observation of microparticle variations which provide a means of dating ice sheets. The ages of the bottom ice from microparticle concentration variations, assuming an annual cycle, are much less for both cores than those ages determined from oxygen isotope and flow theory calculations. Based upon preliminary microparticle studies the age of the bottom ice at Byrd Station is estimated to be 30,000 years.

In both deep ice cores the possibility of a break in the stratigraphic record exists. In the Camp Century core the compositions

of particles in the lower 200 meters of core are atypical, as they contain substantial amounts of Sn, Mo and Cu. Particles with this same combination of elements are not found in the ice above 1200 meters nor in the entire Byrd Core. It can be argued that the source for these particles is the bedrock over which the Greenland Ice Sheet flows. If these particles below 1200 meters depth are from the bottom then the microparticle and oxygen isotope stratigraphies below this depth can not be meaningfully interpreted.

The microparticle data from these deep ice cores offer renewed support for the volcanic theory of glaciation. Therefore, a volcanic theory of climatic change is put forth. Volcanic activity can be effective in causing glaciations only when the prerequisite of appropriate land and ocean distribution exists. The large thermal inertia of the oceans, as well as orogenic and epeirogenic movements, probably play equally important roles in producing glacial climates.

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The sections of the Byrd Station and Camp Century ice core were provided by Dr. C.C. Langway, Jr., then of the U.S. Army Cold Regions Research and Engineering Laboratory, (CRREL), Hanover, New Hampshire. I am grateful to Dr. W. Dansgaard whose earlier research on stable isotope variations has been extremely valuable and a constant source of stimulation.

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## INTRODUCTION

### Value of Microparticle Studies

The ice sheets of the world provide ideal locations for the investigation of changes over time in such climatic parameters as temperature and variations in the concentration, size distribution, and composition of microparticles. The two major ice sheets, Greenland and Antarctica, are unique in that they were completely remote areas until recently, in which were recorded annual variations of the atmospheric impurities.

As early as 1883, Nordenskiöld (1886) noted visible summer dust layers on the Greenland Ice Sheet, while in 1922, Wright and Priestley (1922) working in Antarctica noted visible silt layers on the coast of Robertson Bay which they interpreted to be annual, wind deposited features. Marshall (1959) was among the first to demonstrate the presence and stratigraphic variation of particles in the Greenland Ice Sheet. A cyclic variation in particle concentration within ice samples from Byrd Station, Antarctica was suggested by Marshall (1962) who proposed that annual layers might be detected from these cyclic variations. Thompson (1973) and Thompson and others (1975) suggested that the cyclic particle variations found in the Byrd Deep ice core are annual.

Hamilton and Langway (1967) compared the variations of the particle concentration and oxygen isotope data with depth in a 700-year-old ice samples from Camp Century, Greenland. Their results showed that spring snowfall contains a much greater abundance of particles per unit volume than does the snowfall from other times of the year.

The ice sheets of the world provide ideal locations to study the variations in microparticles with time. First, on the ice sheets annual snow accumulation averages from a few cm to over 100 cm, providing a much more expanded time record than is available in deep sea or lake bottom cores. Second, due to the development of ice core drilling techniques, it has been possible to recover ice cores to bedrock from Camp Century, Greenland (1966, 1387 meters) and Byrd Station, Antarctica (1968, 2164 meters). In addition, cores of intermediate depth have been retrieved from various sites in Antarctica, Greenland, Meighen Island, Devon Island, and Iceland. Cores to be taken in the future from East Antarctica may yield very long-term records of variations in the concentration, size distribution and composition of microparticles covering perhaps 200,000 years.

Recent theoretical studies (Pollack and others, 1976) and empirical evidence (Newell, 1970) indicated that variations in the concentration of particles in the atmosphere is an important component in the heat

balance of the earth-atmosphere system. Some investigators (McCormick and Ludwig, 1967; Bryson, 1968; Rasool and Schneider, 1971) attributed the decrease in mean air temperatures of the northern hemisphere since the 1940's to an increase in the atmospheric aerosol load. It is generally agreed that the effect of high altitude aerosols, such as stratospheric dust veils of volcanic origin, is theoretically one of surface cooling; however, the role of aerosols concentrated in the lower troposphere in the radiation budget is still open to speculation. Neumann and Cohen (1972) predicted net cooling as the end result, while Charlson and Pilat (1969) predicted net warming.

Pollack and others (1976) showed that an increase in the number of silicate and sulfuric acid particles of volcanic origin would result in an increase in the global albedo, and ultimately, in a reduction of world surface temperatures. They suggested that changes in the level of volcanic activity may be a significant factor in producing temperature changes over the last century and during portions of the last major glacial stage, The Wisconsin. Lamb (1970) and Mitchell (1975) illustrated the temporal relationship between volcanic activity and atmospheric temperature fluctuations in the northern hemisphere. Mitchell (1975) found a correlation between an increase in the quantity of volcanically-injected stratospheric material and a decrease in mean temperatures in the northern hemisphere for the period since 1880 A.D.

#### Reasons for Variations in Particle Concentration

The Antarctic continent is completely covered by a permanent ice sheet and is located in the polar easterlies while the central and southern regions of the Greenland Ice Sheet are situated in the westerlies. Greenland, unlike Antarctica, serves as a permanent obstacle to the mid-latitude low pressure systems approaching from the west, which consequently are often diverted to the north. Since Greenland and Antarctica have different geographical positions the mechanisms responsible for microparticle deposition are likely to be dissimilar.

For the Greenland ice sheet Putnins (1970) illustrated that the frequency of precipitation is fairly evenly distributed on an annual basis. However, for the coastal regions of Greenland the mean seasonal snow accumulation generally is greatest in April and least in August.

In Antarctica Rubin and Weyant (1965) found that the mean monthly weather conditions vary from year to year. In addition, Schwerdtfeger (1970) noted that no one season consistently exhibits the greatest abundance of snowfall, although cyclones do tend to be more intense in the winter. At present the seasonal variations in snowfall as well as the general circulation patterns over Antarctica are not well documented.

Particles are transported aerodynamically and hence, their distribution and subsequent deposition is related closely to the local and global atmospheric circulation regimes. Seasonal variations in circulation patterns are produced primarily by seasonal redistribution of the latitudinal radiation balance. Thus the circulation regime of the polar summer is not synonymous with that of the polar winter as each responds to thermal gradients of different magnitude and spatial distribution. Due to the variations in intensity and direction of the seasonal circulation regimes, the prevailing surface winds draw upon different source regions which in turn affects the abundance and nature of the particles aerodynamically transported and subsequently deposited on the ice sheets.

Hamilton (1969) noted that variations in particle deposition patterns are affected by seasonal changes in air temperature, precipitation and saturation mixing ratios. To serve as condensation nuclei smaller particles (radii  $< 10^{-2} \mu\text{m}$ ) require supersaturations in excess of 200% (Mason, 1962) while larger particles ( $0.1 \mu\text{m} < \text{radii} < 1.0 \mu\text{m}$ ) readily serve as condensation nuclei at much lower saturations. The saturation of the air is partially dependent upon temperature. Sood and Jackson (1970) determined the elemental scavenging efficiency of snow and ice crystals for particles of  $0.5 \mu\text{m}$  to  $2.0 \mu\text{m}$  radius and found it to be four times greater than that of raindrops of the same circumscribed diameter. Scavenging also is dependent upon a sufficient abundance of particles of different sizes. Thus, temperature, precipitation and saturation interact and, along with the concentration of particles in the atmosphere, determine the particle concentration found in polar ice sheets.

Aside from the seasonal changes in particle concentration over the polar regions, the frequency and intensity of volcanic eruptions and continental aridity produce temporal fluctuations in the number of particles available for deposition. If the volcanic or arid region sources are not directly adjacent to the polar ice sheet under consideration, then the deposition of the particles at that location probably will be cyclic as a result of seasonal changes in both the local and global circulation patterns.

It is possible that on smaller ice caps the deposition of particles is temporally more irregular due to the preponderance of sources. On the other hand, for some ice caps such as the Quelccaya Ice Cap in Peru, the deposition of particles may vary on an annual basis due to the operation of factors other than those discussed above (Thompson and Dansgaard, 1975). The climatic regime of the central Peruvian Andes exhibits marked seasonal variations. During the two months of the dry season the glacier surface is exposed to direct solar radiation while during the ten months of the rainy season the ice cap receives its annual snowfall under heavy cloud cover.

Generally, it appears reasonable to conclude that ice caps located in areas with marked seasonal variations in climatic regime possess the greatest potential for producing meaningful microparticle stratigraphic records. It is necessary to bear in mind that reworking of fallen snow by the wind, volcanic eruptions in the vicinity or anomolous weather conditions can disrupt the stratigraphic record of any snow field.

This investigation was conducted to determine the nature of variations in particle concentrations, size distributions and compositions in ice cores. Microparticle analyses have been conducted for surface cores as well as for 23 core sections (27.4 meters of ice) from the Byrd core and 30 core sections (14.15 meters of ice) from the Camp Century core.

The laboratory techniques employed in these analyses are presented in the second chapter. Chapter III discusses the analyses of surface firn cores to establish the existence of an annual cycle in particle concentration variations. This cycle is utilized to establish chronologies for the Byrd and Camp Century deep ice cores which are presented in Chapter IV. In Chapter V the microparticle elemental compositions and morphologies are examined to determine changes in particle sources with time. Chapter IV compares the microparticle concentrations, compositions and morphologies with oxygen isotope variations over millennial time intervals. The relationships between volcanic activity and climatic change are discussed in Chapter VII.

## LABORATORY TECHNIQUES

### Particle Concentration and Size Distribution

The fundamental procedures for analyzing samples from ice cores for particle concentration were established by Marshall (1962) and greatly refined by Bader and others (1965). Taylor and Gliozzi (1964) modified the original techniques in the clean room facilities in the Institute of Polar Studies at The Ohio State University. Further modifications in these facilities and techniques were developed by Hamilton (1967, 1969) and Thompson (1973, 1975a). As of June 1975 the speed and reliability of microparticle examination in glaciological samples was greatly enhanced by the establishment of a new clean room and the upgrading of equipment (Figures 1 and 2).

The laboratory is designed to determine microparticle variations in snow and ice samples. The equipment includes two multichannel particle counters (Coulter Model TA II counter, with X-Y recorders); a Coulter 550 air-contamination counter system that continuously monitors room air; a Milliport Milli-Q3 system which purifies tap water to general laboratory requirements; and a Milli-Q2, which produces reagent grade water. The combined use of the Milli-Q2 and -Q3 provide a large quantity of consistently high-quality deionized and filtered water (henceforth referred to as DFW). The Milli-Q2 water system removes insoluble particles of diameter greater than 0.22  $\mu\text{m}$ .

When lab workers wear proper clean-room clothing (i.e. full length smock, hat, and shoe covers) the laboratory maintains class 100 conditions for measuring particle variations in ice cores. With this equipment we have been able to analyze as many as 130 samples per day, the equivalent of  $\sim 5$  meters of ice core.

### Sample Preparation

Cores analyzed in the laboratory are either ice or firn. Firn samples taken with a SIPRE auger are approximately 7 cm in diameter; sections from the deep ice cores are approximately quarter-core segments cut from the parent cores which are 10.8 cm in diameter. Because the physical properties of ice and firn are dissimilar, different laboratory cleaning procedures are required.

The initial step in sample preparation is to make a detailed sketch of the structure of the intact core sections including notation of the top-to-bottom orientation. All visible fractures and the number and orientation of air bubbles are included in this sketch. Stratigraphic features such as horizontal color variations and dust bands in the ice sections are recorded. Particular attention is given to the dust bands;

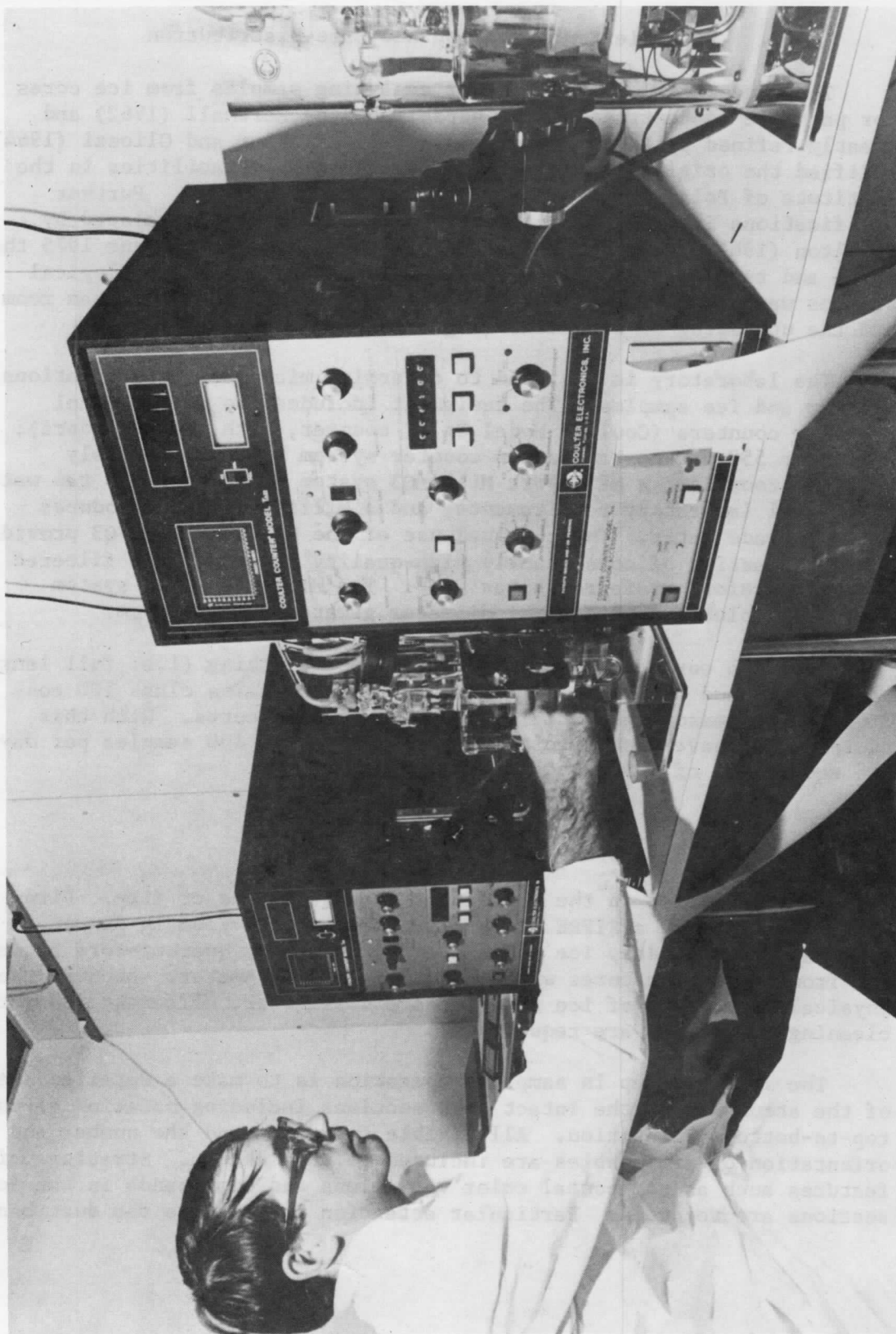


Figure 1. Two Coulter Model TA II counter used in microparticle analysis.



Fig. 2. Sample preparation in class 100 clean room.

their exact location and the number of degrees by which the bands deviate from the horizontal are specifically noted.

With ice sections the second step is to cut the section into samples with a coarse-toothed band saw. Discretion is required when choosing the sample size as it is important that the sample be small enough to insure that changes over small distances in particle variations will not be lost. However, from a practical point of view, the samples should be as large as possible to provide analyses over longer sections of the core. In this investigation sample sizes have ranged from 0.7 to 2.5 cm. In the sections containing visible stratigraphic features (i.e., dust bands) the samples are usually cut parallel to the features assuming that the features were originally horizontal.

In the laboratory ice samples are rinsed thoroughly in DFW to remove surface contamination. During rinsing the samples are handled only with glass-tipped tongs and approximately 2 mm of ice is melted from all sides of the ice sample to insure removal of possible contamination.

Firn cores require special handling because their porous nature renders them more susceptible to contamination at all stages. Firn cores are cut into 4 to 8 cm samples (sample size being largely a function of the annual accumulation at a given site) and transported to the clean room, where a 4.5 cm diameter plug is drilled from the center of each sample with a polyethylene hand corer. This procedure provides a sample which has not been in contact with the core driller or the shipping container.

Some problems were encountered while rinsing samples from the Byrd core between 400 and 900 meters depth where the ice is fragile and badly fractured (Gow and others, 1968). The fractures allow the drilling fluid, in this case ethylene glycol, to penetrate and thus to contaminate the ice sample. With samples from this section removal of all contamination was impossible; particle counts for these samples should be viewed with caution. This is illustrated by the core sections from 565 and 953 meters shown in Appendix A. The data from the fractured sections are not used in subsequent discussions of the Byrd core micro-particle analyses.

After rinsing, the ice sample is allowed to drip and is then allowed to melt in bottom-weighted polyethylene containers which have been rinsed with DFW. Samples initially 0.7 and 2.5 cm thick yield 7 and 25 ml of liquid, respectively. For use in the Coulter counter the melted sample is made conducting by converting it into a 2 percent NaCl solution by adding the appropriate amount of filtered concentrated NaCl solution. The sample is then stirred with a precleaned dropper immediately prior to analysis in the particle counter.

The melted ice core samples were analyzed in random order and only one half of each ice core section was analyzed at a time. The second half of each section was analyzed approximately two months after the first half in order that any apparent particle variation trends due to laboratory procedures might be detected.

### Contamination

The elimination of all possible contamination is a major objective in microparticle studies. Contamination can be introduced during coring, handling, shipping, and cutting of samples. Owing to their high porosity, firn cores are especially difficult to handle. In the field the collection of firn cores must be conducted upwind from any possible contaminant source.

Hamilton (1969) enumerated a number of possible sources of contamination in the laboratory. Among these were impurities from the DFW supply, the sample containers, the concentrated NaCl solution, and insufficient cleaning of the ice samples during rinsing, particles in the air of the laboratory and from particulate carry over on the aperture tube. Of the above, the contamination from the DFW supply and the concentrated NaCl solution are the most serious. The contamination from microorganisms may be controlled, without affecting the analyses, by using small amounts of sodium azide to kill bacteria. To insure that contamination from the concentrated NaCl and DFW supply is minimized, a count is taken at the beginning of each day on a blank sample made from the DFW and the concentrated NaCl. When the class 100 clean room was established in June 1975 the contamination potential from the DFW and particles in room air was greatly reduced. Federal standard 209b defines a class 100 laminar flow clean room as one in which there are 100 or less particles per cubic foot of air with diameters greater than 0.5  $\mu\text{m}$ .

### Particle Counters

We have used both the Coulter Model T and the Model TA II counters in our analyses. The Model T is greatly improved over the earlier Coulter models in that it electrically separates particles into fifteen size ranges. In this investigation threshold values were set to range from less than 0.52  $\mu\text{m}$  diameter for channel 14 to greater than 14.77  $\mu\text{m}$  for channel 0. The Model T is adjusted electrically so that the average particle volume has an incremental factor of 2 for adjacent channels. The data from counts for samples are then available in 4 basic forms: population differential, population cumulative, volume differential and volume cumulative. The instrument also clocks the time for the individual counts in all four modes.

In the class 100 clean room there are two newer Coulter counters, both Model TA II. These counters electronically separate particle sizes into 16 ranges between 0.4 and 18  $\mu\text{m}$  in diameter. All particle concentration and size distribution data presented in this manuscript are restricted to particle diameters greater than 0.5  $\mu\text{m}$  since the counts below this diameter are subject to inaccuracies due to electronic noise.

All Coulter counters operate on the Coulter principal which allows determination of the concentration and size distribution of particles suspended in an electrolyte by forcing the sample to flow through a small aperture and monitoring an electrical current which also passes through the aperture. Current passes between two electrodes submerged in the liquid sample, one inside and one outside the aperture tube.

The size range of particles to be counted dictates the size of aperture employed; in this investigation a 30  $\mu\text{m}$  aperture tube was used. A known volume of the sample, usually 500  $\mu\text{l}$ , is pulled through the aperture by a pressure difference established by means of a mercury manometer. As the particles suspended in the sample pass through the aperture they displace the electrolyte, momentarily increasing the resistance in the path of the current. This produces a current pulse of short duration, with a magnitude proportional to the particle cross-sectional area. The current pulses then are electronically scaled and counted. Bader and others (1965), Hamilton (1969) and the Coulter Counter Model TA II Manual (1975) provide additional technical information on the method.

According to the Coulter theory, the particle resistivity has very little effect on the response, unless that resistivity is very close to that of the electrolyte. The manufacturer has shown that if the particle conductivity changes from one millionth to one hundredth of that electrolyte, there is less than 1 percent change in the response.

If two particles traverse the aperture nearly simultaneously only one may be counted. Mattern and others (1957) have shown that the proportional coincident passages of particles are a function of the particle concentration. They plotted the instrument count against the true count for whole blood and found little difference in the two counts at low concentrations. Since the average particle concentration in the ice samples analyzed is about 100 ppm, counting errors due to the coincidence effect may be assumed negligible.

#### Laboratory Procedures

A warming time of at least 30 minutes is used for both the Model T and TA II counters. The instrument is then calibrated using monosized (1.305  $\mu\text{m}$  diameter) polystyrene latex spheres in a 2.0 percent electrolyte solution.

By using two TA II particle counters an internal check of the counting accuracy can be made periodically by counting the same sample on each machine and comparing the results. Once the counters have been calibrated a check of the calibration is made by using a standard of different diameter to determine if the peak concentration occurs at the expected threshold. After the calibration, the aperture tube is flushed with DFW to remove possible contamination by the calibration spheres. Concentration and size distribution analyses are then performed with the counters, after which the remaining sample is filtered through a 0.45  $\mu\text{m}$  Millipore filter to retain the particles for analysis of composition.

### Particle Composition

An important aspect of a particle analysis is the composition of the particles. Several methods to determine particle composition have been employed in earlier studies. Windom (1969) used X-ray diffraction on dust found in various permanent snowfields from sites all over the world. He found that the particles in samples taken from a shaft at New Byrd Station, Antarctica, included small amounts of quartz and perhaps feldspar. Wright and others (1963), Hodge and others (1964), Langway (1963), Langway and Marvin (1964) and Langway (1970) chemically analyzed 115 spherules from Site 2 and from Camp Century, Greenland, using an electron probe microanalyzer. They found that the elements Fe, Si, and Mn occurred most often, while Al, K, Ca and Ti occurred in lesser quantities. Langway (1970) used microscopic techniques and X-ray diffraction to determine the composition of cosmic dust in a deep ice core from Greenland. In that study metallic spherules, which are considered extraterrestrial, were found to consist primarily of magnetite with some hematite and wüstite. In the present study a light microscope, scanning electron microscope (SEM), and energy dispersive X-ray analysis system (EDS) have been used in analyzing the particles.

### Sample Preparation

The particles which are retained upon the 0.45  $\mu\text{m}$  filters are stored in sterile petri dishes in the clean room. All preparations are conducted in the clean room, and all samples are handled using only utensils cleaned in an ultra-sonic cleaner. For this initial study one filter from each of the sections of the two cores listed in Table 4 (Appendix C) and Table 5 (Appendix D) has been examined. Appendix C illustrates the elemental composition and morphology of the microparticles from the Byrd Station deep ice core while Appendix D presents the elemental composition and morphology of the microparticles from the Camp Century deep ice core.

The selection of the filters for study is difficult as the same number of samples have not been filtered upon each filter. The data obtained by the Coulter counter provide a frequency analysis according

to the particulate diameter. The criterion for sample filter selection is established to provide a sufficient number of particles to warrant the time and expense of analysis (at least 350 particles with radii greater than  $2.0\text{ }\mu\text{m}$ ), but not so many particles as to hinder visual study. Rarely are filters with more than 15 samples selected, nor are filters with only one sample out of a core section appropriate. This is done to reduce the chance of obtaining an unrepresentative sample for each core section.

The filters are cut into fifths and one of these segments is affixed to an aluminum stub with a silver colloid cement for subsequent examination and analysis by light and scanning electron microscope. Each aluminum stub has a plastic cover to prevent contamination, and the samples are stored in the clean room. Utmost caution is exercised during the entire procedure to minimize contamination.

The study of microparticles from these ice cores is quite difficult, due to their large concentration and extremely small size. Almost all the particles have diameters less than  $30.0\text{ }\mu\text{m}$ , and their distribution mode centers around  $0.5\text{ }\mu\text{m}$  diameter. The scanning electron microscope and the Ortec X-ray energy dispersive system (EDS) allow the smallest particles to be photographed and analyzed for elemental composition, but the great quantity of particles (generally more than  $2 \times 10^5$ ) on each filter segment makes it impossible to study all the particles individually. For these reasons we have adopted the following procedure.

The filters are initially studied by a Leitz-Wetzler Orthoplan light microscope with a maximum magnification of 750 diameters. The microscope is equipped with a fully automatic Orthomat camera. A particle classification scheme has been devised for this study (Table 1).

The particle population for each filter is sampled to determine the frequency of occurrence of each particle type within each section of the core. The sampling procedure used is adopted from the pollen analysis technique suggested by Faegri and Inversen (1964). The specimen is placed under the microscope and counting and classification proceeds along a horizontal traverse. The preparation is then shifted one and one-half diameters of the field of view perpendicular to the direction of the initial traverse and moved horizontally again along another counting traverse.

To determine the number of particles which should comprise a sample population many consecutive samples of various sizes are counted for the same filter population. The frequency of occurrence is calculated for each particle type in each of the trial samples until the relative frequencies become nearly constant for three consecutive samples. A sample population of 350 particles gives repeatable results. A limiting magnification of 750x for counting and classification dictates a particulate detection limit of  $4.0\text{ }\mu\text{m}$  diameter.

TABLE 1

Particle classification scheme based upon visual properties

- 
- I. Perfect spherules
    - A. Shiny metallic luster
    - B. Clear with smooth surface
  - II. Imperfect spherules - more round than angular
    - A. Shiny metallic luster
    - B. Clear with smooth surface
  - III. Irregular particles - angular
    - A. Translucent with vitreous luster
      - 1. Orange
      - 2. Light red
      - 3. Dark red
      - 4. Smoky-gray
      - 5. Yellow
      - 6. Brown
    - B. Opaque with vitreous luster
      - 1. Black - flaky
      - 2. Black - non-flaky
    - C. Opaque with metallic luster
      - 1. Black - flaky
      - 2. Black - non-flaky
    - D. Clear without inclusions
    - E. Clear with inclusions
-

The scanning electron microscope produces two dimensional images with a large depth of field (about 300  $\mu\text{m}$ ) by scanning the specimen with a fine electron beam synchronized with the electron beam of a cathode ray picture tube. In addition to the secondary electrons producing the CRT image, the particle emits X-rays which are characteristic of the atoms in the particle under analysis. The SEM is equipped with an X-ray energy dispersive system (EDS) containing a crystal spectrometer which collects the characteristic X-rays for all elements with atomic number greater than 10. The X-ray impulses are passed to a solid-state detector with a pulse-height analyzer which produces an energy spectrum which is, therefore, representative of the various types and abundance of atoms within the sample. This method of X-ray emission chemical analysis was described first by Moseley (1913), but this microprobe technique has been refined greatly by the development of a narrow or small-diameter probing beam and by the new principles developed to obtain quantitative results.

To a first approximation, the concentration of an element (A) in a sample may be obtained by comparing the intensity of the characteristic spectral line of element (A) emitted from the unknown sample under specific instrumental conditions with the intensity of the corresponding characteristic line emitted from a sample of pure element (A) under identical instrumental conditions (Castaing, 1963). In quantitative microprobe analysis the characteristic line intensities from both the sample and the standard must be corrected for physical processes which occur within the specimen during the analysis. The three primary correction factors which generally are applied to all X-ray spectra account for absorption, fluorescence and for atomic number effects. The last includes losses of intensity due to backscatter and to ionization-penetration (Reed, 1973).

To apply these corrections to the X-ray energy dispersive data, the microprobe computer program, MAGIC IV, is employed. MAGIC, which is an acronym for microprobe analysis general intensity corrections, was first described by Colby (1969) and was later updated to handle standards containing up to five elements (Colby, 1971). Using The Ohio State University IBM-370, MAGIC IV, and a set of pure elemental standards, the EDS spectral data for each sample are converted into the weight percentages of elements present in the particle under analysis. These results for particles from the Byrd and Camp Century cores are listed in Appendices C and D.

A total of 5000 particles from each core were classified visually, using the light microscope sampling and classification technique described earlier. A total of 105 particles from 14 sections of the Byrd core and 97 particles from 13 sections of the Camp Century core were analyzed by the energy dispersive technique. The color light micrographs taken earlier were used to locate particles of a specific type for SEM photography and X-ray analysis. Approximately 8 to 10 particles upon each filter section and at least one of each of the most abundant particle types were examined and probed.

The data for each particle were obtained from the X-ray analyzer by teletype output of the selected regions of interest (ROI) rather than by a complete printing of all 1024 collection channels. The same technique was employed for each spectrum and all data reduction methods were applied consistently to all data sets. The results of the above analyses are discussed in Chapter IV and listed in Appendices C and D.

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## TYPES OF PARTICLES AND THEIR SEASONAL VARIATION

### Particle Types

Particles deposited on the ice sheets can be separated into water soluble and non-soluble particles. Microparticles deposited on the ice sheets, either as impurities within the snow or as dry-air fallout, have four primary sources: cosmic, continental (eolian), marine and volcanic. In addition during the last century, anthropogenic activities have introduced increasing amounts of particulate material into the atmosphere. Because the ice samples are melted before analysis and soluble particles, such as marine salts pass into solution, this discussion is restricted to non-soluble particles.

In many cases, non-soluble particles from the above sources may be distinguished on the bases of morphology and elemental composition. This point is illustrated by Plates I and II, which compare six particles, two from each possible source and which list the elements present in order of decreasing abundance. Nordenskiöld (1886) called mineral particles deposited on ice sheets cryoconites and concluded that the non-soluble particles came from the three main sources illustrated in Plates I and II. The two types of cosmic particles shown in Plate II also differ in origin. The first spherule is rich in silicon and iron and, therefore, may be the ablation product of a stony meteorite, while the second spherule is iron-rich suggesting that it is the ablation product of an iron meteorite (Wright and others, 1963).

The size distribution of suspended particulate material is a function of the nature of the distance from contribution sources, wind velocities and humidity, all of which determine the rates of sedimentation, coagulation and precipitation scavenging (Junge, 1963). Figure 3 displays the typical logarithmic size distribution of both continental and maritime particulate material (U. S. Air Force, 1960). The shaded portion of the diagram indicates the particle size range measured in ice core samples by the Coulter counter technique.

### Variations of Particle Concentrations

In following sections surface ice cores are examined to establish a cyclic variation in particle deposition. Because the deposition of particles is cyclic the particle variations can be used as a time record. All particle profiles are the results of analyses conducted in this investigation. Oxygen isotope profiles are presented for comparison with the particle profiles. The oxygen isotope values are given as the  $\delta^{18}\text{O}$  ratio, where;

# PLATE I

## Major Types of Non-Soluble Particles in Deep Ice Cores

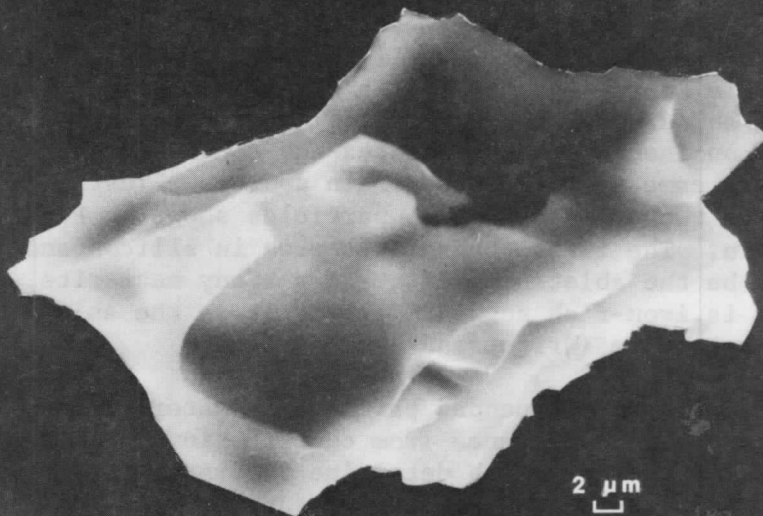


B-875 Mag. 3000

### Elements

Fe  
Al  
Si

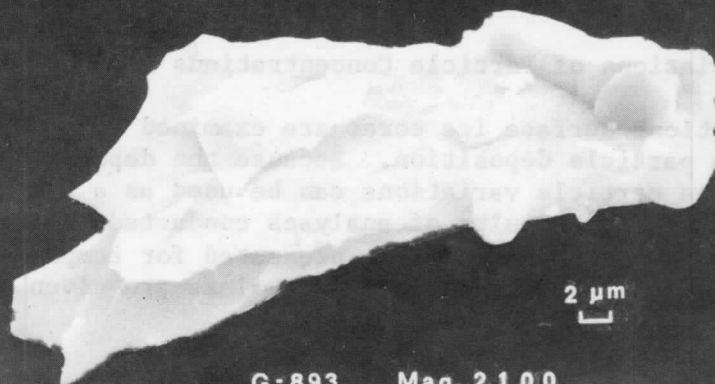
VOLCANIC



B-875 Mag. 2900

### Elements

Si  
Al  
K  
Fe  
Ca



G-893 Mag. 2100

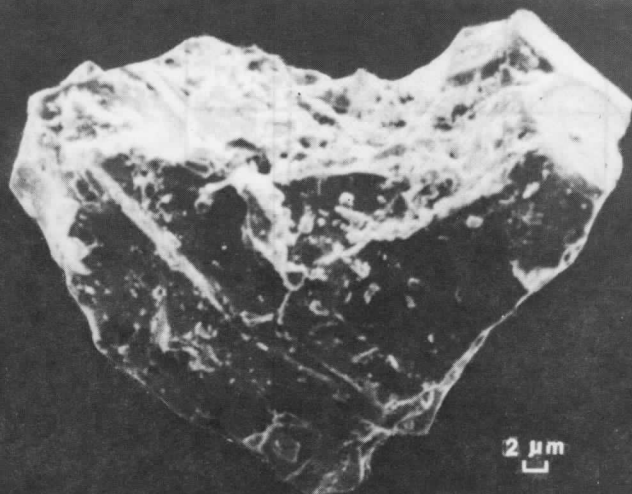
### Elements

Si  
Mg  
Al  
Fe  
K  
Ti

CONTINENTAL (clay)

# PLATE II

## Major Types of Non-Soluble Particles in Deep Ice Cores



2  $\mu$ m

CONTINENTAL

B-74 Mag. 2100

Elements

Si

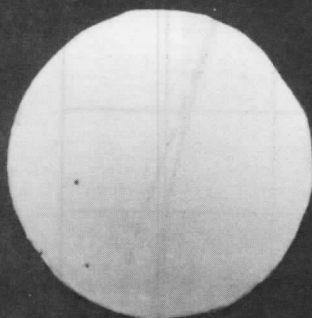
Al

Fe

K

Ca

Ti



2  $\mu$ m

B-331 Mag. 6000

Elements

Si

Fe

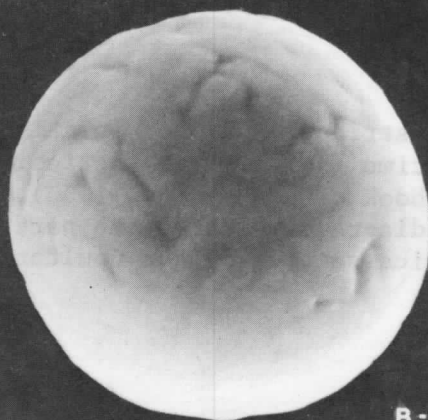
Al

K

Ti

Ca

COSMIC



2  $\mu$ m

B-751 Mag. 5500

Elements

Fe

Al

Si

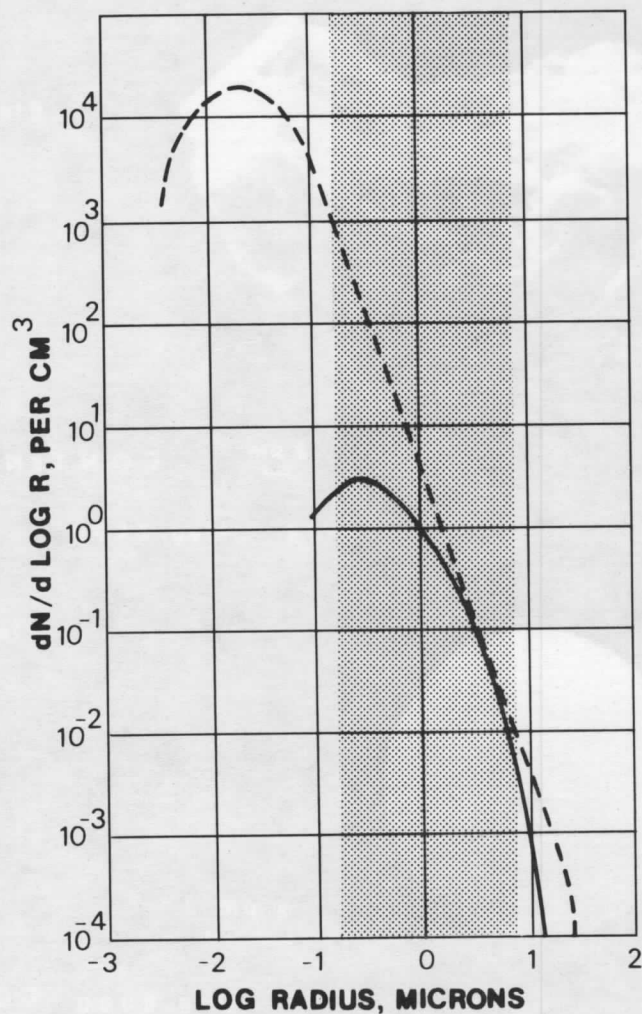


Figure 3. The logarithmic size distribution of both continental (dashed line) and maritime (solid line) particulate material. (USAF, Handbook of Geophysics, 1960). The shaded portion of the diagram indicates the particle range measured in the ice samples by the Coulter Model TA II counter.

$$\delta = \left[ \frac{(\text{H}_2\text{O}^{18} / \text{H}_2\text{O}^{16})_s}{(\text{H}_2\text{O}^{18} / \text{H}_2\text{O}^{16})_{\text{SMOW}}} - 1 \right] \times 10^3 \quad \text{o/oo}$$

where (s) indicates sample and (SMOW) indicates Standard Mean Ocean Water (Craig, 1961).

### Antarctic Peninsula

Microparticle concentration and size distribution have been measured for firm cores from the Antarctic Peninsula which were collected by the British Antarctic Survey from the Graham Land plateau (66°00'W; 67°32'S). Two cores, collected from sites five meters apart, encompass the depth interval of approximately 3 to 8 meters below the 1974-75 snow surface. The results of the analyses of these cores are presented in Figs. 4 and 5.

The particle profile in Fig. 4 is the more reliable of the two, as the firm core illustrated in Fig. 5 experienced varying degrees of melting during transport. Those sections exhibiting melting are shown on the right of each profile. The consistent background levels in the total particle count in Fig. 4 further attests the lack of melting and resultant contamination. Note that for both profiles approximately 20 samples represent each year of accumulation and that the concentration of particles is presented as the total number greater than 0.62  $\mu\text{m}$  diameter per 500  $\mu\text{l}$  of sample.

From the particle profiles for this depth interval a mean annual thickness of 0.82 m and 0.84 m is calculated for profiles in Figs. 4 and 5, respectively. The variation in the two calculated annual accumulation rates probably results from differences in snow redistribution by wind following deposition.

The above interpretation of the microparticle variations suggest that the mean annual layer thickness of snow in this area of the Antarctic Peninsula averages 0.82 m. Furthermore, these data suggest that the microparticle dating technique is valid for this relatively warm snow area and very probably could be used for determining annual layers in ice from deeper in the glacier.

### Marie Byrd Land

During the 1973-74 field season surface samples were collected from points along the flow line upglacier from Byrd Station, Antarctica. These cores were analyzed at The Ohio State University for

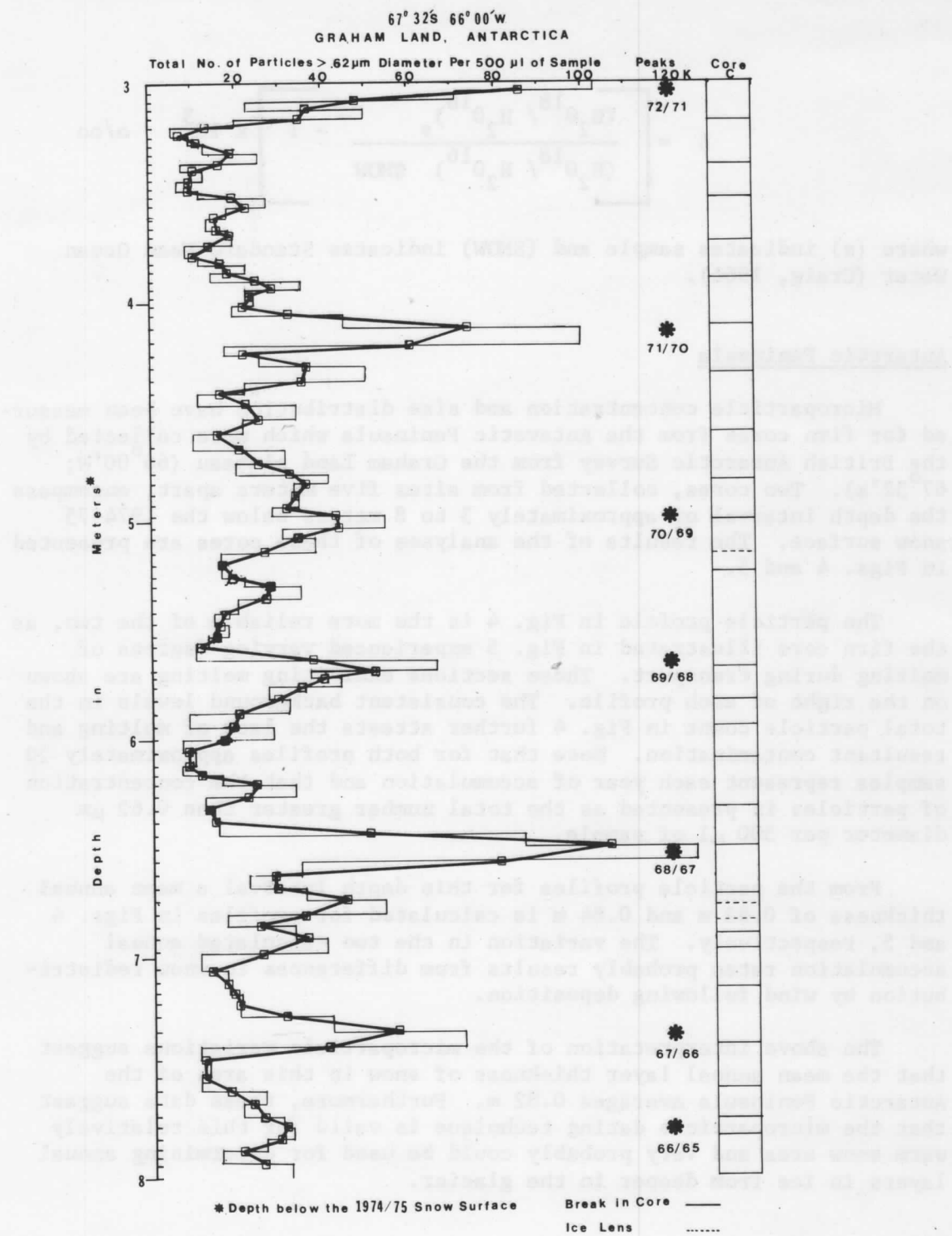


Figure 4. Variations in total microparticle concentration in a 5 meter core from the Antarctic Peninsula. The core stratigraphy is illustrated on the right of the diagram.

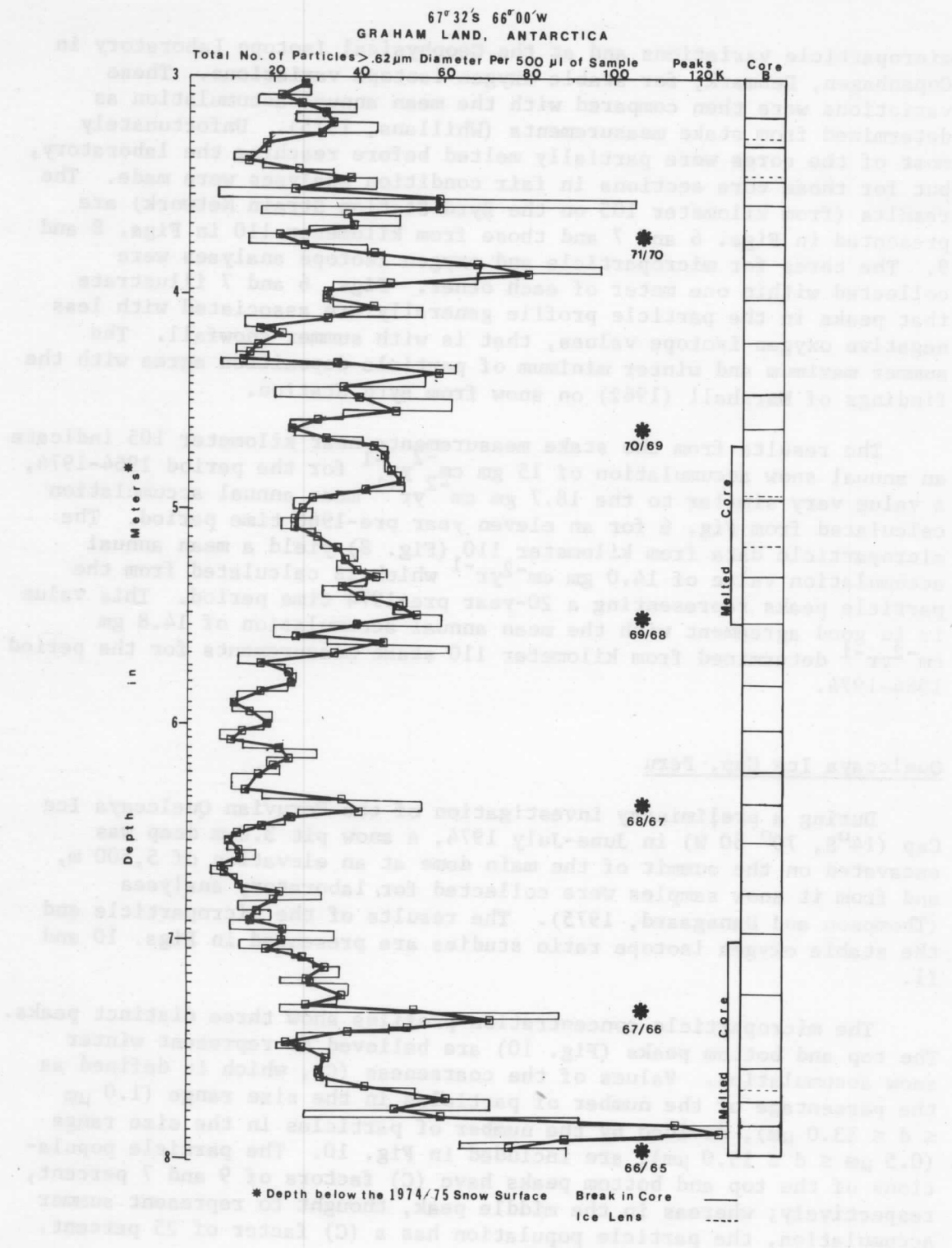


Figure 5. Variations in total microparticle concentration in a 5 meter horizontal distance from the core site for Figure 4. The core stratigraphy is illustrated on the right of the diagram.

microparticle variations and at the Geophysical Isotope Laboratory in Copenhagen, Denmark, for stable oxygen isotope variations. These variations were then compared with the mean annual accumulation as determined from stake measurements (Whillans, 1975). Unfortunately most of the cores were partially melted before reaching the laboratory, but for those core sections in fair condition analyses were made. The results (from kilometer 105 on the Byrd Station Strain Network) are presented in Figs. 6 and 7 and those from kilometer 110 in Figs. 8 and 9. The cores for microparticle and oxygen isotope analyses were collected within one meter of each other. Figs. 6 and 7 illustrate that peaks in the particle profile generally are associated with less negative oxygen isotope values, that is with summer snowfall. The summer maximum and winter minimum of particle deposition agree with the findings of Marshall (1962) on snow from Byrd Station.

The results from the stake measurements near kilometer 105 indicate an annual snow accumulation of  $15 \text{ gm cm}^{-2} \text{ yr}^{-1}$  for the period 1964-1974, a value very similar to the  $18.7 \text{ gm cm}^{-2} \text{ yr}^{-1}$  mean annual accumulation calculated from Fig. 6 for an eleven year pre-1966 time period. The microparticle data from kilometer 110 (Fig. 8) yield a mean annual accumulation value of  $14.0 \text{ gm cm}^{-2} \text{ yr}^{-1}$  which is calculated from the particle peaks representing a 20-year pre-1974 time period. This value is in good agreement with the mean annual accumulation of  $14.8 \text{ gm cm}^{-2} \text{ yr}^{-1}$  determined from kilometer 110 stake measurements for the period 1964-1974.

#### Quelccaya Ice Cap, Peru

During a preliminary investigation of the Peruvian Quelccaya Ice Cap ( $14^{\circ}\text{S}$ ,  $70^{\circ} 50' \text{W}$ ) in June-July 1974, a snow pit 3.8 m deep was excavated on the summit of the main dome at an elevation of 5,500 m, and from it snow samples were collected for laboratory analyses (Thompson and Dansgaard, 1975). The results of the microparticle and the stable oxygen isotope ratio studies are presented in Figs. 10 and 11.

The microparticle concentration profiles show three distinct peaks. The top and bottom peaks (Fig. 10) are believed to represent winter snow accumulation. Values of the coarseness (C), which is defined as the percentage of the number of particles in the size range ( $1.0 \mu\text{m} \leq d \leq 13.0 \mu\text{m}$ ), divided by the number of particles in the size range ( $0.5 \mu\text{m} \leq d \leq 13.0 \mu\text{m}$ ), are included in Fig. 10. The particle populations of the top and bottom peaks have (C) factors of 9 and 7 percent, respectively; whereas in the middle peak, thought to represent summer accumulation, the particle population has a (C) factor of 25 percent.

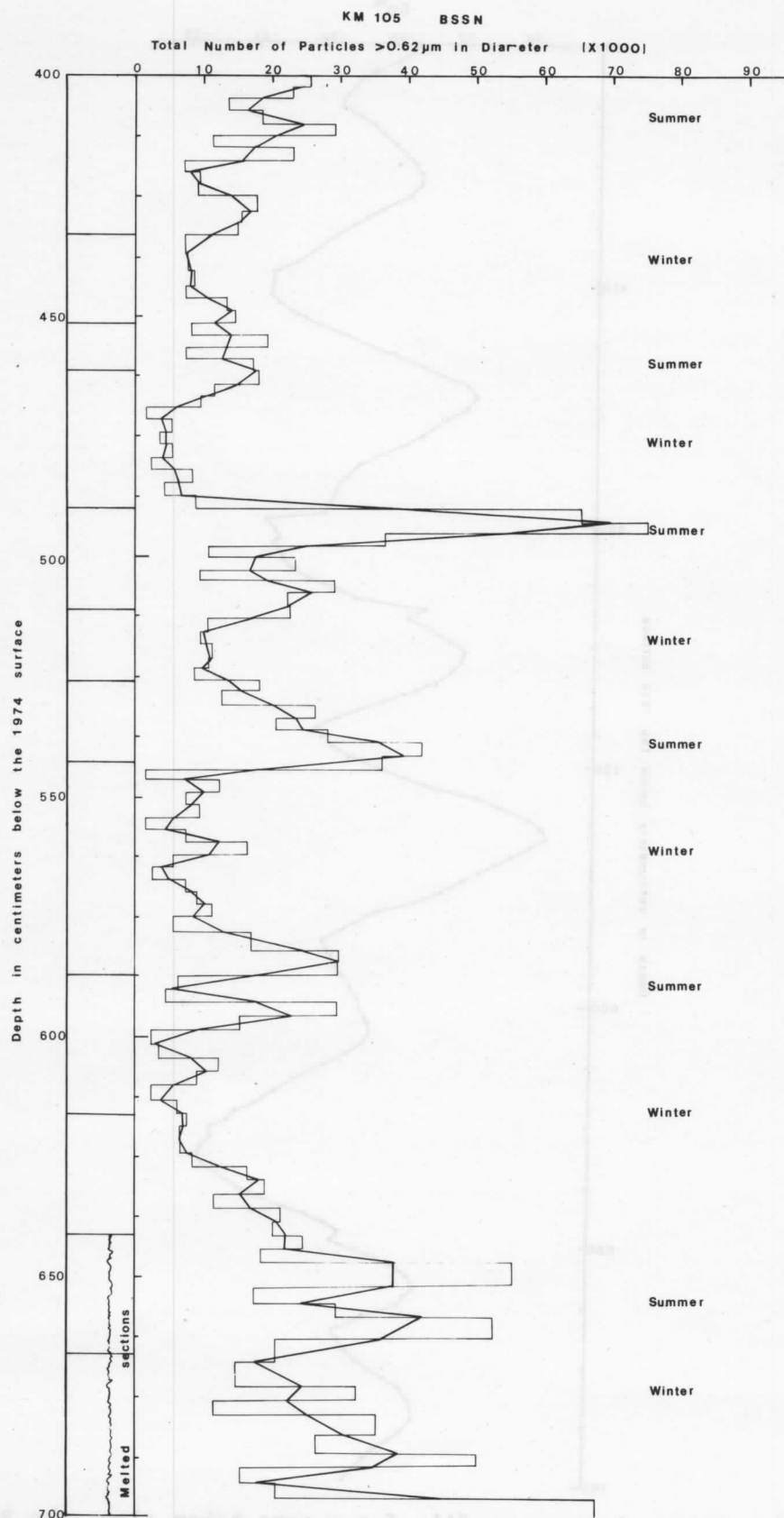


Figure 6. Variations in total microparticle concentration in a 3 meter core section taken along the Byrd Station Strain Network, 105 kilometers from the ice divide in West Antarctica. The summer/winter interpretation comes from comparing the particle profile with the oxygen isotope profile (Fig. 7) which was obtained for the same depth interval.

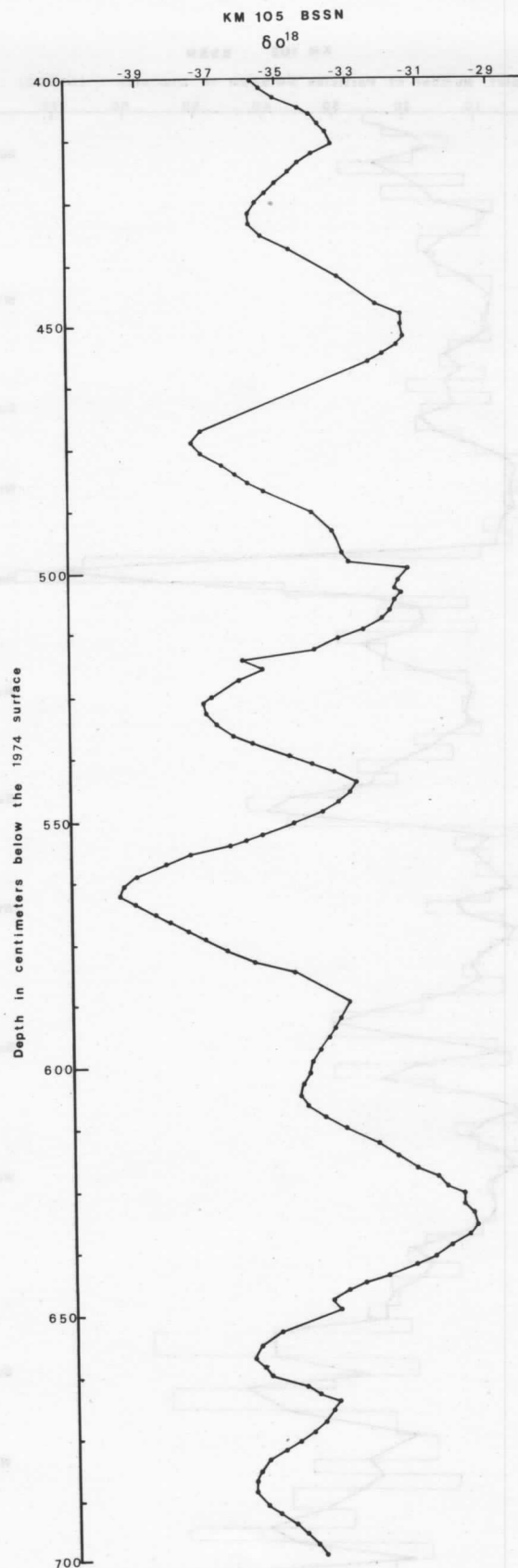


Figure 7. Oxygen isotope profile for a core taken along the Byrd Station Strain Network, 105 kilometers from the ice divide in West Antarctica. The depth interval is the same as that for the microparticle profile in Fig. 6. This core site is separated by less than 1 meter horizontal distance from the core site of Fig. 6. (Oxygen isotope data provided by I. Whillans and W. Dansgaard).

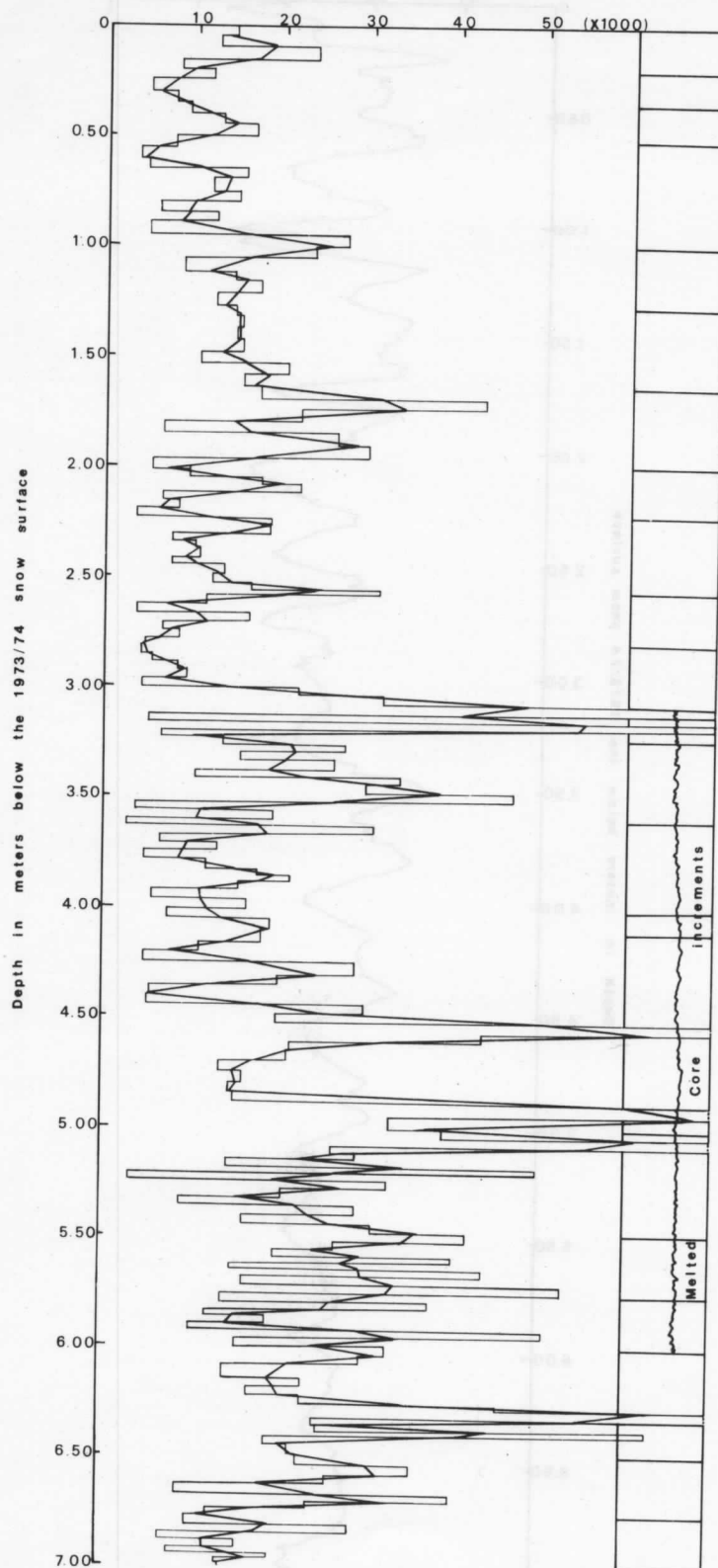
Total Number of Particles  $> 0.62 \mu\text{m}$  in Diameter

Figure 8. Variations in total microparticle concentration in a 7 meter surface core taken along the Byrd Station Strain Network, 110 kilometers from the ice divide in West Antarctica. Some sections of this core underwent partial melting in transport to the laboratory. These sections are labelled in the core stratigraphy on the right.

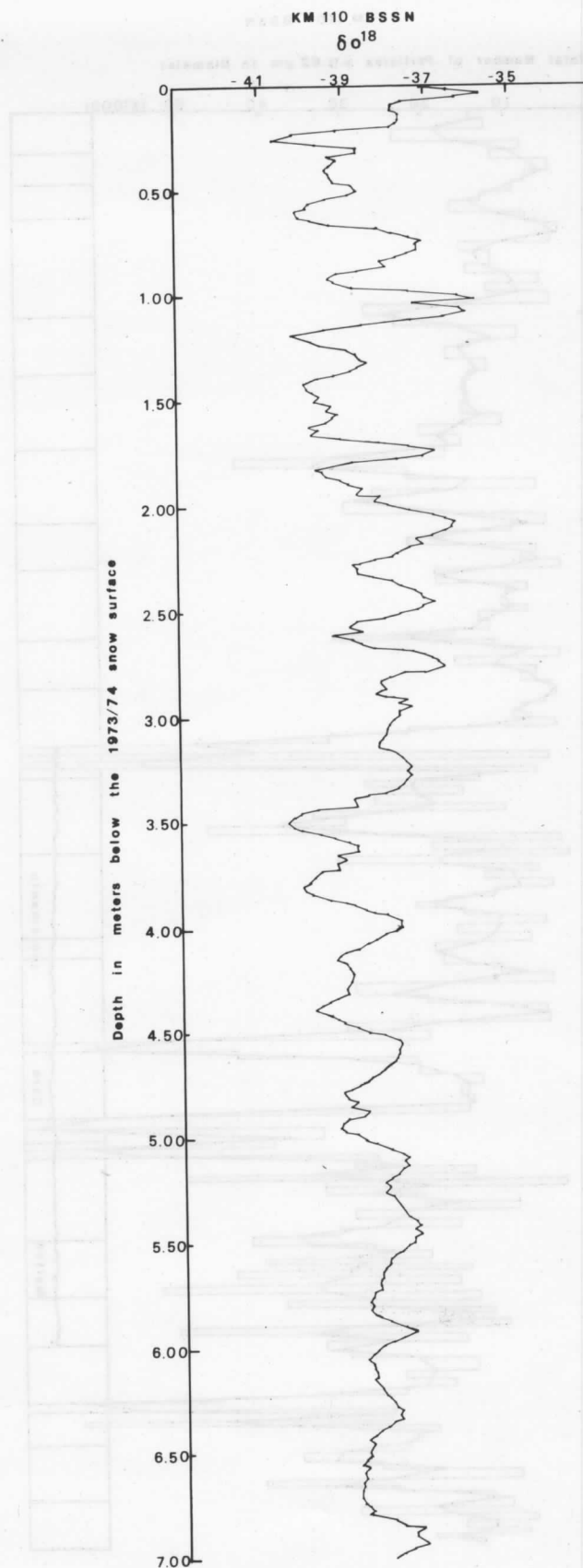
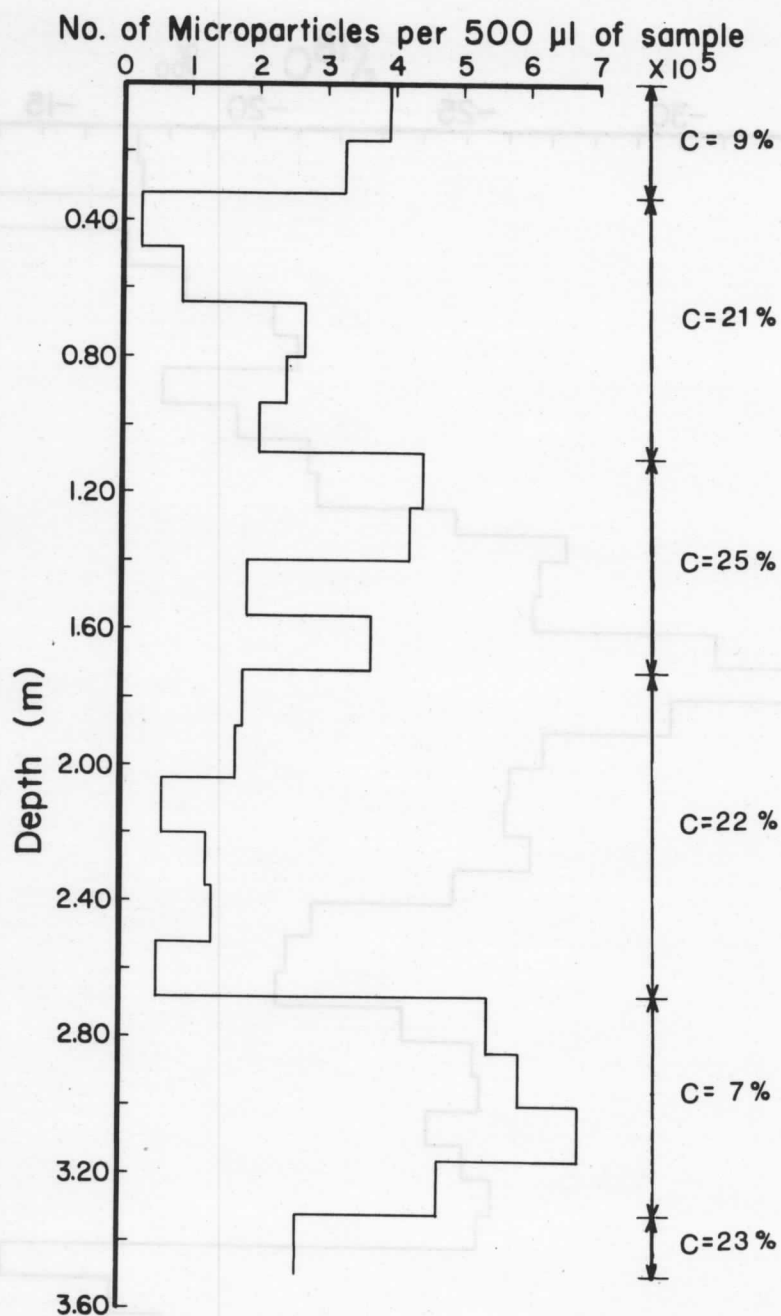


Figure 9. Oxygen isotope profile for a surface core taken along the Byrd Station Strain Network, 110 kilometers from the ice divide in West Antarctica. This core site is separated by less than 1 meter horizontal distance from the core site for Fig. 8. (Oxygen isotope data provided by I. Whillans and W. Dansgaard).



$$\%C = \frac{\text{No. Particles } (1.0 \mu\text{m} \leq d \leq 13.0 \mu\text{m})}{\text{No. Particles } (0.5 \mu\text{m} \leq d \leq 13.0 \mu\text{m})} \times 100$$

Figure 10. Results of the microparticle analyses of snow samples from the Quelccaya Ice Cap, Peru. The microparticle concentration profile shows three distinct peaks. In the top and bottom peaks representing snow which is believed to be winter accumulation, the particle populations have coarse-ness factors (C) of 9 and 7 percent, respectively; whereas in the middle peak, thought to represent summer accumulation the particle population has a (C) factor of 25 percent. (Thompson and Dansgaard, 1975).

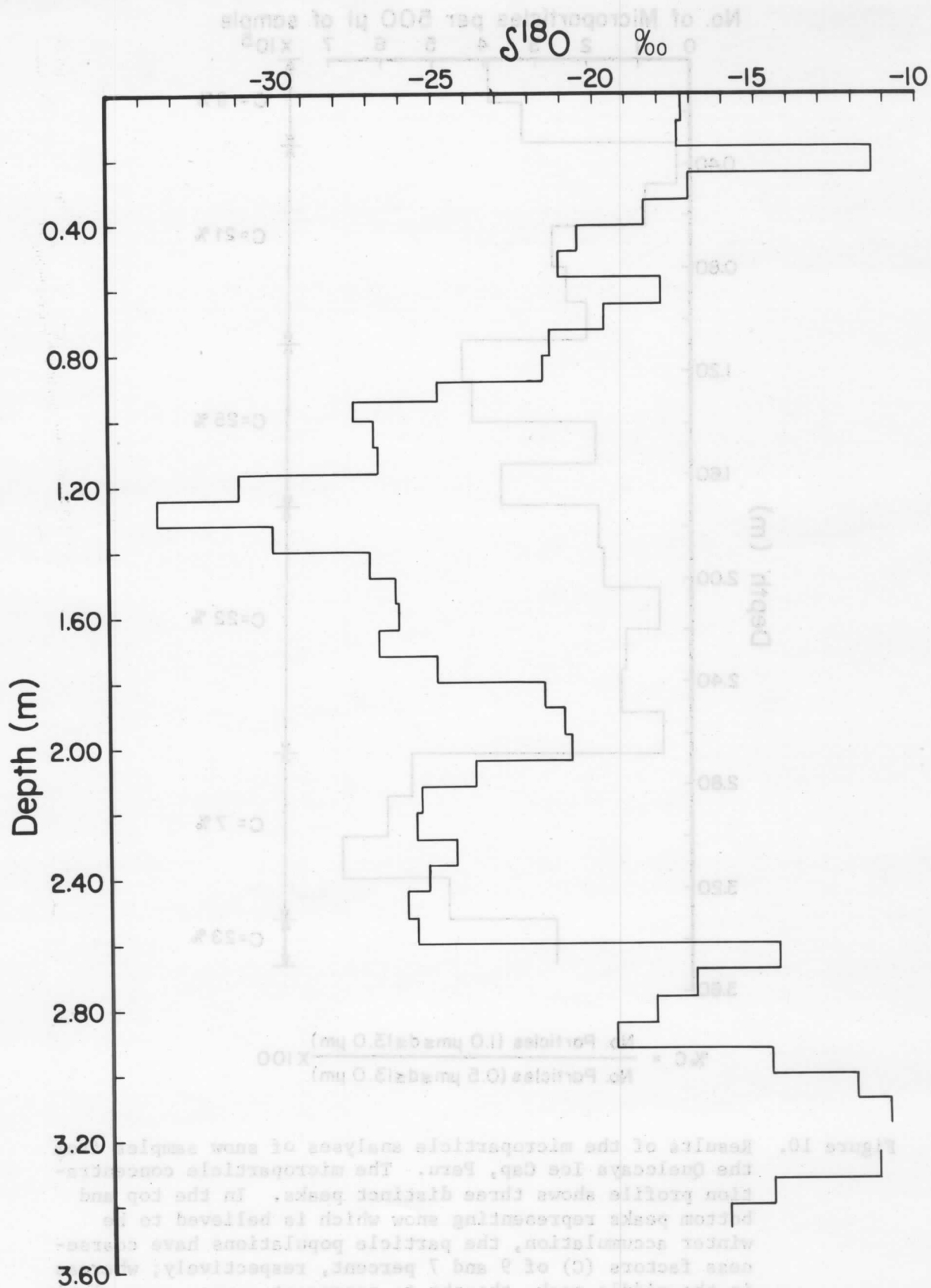


Figure 11. The profile of the stable isotope ratios also shows an apparent seasonal variation. The least and most negative values of  $\delta$  seem to occur in winter and summer snow, respectively. (Data provided by W. Dansgaard).

Hamilton (1969) found in the microparticle analyses on ice cores from Greenland and Antarctica that particles generally are larger in summer and smaller in winter. Thus, the ratio of coarse particles to fine particles could be used to distinguish summer snowfall from winter snowfall. This relationship appears to be valid for the Quelccaya Ice Cap microparticle samples. However, the coarseness ratios calculated for the Graham Land and the Byrd Station Strain Network samples presented in this chapter show no distinct change in the coarseness of particles deposited in summer compared to winter. Consequently, it may be concluded that the coarseness of particles deposited during a year in any given area is largely a function of the regional environment (i.e., the amount of land exposure and seasonal changes in the direction and intensity of air circulation patterns).

The profile of the stable oxygen isotope ratios also shows an apparent seasonal variation. However, it should be noted that the least- and most-negative values of  $\delta^{18}O$  seem to occur in winter and summer snow, respectively. This is an unusual exception to the normal seasonal  $\delta^{18}O$  variations, although it is not the only exception recorded (Dansgaard, 1964). The  $\delta^{18}O$  profile in Fig. 11 exhibits a range of 22 ppm which is greater than any range reported from the polar regions (Thompson and Dansgaard, 1975). In addition, the mean  $\delta^{18}O$  value, -21 ppm, is remarkably low for such a low-latitude site where the mean annual air temperature is suspected to be within a few degrees of  $0^{\circ}C$ .

A comparison of the microparticle and oxygen isotope profiles suggests a value of 3 m for the annual snow layer thickness. However, this must be confirmed with additional cores from future field seasons during which samples should be collected for microparticle, oxygen isotope and  $\beta$  activity measurements.

#### Annual Layers in Deep Ice Cores

Since there is no direct means of determining the period of the particle concentration variations at depth in the Byrd ice core, the assumption is made that the past variations, as preserved in the ice cores, have the same periodicity as they do at the same site in recent snow deposits. In some core lengths the particle cycles are quite clear, while in others they are less distinct. An example of this situation is illustrated in a section from 199 m depth in the Byrd core (Appendix A). In the upper 60 cm of this section the separation between particle peaks appears to be random while in the lower 60 cm cycles appear to be preserved. This can be explained by a reworking of the initial snow after deposition to the degree that the annual layers were destroyed in the upper segment of this core section, while in the lower segment the initial snow deposition remained relatively undisturbed. On the other hand, the annual accumulation at Byrd Station

since 1955 has varied by more than 2:1 from maximum to minimum such that the particle peaks in the upper 60 cm may also represent annual depositions.

In some core sections, such as one from 401 m in the Byrd core (Appendix A), there appear to be well preserved cycles. Even at great depth, such as at 1634 meters, a well developed cycle in the particle profile occurs. In each of these core sections it is necessary to determine whether a cycle exists and, if so, evaluate its properties. The degree to which this can be done depends upon the length and the continuity of each core section.

For the Camp Century core the period of the particle cycle can be checked for approximately the last 8000 years, for which the annual oxygen isotope variations are still present (Johnsen and others, 1972). The layer thicknesses determined by these two independent means are in essential agreement in the upper 1060 meters of the Camp Century core as illustrated in Fig. 15 in Chapter IV; this strongly suggests that the oxygen isotope and microparticle cycles measured are identical.

The choice of sample size for analyses conducted at depth in any deep ice core is crucial for detecting annual cycles as is illustrated by the core increment from a depth of 898 m in the Camp Century core (Appendix B). In the lower 50 cm of this core section, a sample size of 0.7 cm was used, and for this segment the annual particle variations are well defined. For the upper 60 cm of this same core section the sample size was 2.5 cm and for this section the annual variations, although still defined, are less distinct than in the lower section. With increasing depth, sample size becomes more crucial because too large a sample size results in missing annual layers by lumping.

It can be argued, for example, that the sample size employed in the core section from a depth of 1066 meters, the microparticle profile for which is illustrated in Appendix B, is too large to detect all the annual layers, so that the variations noted are not annual. However, it is difficult to visualize another sedimentary process which would produce the cycle shown by this particle profile. In the particle profiles at depth in both cores it is possible some annual cycles have been missed and the cyclicity determined is a function of the concentration of particles in specific horizons which result from the migration of particles during recrystallization. The period of these cycles is of primary importance for determining the age of deep ice cores. In Chapter IV the particle variations given in Appendices A and B are employed to establish a chronology for these two deep ice cores.

## CHRONOLOGIES FOR THE BYRD AND CAMP CENTURY DEEP ICE CORES

In the past decade considerable attention has been given to determining the age of the bottom ice for the Byrd Station and the Camp Century deep ice cores. A significant advance was made in this direction when the  $\delta^{18}O$  curves for these two cores were applied to a logarithmic time scale (Johnsen and others, 1972). They calculated a maximum age for the bottom ice in the Byrd Station core to be greater than 84,000 years B.P. and in the Camp Century core to be greater than 120,000 years B.P. The calculations of the time scales for these cores were made assuming a constant thickness of ice, a constant flow pattern and a constant thermal regime back through the time intervals indicated. Since these factors have changed, it means their time scale is inaccurate (Mörner, 1972). However, it remains to be determined just in what direction and to what degree the time scale is in error. In this chapter the micro-particle variations are used to determine independent age estimates for the bottom ice in both the Byrd Station and Camp Century deep ice cores.

The deposition of microparticles has been shown in the previous chapter to be cyclic and of an annual nature at some locations in Antarctic and Greenland. Thus, assuming that post-depositional processes have not greatly disrupted this cyclic pattern, using the Byrd and Camp Century deep ice core microparticle data, it is possible to calculate an age for the bottom ice in both the West Antarctica Ice Sheet and with less reliability, for the Greenland Ice Sheet. Although this could be accomplished by counting the number of particle concentration peaks (or lows) for the entire core, in practice the analysis of the entire deep core is not feasible. Therefore, core sections from various depths in both deep ice cores were analyzed and annual layer thicknesses determined.

### The Byrd Station Deep Ice Core

In this investigation 23 core sections (totaling 27.4 meters) selected from various depths of the Byrd Station deep ice core and 10 meters of surface cores were analyzed to determine the microparticle concentrations and size distributions. The length and depth of the individual Byrd Core sections are given in Table 2 and the particle profiles resulting from the analyses of these sections are presented in Appendix A. For each profile the number of particle peaks were counted and the average thickness of the annual layer of ice ( $a_1$ ) calculated.

Each unanalyzed section is divided in half, and the annual accumulation in each half is assumed to be the value estimated in the adjacent analyzed section. The calculated  $a_1$  values are plotted against depth in Fig. 12. A gradual reduction in the annual layer thickness toward the bottom of the core is due to thinning by vertical strain.

TABLE 2.

Byrd Station, Antarctica data

A	B	C	D	E	F
0	7.00	100.0	821	6700	-32.0
61	180.00	31.0	391	1665	-33.0
74	199.33	127.0	280	5494	-33.5
144	298.75	135.0	231	10384	-32.9
216	401.29	137.0	684	14519	-32.8
552	896.94	101.5	650	11282	-34.3
587	952.91	136.0	477	1993	-34.7
651	1046.79	154.5	515	17433	-34.5
751	1194.06	74.0	977	8716	-37.3
767	1225.75	151.0	699	12062	-37.0
785	1252.53	150.0	910	8639	-38.8
820	1304.72	130.0	3770	19456	-40.5
852	1342.71	151.0	3533	31912	-41.2
875	1387.00	76.0	3345	42302	-42.2
884	1389.66	150.0	3942	23017	-42.2
982	1547.39	153.0	5586	46270	-40.8
1016	1599.00	56.5	3186	25367	-40.5
1049	1633.72	142.5	2130	20112	-39.5
1115	1748.00	151.0	2390	11125	-40.5
1229	1900.00	146.00	1710	7484	-40.5
1297	2022.31	148.0	732	5383	-38.7
1393	2139.61	139.3	771	6811	-34.3

A = Core Tube Number

B = Depth to top of core from 1968 surface in meters

C = Length of section analyzed in centimeters

D = Concentration of 0.65 to 0.82  $\mu\text{m}$  diameter particles per 500  $\mu\text{l}$  of sample in cleanest 10% of each sectionE = Mean total particle concentration greater than 0.65  $\mu\text{m}$  in diameter per core section

F = Mean oxygen isotope values for core sections analyzed in C

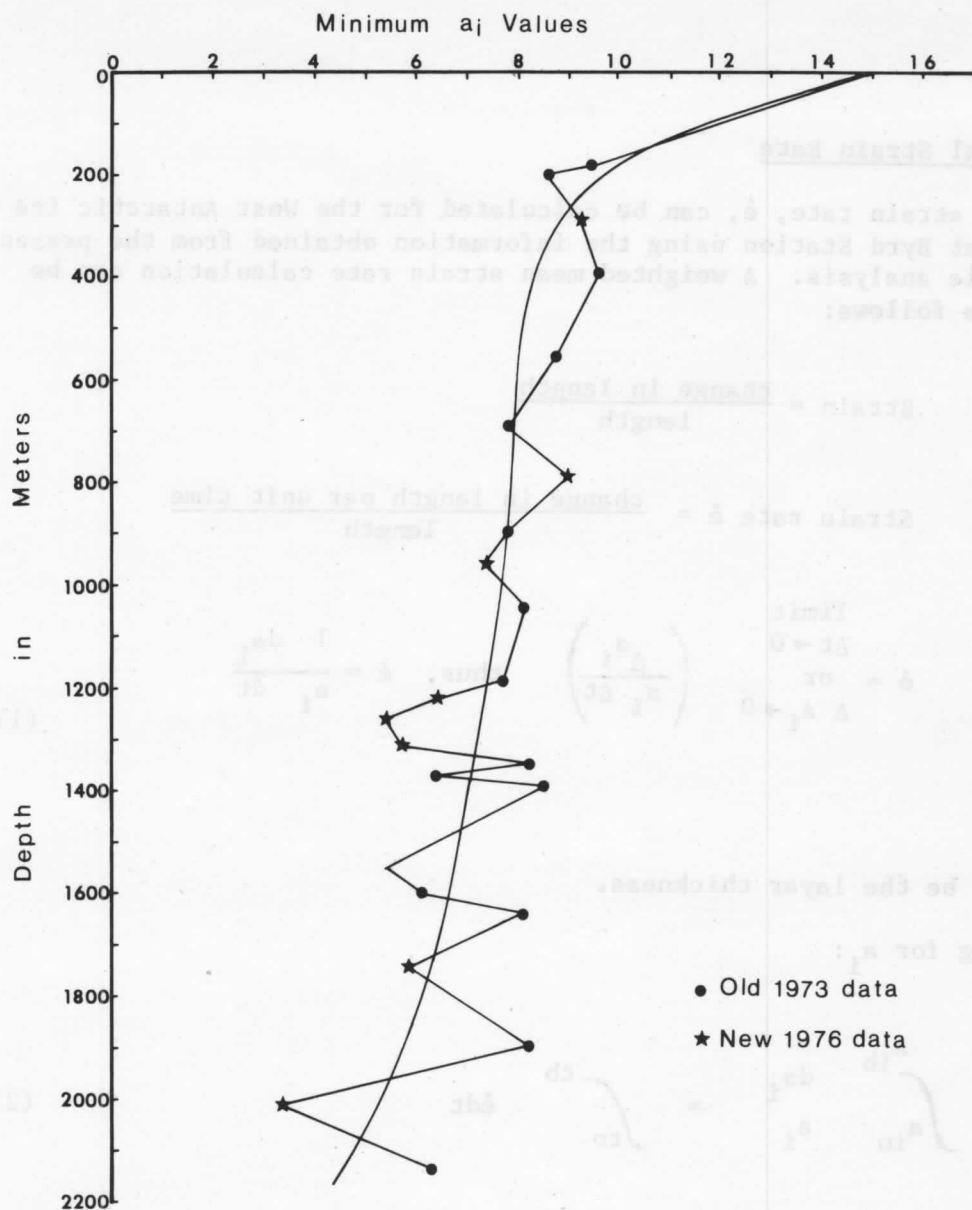


Figure 12. The annual accumulation of ice in cm ( $a_i$ ) for the Byrd Station core as determined from the maximum particle peak counts on the particle profiles in Appendix A are plotted according to depth in meters. The solid curved line represents the average  $a_i$  values. The old data are those reported by Thompson (1973) and the new data are those from core sections analyzed since 1973.

### Vertical Strain Rate

The strain rate,  $\dot{\epsilon}$ , can be calculated for the West Antarctic ice sheet at Byrd Station using the information obtained from the present particle analysis. A weighted mean strain rate calculation can be made as follows:

$$\text{Strain} = \frac{\text{change in length}}{\text{length}}$$

$$\text{Strain rate } \dot{\epsilon} = \frac{\text{change in length per unit time}}{\text{length}}$$

$$\dot{\epsilon} = \lim_{\Delta t \rightarrow 0} \text{ or } \lim_{\Delta a_i \rightarrow 0} \left( \frac{\Delta a_i}{a_i \Delta t} \right) \quad \text{thus, } \dot{\epsilon} = \frac{1}{a_i} \frac{da_i}{dt} \quad (1)$$

Let  $a_i$  be the layer thickness.

Solving for  $a_i$ :

$$\int_{a_{io}}^{a_{ib}} \frac{da_i}{a_i} = \int_{t_o}^{t_b} \dot{\epsilon} dt \quad (2)$$

Finally,

$$\ln a_{ib} - \ln a_{io} = \int_{t_o}^{t_b} \dot{\epsilon} dt \quad (3)$$

where  $a_{ib}$  is the annual layer thickness at depth  $b$ , as determined from microparticle data (Fig. 12),  $a_{io}$  is the surface balance assuming a condition of ice sheet equilibrium and  $t_b - t_o$  is the time interval between the present and the time of deposition on the surface of the snow now at depth  $b$ . The value of  $t_b - t_o$  is obtained directly from the particle time scale derived for the Byrd core (Fig. 13). The Byrd Station surface balance,  $a_{io}$ , varies between  $14 \text{ gm cm}^{-2} \text{ yr}^{-1}$  (Bull, 1971) and  $10.4 \text{ gm cm}^{-2} \text{ yr}^{-1}$  (Cameron, 1971).

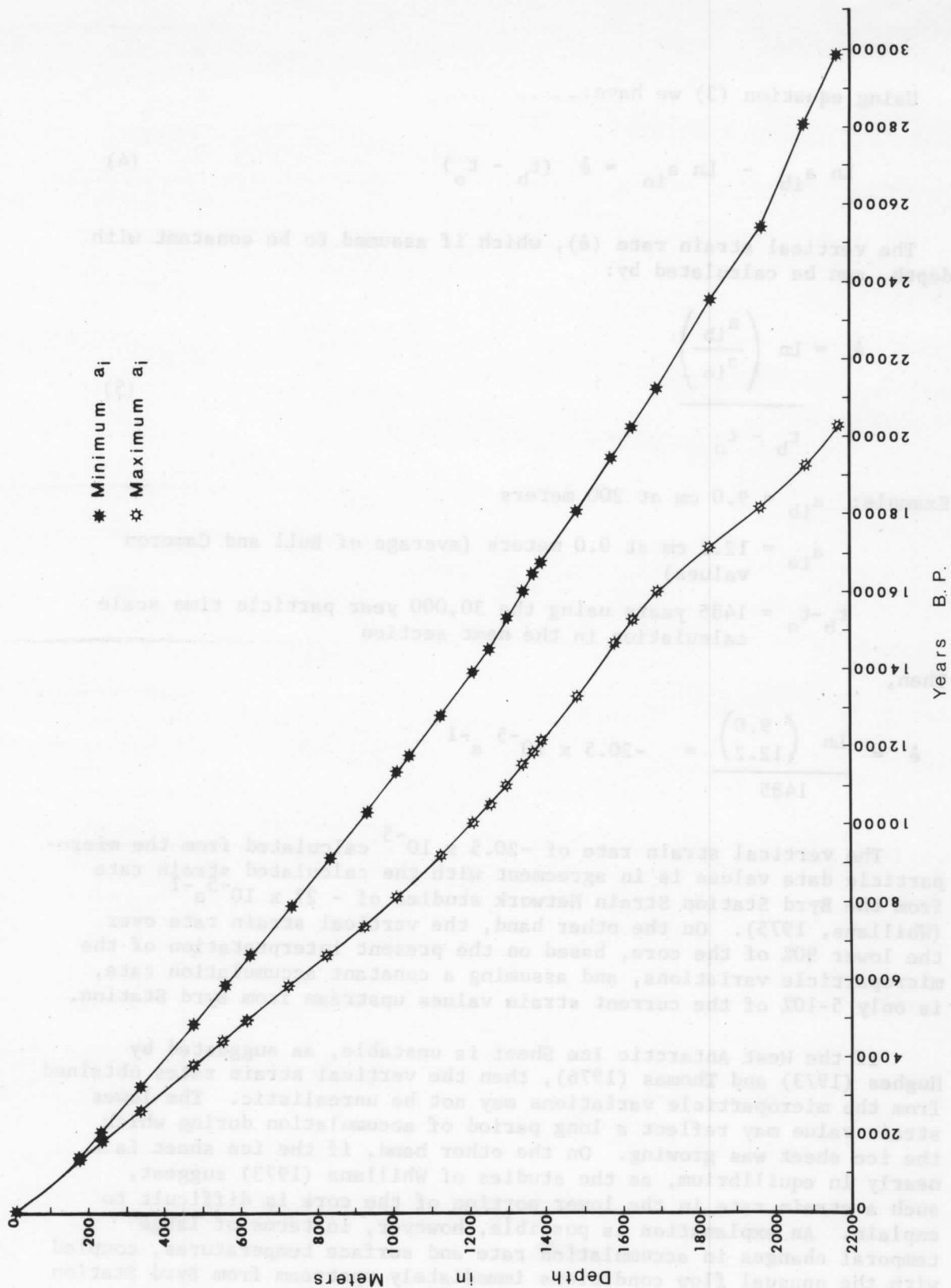


Figure 13. Time-depth relationships for the Byrd core established by using minimum  $a_i$  (solid stars) and maximum  $a_i$  (open stars) obtained from the particle profiles in Appendix A.

Using equation (3) we have:

$$\ln a_{ib} - \ln a_{io} = \dot{\epsilon} (t_b - t_o) \quad (4)$$

The vertical strain rate ( $\dot{\epsilon}$ ), which if assumed to be constant with depth, can be calculated by:

$$\dot{\epsilon} = \frac{\ln \left( \frac{a_{ib}}{a_{io}} \right)}{t_b - t_o} \quad (5)$$

Example:  $a_{ib} = 9.0$  cm at 200 meters

$a_{io} = 12.2$  cm at 0.0 meters (average of Bull and Cameron values)

$t_b - t_o = 1485$  years using the 30,000 year particle time scale calculation in the next section

Then,

$$\dot{\epsilon} = \frac{\ln \left( \frac{9.0}{12.2} \right)}{1485} = -20.5 \times 10^{-5} \text{ a}^{-1}$$

The vertical strain rate of  $-20.5 \times 10^{-5}$  calculated from the micro-particle data values is in agreement with the calculated strain rate from the Byrd Station Strain Network studies of  $-27 \times 10^{-5} \text{ a}^{-1}$  (Whillans, 1975). On the other hand, the vertical strain rate over the lower 90% of the core, based on the present interpretation of the microparticle variations, and assuming a constant accumulation rate, is only 5-10% of the current strain values upstream from Byrd Station.

If the West Antarctic Ice Sheet is unstable, as suggested by Hughes (1973) and Thomas (1976), then the vertical strain rates obtained from the microparticle variations may not be unrealistic. The lower strain value may reflect a long period of accumulation during which the ice sheet was growing. On the other hand, if the ice sheet is nearly in equilibrium, as the studies of Whillans (1973) suggest, such a strain rate in the lower portion of the core is difficult to explain. An explanation is possible, however, in terms of large temporal changes in accumulation rate and surface temperatures, coupled with the unusual flow conditions immediately upstream from Byrd Station as a result of the bottom topography. It is quite likely that the

annual accumulation rate has changed significantly in the last 30,000 years, but such variations alone would not explain easily the apparent  $a_i$  pattern.

### Age Calculation

The number of years represented in each ice core segment can be calculated by the following simple relationship:

$$\Delta t = \frac{\Delta h}{a_i} \quad (6)$$

where:  $\Delta t$  = age interval;

$\Delta h$  = the depth interval measured in cm over which a given  $a_i$  value is assumed to be representative; and

$a_i$  = the distance between annual layers, i.e., net annual accumulation measured in cm as determined from peak counts for each section.

Based on preliminary studies, the age of the bottom ice at Byrd Station was estimated at 30,000 years (Thompson, 1973; Thompson and others, 1975). Since these initial studies were made 10 more core sections have been analyzed. The new and old data are plotted in Fig. 12 which illustrates that the new data are in essential agreement with the old and, hence, support the 30,000 year age estimate. Fig. 13 provides a comparison of the age-depth relationship in the Byrd Core as calculated from both minimum and maximum particle  $a_i$  values.

### Discussion

It must be emphasized that the age calculation based upon the separation of the peaks in the particle profile depends on the period of the cyclic deposition of particles, which is assumed to be constant with time, as well as on the degree of post-depositional changes in the stratigraphy. Sample size in the lower portions of the core introduces a potential problem (because thinning with depth may lead to lumping of annual layers) but experiments in which the sample size was reduced resulted in no change of the  $a_i$  values.

The new laboratory enables sample size to be reduced to 0.7 cm length so that cycles of 2.0 cm wavelength would be apparent. Even if the vertical strain retains its near-surface value throughout the

profile and accumulation rates are unchanged, the values of  $a_1$  should exceed 2 cm for 80% of the ice thickness. This is demonstrated in core sections from depths of 1500, 1634, 1749, 1900, 2022, and 2140 meters, the microparticle profiles for which are shown in Appendix A. A varying sample size has been used for these core sections, and it is instructive to note that for the core sections from a depth of 1749 meters the reduction of sample size to 0.7 cm serves to better define the larger cycle of 13 cm wavelength. A large cycle of 12 to 14 cm wavelength can be found also in the core samples from 1634, 1900, 2022 and 2140 meters depth.

The cyclicity of the 13 cm wavelength may be due to the concentration of particles within specific horizons as a result of the migration of particles during recrystallization. However, this cannot be confirmed. Another possible explanation for the large  $a_1$  values at depth is that the variations represent a larger period cycle. Core profiles from depths of 1105, 1226 and 1254 meters suggest the existence of some longer-term variations which at depth may be misinterpreted as annual cycles.

The explanation for the relative low age estimated for the bottom ice surface compared to the oxygen isotope age of greater than 84,000 years B.P. is possibly due to melting of the bottom ice by geothermal and frictional heat. The temperature gradient near the bottom of the Byrd hole obtained by Ueda and Garfield (1968), appears to be  $3.25^{\circ}\text{C}$  per 100 meters and this implies that basal ice is melted each year. Indeed many other mechanisms could also suffice for removing many meters of ice from the bottom of the ice sheet; for example, through the ice flow associated with ascending convection flow as suggested by Hughes (1970, 1972). Furthermore, it is possible that a break in the sedimentary record may exist in the Byrd core or in the Camp Century core discussed later.

#### The Camp Century Deep Ice Core

The 30 Camp Century deep ice core sections (totaling 14.15 meters) from various depths were analyzed for particle concentration and size distribution. Table 3 gives the length of each Camp Century section. Examination of Tables 2 and 3 illustrates that the individual Camp Century sections are smaller than the Byrd Core sections which results in less accurate individual  $a_1$  values for the Camp Century core.

The same procedure is used to determine the age for the Camp Century core as used for the Byrd core. The individual  $a_1$  values plotted in Fig. 14 are derived from the data in Appendix B. The stars in Fig. 14 indicate  $a_1$  values determined from the oxygen isotope variations of Johnsen and others (1972), while the dots indicate the microparticle  $a_1$

TABLE 3.

## Camp Century, Greenland data

A	B	C	D	E	F
141-145	80.50	436.0	1638	11273	-28.5
523	595.44	54.0	719	10667	-28.9
655	791.05	28.0	1146	12306	-28.9
728	898.03	48.0	971	6630	-28.4
828	1043.74	72.0	3872	31712	-28.8
839	1061.00	35.0	1004	65044	-28.8
842	1065.97	43.0	3224	8882	-29.5
858	1086.63	30.0	3278	54647	-29.6
873	1108.97	13.0	14544	22330	-30.7
893	1135.09	42.0	3042	12806	-32.6
908	1155.56	28.0	7268	46553	-35.2
915	1166.00	32.0	75307	374147	-39.4
921	1178.71	32.0	36163	372080	-40.6
929	1185.98	20.0	98402	551526	-40.9
937	1197.68	20.0	51258	202946	-41.6
942	1204.87	20.0	78616	499722	-41.6
949	1214.97	20.0	97786	289469	-41.5
955	1223.88	22.0	20153	64107	-37.4
962	1234.48	30.0	48405	154061	-38.2
969	1244.98	12.0	20145	53387	-37.9
976	1255.14	20.0	16193	50643	-38.4
983	1265.46	22.0	22445	106417	-38.5
992	1278.82	20.0	67592	216619	-42.8
997	1285.49	18.0	12660	38520	-38.5
1004	1295.59	20.0	7731	24016	-36.7
1011	1305.97	15.0	17255	51399	-39.2
1024	1324.76	17.0	14568	54816	-36.4
1031	1334.69	32.0	5248	27634	-30.4
1038	1345.43	20.0	31613	108081	-28.8
1045	1354.86	30.0	9863	59904	-27.6

A = Core Tube Number

B = Depth to top of core from 1968 surface in meters

C = Length of section analyzed in centimeters

D = Concentration of 0.65 to 0.82  $\mu\text{m}$  diameter particles per 500  $\mu\text{l}$  sample in cleanest 10% of each sectionE = Mean total particle concentration greater than 0.65  $\mu\text{m}$  in diameter per core section

F = Mean oxygen isotope values for core sections analyzed in C

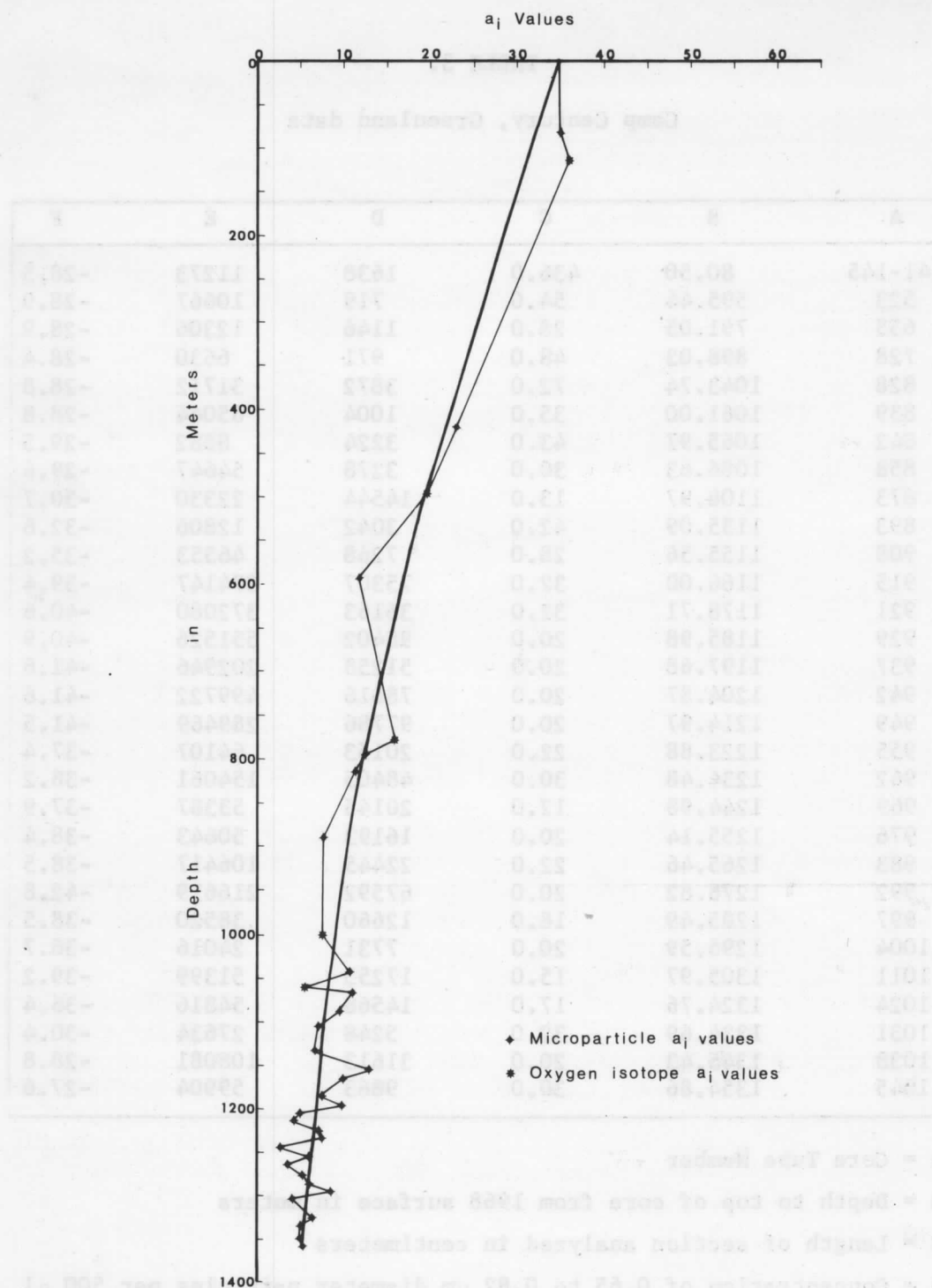


Figure 14. The annual accumulation of ice in cm ( $a_i$ ) for the Camp Century core as determined from the particle counts on the particle profiles in Appendix B are plotted according to depth in meters. Also average  $a_i$  values as obtained from detailed oxygen isotope analysis (Johnsen and others, 1972) are plotted according to depth. The solid curved line represents the average  $a_i$  value.

values obtained in this work. From Fig. 14 it is apparent that the microparticle  $a_i$  values and the oxygen isotope  $a_i$  values agree well for the upper 1000 meters of the Camp Century core.

#### Vertical Strain Rates

Vertical strain rates have been calculated for the Camp Century deep ice core as they were for the Byrd core earlier. Using the  $a_i$  values in Fig. 14 the  $a_{i0}$  value of  $35 \text{ gm cm}^{-2}\text{yr}^{-1}$  (Croaz and Langway, 1966) and the value of  $(t_b - t_0)$  derived for this core from Fig. 15, the vertical strain rate was calculated to be  $-16.5 \times 10^{-5} \text{ a}^{-1}$  for the entire Camp Century core. In the Camp Century core the calculated vertical strain rates are constant with depth, unlike those calculated for the Byrd Core.

#### Age Calculation

The microparticle age calculation for the Camp Century deep ice core cannot be calculated with the same reliability as for the Byrd core because of the shortness of the individual core sections analyzed. However, the microparticle and oxygen isotope  $a_i$  values in the ice core agree well (Fig. 14) and lend support to the validity of individual microparticle  $a_i$  determinations.

Where annual variations are detectable in the oxygen isotope and microparticle data, i.e., down to a depth of about 1000 meters (6,200 years B.P.) the age calculations are in agreement with the logarithmic time scale. Below this depth the disparity increases with depth. At a depth of 1100 meters the microparticle age is 7,300 years B.P. while the logarithmic time scale yields 8,500 years B.P.; at 1200 meters depth the microparticle age is 8,488 years B.P. while the logarithmic time scale gives 15,000 years B.P. Similarly, at 1300 meters the microparticle age is 10,500 years B.P. while the logarithmic time scale yields 30,000 years B.P. and at the bottom of the core (1387 meters) the microparticle age is 11,803 years B.P. while the logarithmic time scale age has increased to 120,000 years B.P. Using the logarithmic time scale, 100,000 years of the entire age estimate occurs within the lower 100 meters of the Camp Century core.

#### Discussion

The preliminary time scale for the Camp Century core was established by model calculations (Dansgaard and others, 1971) by means of the flow theory presented by Nye (1959) along with certain assumptions concerning parameters which influence it. The preliminary time scale was then corrected by means of Fourier spectral analysis employing the

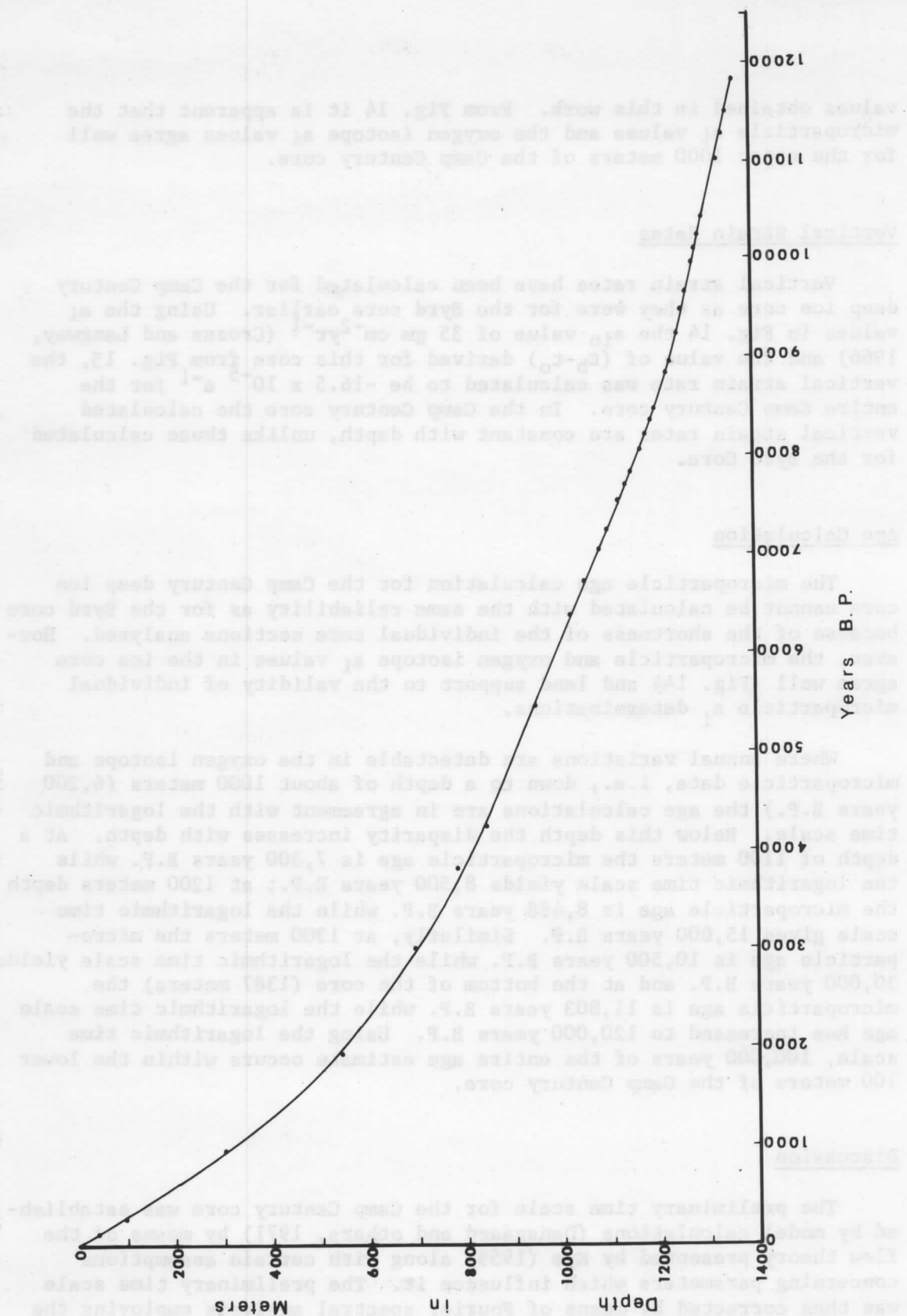


Figure 15. Time-depth relationship for the Camp Century core established by using the  $a_i$  values obtained from the particle profiles in Appendix B and from the oxygen isotope  $a_i$  values as determined by Johnsen and others, 1972.

assumptions that the observed oscillations in the oxygen isotope data are caused by varying solar activity and that the periodicity of these solar fluctuations have been constant in time (Dansgaard and others, 1971). The climatic oscillations thought to be detected in the Camp Century core exhibited periods of 2,400 400, 180 and 78 years.

Much interest has been expressed in the study of climatic cycles some of which are gaining general acceptance. However, LaMarche (1975, personal communication) cautioned that careful reading of the literature reveals a certain degree of circularity in reasoning. LaMarche gives the 2,400 year cycle as a case in point. In the Camp Century analysis a spectral peak was found between 2000 and 3000 years based on the dates for the core obtained by the flow model. Then the time scale was adjusted so that the spectral peak matched the 2,400 year peak in the radiocarbon data obtained from tree ring studies. Conversely, Suess (1970) accepted the validity of a 2,400 year radiocarbon cycle because it was found in the Camp Century oxygen isotope data. The existence of a 2,400 year cycle remains to be proven.

Similarly Denton and Karlén (1973) have suggested a 2,400 year cycle in the neoglaciation record. Their study areas are geographically restricted, and there are some inherent problems with the recognition and dating of advances of valley glaciers. Therefore, given the uncertainty of their placement in time, these minor glacial events might easily be fitted into a cyclic pattern, which, in fact, is non-existent.

In the Camp Century deep ice core a very real possibility of a break in the sedimentary record exists. The compositions of particles in core section 962 from a depth 1234.48 meters in the Camp Century deep ice core (Appendix D, Table 5) are atypical. Six out of the seven particles analyzed from this core section contain substantial amounts of tin (Sn). Sn only occurs in one particle of those analyzed above this depth in the Camp Century core and in only one particle in the entire group of particles analyzed for the Byrd Station core. Molybdenum (Mo) is even more unusual in that it occurs in 4 out of the 6 particles analyzed in this same section and is not present in any of those particles analyzed above the 1234.48 meter depth in the Camp Century core or in the entire group of particles analyzed for the Byrd core. However, Sn and Mo do appear together again in the lowest section (Core number 1038, 1345.43 meters depth) of the Camp Century core (Appendix D, Table 5).

The elements of Mo and Sn are not naturally common. Therefore, the first possible source for these particles which must be considered is contamination. One of the common sources for Sn is solder. Solder consists of Sn combined in some proportion with Pb. However, in the analyses no Pb was detected in the particles. In addition, if solder is the source for Sn then it still would be difficult to find a source

for Mo. Another possible source would be through some fluke in the SEM analyses. However, there are particles which were analyzed in the 962 and 1038 core sections which contain no Sn or Mo (See Appendix D). A thorough investigation of the laboratory and SEM facilities has not revealed any contaminate source for these particles.

The elements Sn and Mo occur naturally together in hypothermal deposits, that is in deposits formed at great depth and high temperatures (300-500°C). An example of this type of deposit is the tin-copper district of Cornwall, England (Park and MacDiarmid, 1970). Here Sn occurs as the mineral Cassiterite ( $\text{SnO}_2$ ) and in the mineral Stannite ( $\text{Cu}_2\text{Fe Sn S}_4$ ) and Mo occurs in the mineral Molybdenite ( $\text{MoS}_2$ ).

Since Sn and Mo are only present in the bottom 200 meters of the Camp Century core, it can be argued that the source of the Sn and Mo is the bedrock over which the Greenland Ice Sheet moves. Perhaps the particles were entrained in the ice as it moved over an irregular bedrock surface. If this is the case then the source of the particles should be some distance along the flowline upglacier from the Camp Century borehole. The surface velocity at Camp Century is 3.3 meters per year (Mock, 1968). Therefore, ice at a depth of 1200 meters, that is ice which is 8,500 years old by the microparticle chronology or 15,000 years old by the oxygen isotope chronology, would have originated no further than 50 kilometers up flowline from the Camp Century borehole. Radio echo sounding interpretations (Robin and others, 1969; Gudmandsen, 1973) indicate little bottom relief in the borehole area. However, relatively large bedrock relief does lie within 30 kilometers south and southwest of the borehole but in the opposite direction of the present flowline. The above suggests that the present flowline may be quite different from that which existed 8,000 to 15,000 years ago. Another mechanism by which particles could be transported upward two hundred meters would be through ascending convection flow as suggested by Hughes (1972). Whatever the mechanism, if the particles at 1200 meters depth are from the bottom then the microparticle and oxygen isotope stratigraphy below this depth cannot be meaningfully interpreted.

A direct interpretation of the particle  $a_1$  values for the Camp Century core suggests that the basal ice is of late Wisconsin age, but if this is the case it is difficult to explain the decrease in  $\delta^{18}\text{O}$  values at a depth of 1150 meters which traditionally has been interpreted to represent the Holocene/Wisconsin stage transition. However, in the sequence of deglaciation in North America, the breaking up of the Laurentide Ice Sheet occurred between 8,000 and 7,000 years B.P. (Flint, 1971). Before this time Hudson Bay and probably Baffin Bay and Davis Strait were ice covered. Perhaps the sudden increase in  $\delta^{18}\text{O}$  values, a change from approximately -41 to -31, corresponds to the opening of these water bodies and consequently to a local source of water vapor for North Greenland. Present storm tracks of occasional and seasonal importance pass over the northern and southern sections

of the Hudson Bay (Flint, 1971). These same storm tracks would have been blocked or displaced when the Laurentide Ice Sheet existed. Thus, the change in  $\delta$  values may be a geographical effect coupled with the fact the ice in the lower portion of the ice core has its origin at a higher elevation on the ice cap. Snow being deposited today near the ice divide in northern Greenland has delta values which are close to those found in the Camp Century ice core which have been interpreted to represent Wisconsin ice.

It has long been recognized that many factors affect the oxygen isotope ratios of precipitation. The Quelccaya Ice Cap oxygen isotope ratios presented in the previous chapter exhibit a  $\delta$  range of 22 ppm while the mean annual temperature appears to vary by only 2 to 3°C. Thus, it is apparent that other factors besides temperature may play the dominant role in determining the  $\delta$  value of precipitation. In the framework of this interpretation on the large increase in clay-like particles could be related to the wasting of the Laurentide Ice Sheet and therefore a source of outwash which could be wind eroded and deposited on the relatively close Greenland Ice Sheet.

It is, however, possible to interpret the microparticle variations in a more traditional framework. Because of the lack of conclusive data to support this interpretation a future chapter focuses on the relationship between dust and climate over the past 30,000 years using the oxygen isotope time scale for the Camp Century core. The initial work on this core illustrates the necessity for continuous detailed research on deep ice cores from Greenland and Antarctica to ascertain a better interpretation of these cores.

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It has long been recognized that many factors affect the oxygen isotope ratios of precipitation. The Quaternary Ice Cap oxygen isotope ratios presented in the previous chapter exhibit a  $\delta$  range of 25 gms while the mean annual temperature appears to vary by only 1 to 2°C. Thus, it is apparent that other factors besides temperature may play the dominant role in determining the  $\delta$  value of precipitation. In the framework of this interpretation on the large increase in clay-like particles could be related to the waning of the Laurentide Ice Sheet and there-fore a source of outwash which could be wind eroded and deposited on the relatively close Greenland Ice Sheet.

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## MORPHOLOGY AND ELEMENTAL COMPOSITION OF MICROPARTICLES

There are no known direct means for ascertaining the nature of atmospheric processes in the past. Therefore, inferences about these processes must be drawn from other evidence such as the records left in and on surfaces such as lakes, bogs, ice sheets and ocean floors.

Differences in chemical composition of the particles distributed in the deep ice cores indicate changes in the particulate source. The properties of particles which affect the global radiation budget most significantly are the optical depth, the particle size distribution function, and particle optical constants (Toon and Pollack, 1975). Therefore, changes in the composition, the concentration and the size distribution of the particles in the atmosphere could modify the global energy budget and hence, the temperature distributions which serve as driving forces for the general circulation. This chapter presents the data and analyses of the non-soluble particulates found in the Byrd Station and Camp Century deep ice cores.

### Data Analysis

The procedures for the light microscope classification, the SEM examination, and the EDS elemental analysis have been discussed previously. Approximately 5000 particles from each of the two deep ice cores were classified according to the scheme in Table 1; 105 particles from the Byrd core and 97 particles from the Camp Century core were analyzed for elemental composition by the EDS technique.

The relative weight percent concentrations of the elements present in each analyzed particle were obtained from the dispersed X-ray energy spectrum by the Magic IV computer program for microprobe intensity corrections. The elemental concentrations for all analyzed particles from each core were examined to determine the extent of natural groups among them. These natural groups, determined by the computer program Hygroup (Ward, 1963), are based upon intergroup similarity of their scores (elemental weight percents) and on the K variables (the elements) used to describe them.

Hygroup, a hierarchical grouping algorithm, begins by defining each original object (particle) as a group and at each step some pair of groups is combined. The decision as to which two groups to combine is based upon minimizing the within-group variance at each step. Therefore, an error matrix ( $E_{ij}$ ) of the within-group variance for each possible grouping is calculated as follows:

$$E_{ij} = \frac{\sum_{k=1}^m (V_{ki} - V_{kj})^2}{n_i + n_j}$$

where  $E_{ij}$  is the within-group variance for the combination of group  $i$  with group  $j$ ,  $m$  is the number of variables ( $V$  or in this case, chemical elements) describing each particle, and  $n_i$  and  $n_j$  are the number of members in group  $i$  and group  $j$ , respectively.

Initially, the two groups with the lowest  $E_{ij}$  value are combined retaining the label of group  $i$ , and the error measure,  $E_{ii}$ , is moved to the group  $i$  diagonal element. The other matrix elements are then recalculated to reflect the potential error for combination with new group  $i$  (Ward and Hook, 1963). This procedure is repeated until all the objects are combined into 1 group.

At each iteration or grouping the error,  $E_{ij}$ , is listed as the error increase and a sum of the total errors associated with all groupings previously conducted is presented. The error increase reflects the loss of information by employing that particular grouping (Ward and Hook, 1963). A sharp increase in the  $E_{ij}$  indicates that much of the classification system's accuracy has been lost by reducing the number of groups by one at this stage. This information of the relative costs of different numbers of groups provides valuable guidance when the user must choose a specific number of profile categories for classification purposes (Ward and Hook, 1963).

## Results

The results of the Hygroup analysis are presented in Fig. 16, where (A) represents the Byrd core particles and (b) represents the Camp Century core particles. The break between Holocene and Wisconsin particles in both cores is based upon the transition from less negative to more negative  $\delta O^{18}$  values (Tables 2 and 3). The break occurs between core sections 785 and 820 for the Byrd core and sections 908 and 915 for the Camp Century core. As illustrated by the distribution of groups in Fig. 16 (A) the elemental composition of the particles in the upper Holocene sections of the Byrd core differ from those in the Wisconsin sections. In Fig. 17 the groups from Fig. 16 (A) are broken into relative weight percentages of the elements present in those particles placed in groups by the Hygroup program composed of less than 45% Wisconsin, greater than 55% Wisconsin, and between 45% and 55% Wisconsin particles. The particles in the greater than 55% group contained more Si and K than the other two groups. For the Greenland core, on the other hand, the Hygroup program was unable to distinguish any marked stratigraphic change in the elemental composition.

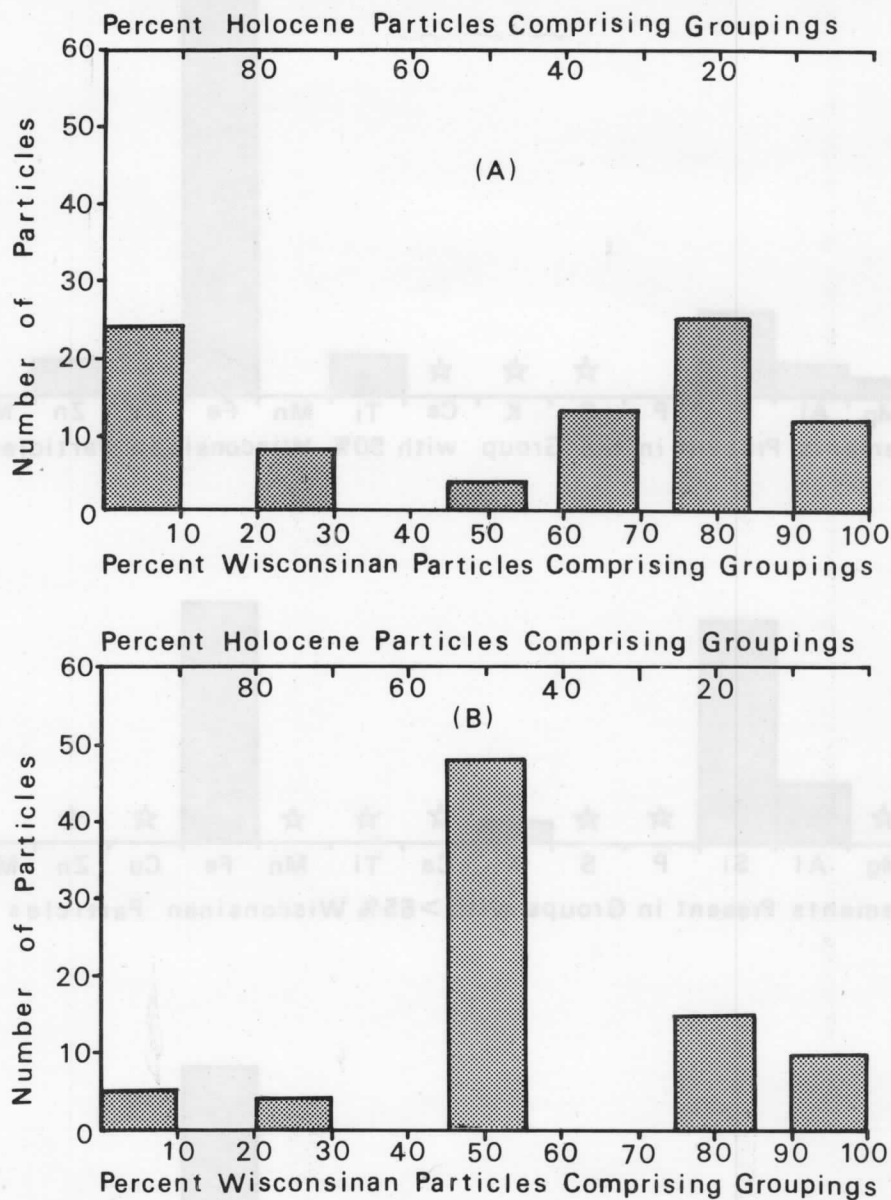


Figure 16. The results of the Hygroup analysis. (A) represents the Byrd core particles; (B) represents the Camp Century core particles. The bimodal distribution in (A) suggests a difference in composition between Wisconsin and Holocene particles. The bell shaped distribution in (B) suggests no change in composition between Wisconsin and Holocene particles in the Camp Century core.

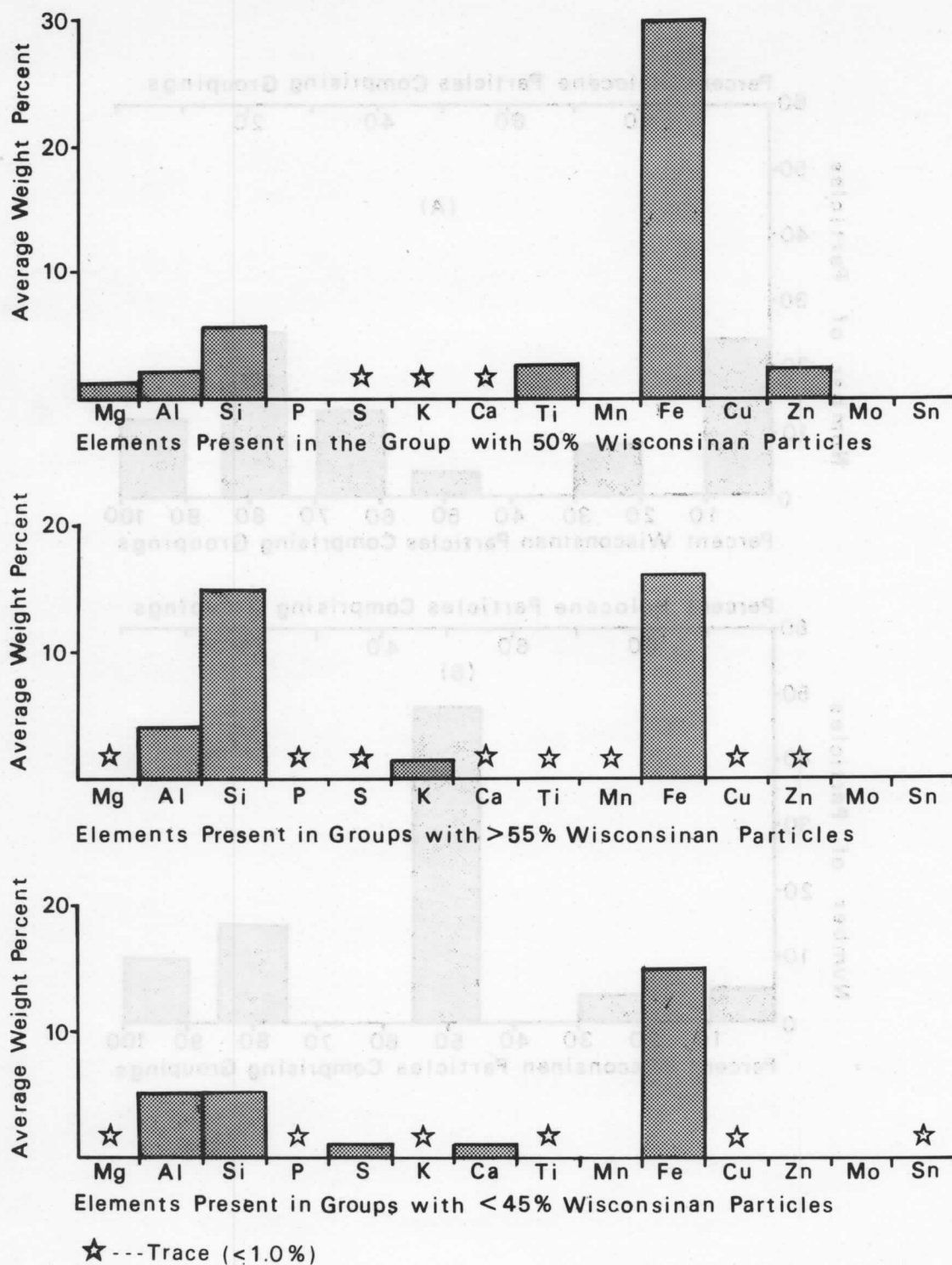


Figure 17. The groups of Fig. 16 (A) are broken into relative weight percentages of the elements present in those particles placed in groups composed of less than 45% Wisconsin, greater than 55% Wisconsin, and between 45 and 55% Wisconsin particles.

This is illustrated in Fig. 16 (B), where the largest group obtained from the Hygroup program is the 45 to 55% Wisconsin group (i.e., the 45 to 55% Holocene group). Those particles which were placed in the greater than 55% Wisconsin group contain relatively higher amounts of Ca, Ti, Sn, Mo, and Cu than the other two groupings (Fig. 18).

In addition to variations in composition, some general distinctions between particles can be made on the basis of particle morphology. In the Byrd core the morphology of Holocene particles is markedly different from that of the Wisconsin particles as illustrated in Appendix C. The Wisconsin particles tend to be much more angular while the particles deposited during the Holocene tend to be more rounded. On the other hand, the majority of the Camp Century particles examined are flat, platy, and clay-like with no marked variation in morphology between the Holocene and Wisconsin age particles as illustrated in Appendix D.

### Discussion

The elemental composition and the changes in morphology suggest that the most of the particles from the Wisconsin sections of the Byrd core are of volcanic origin which agrees with conclusions drawn by Gow and Williamson (1971) using petrographic techniques. The majority of the particles from the Wisconsin sections of the Camp Century core are more likely wind-blown terrestrial material. Volcanic-like particles do occur in the Camp Century deep ice core; however, it is difficult to ascertain the significance of these particles as the total particle concentrations are quite high, as illustrated in the next chapter. In fact, if the volcanic particles have the same concentration in the Wisconsin sections of the Camp Century core as they do in the Wisconsin sections of the Byrd core only one of every 20 particles would be volcanic.

The visual classification of particles into the scheme illustrated in Table 1 coupled with the results of the X-ray energy dispersive analysis system, allows examination of changes in composition of particles with depth in each core. The visual classification scheme was used to classify 350 particles from each core section. Of these particles, the greatest number fell in the clear particle category with the second highest number falling in the black particle category. The variation with depth in the percentage of black particles in each core section in the Byrd Station core is presented in Fig. 19. Fig. 19 (b), (c), (d), (e), and (f) illustrate mean values in the black particles for each core section for variations in Si, Al, Mg, S and Fe, respectively. Several important relationships are suggested. Fig. 19 (a) shows that the percentage of black particles is less in Wisconsin ice. Fig. 19 (b), (c) show a notable increase in Si and Al in these black particles in Wisconsin ice.

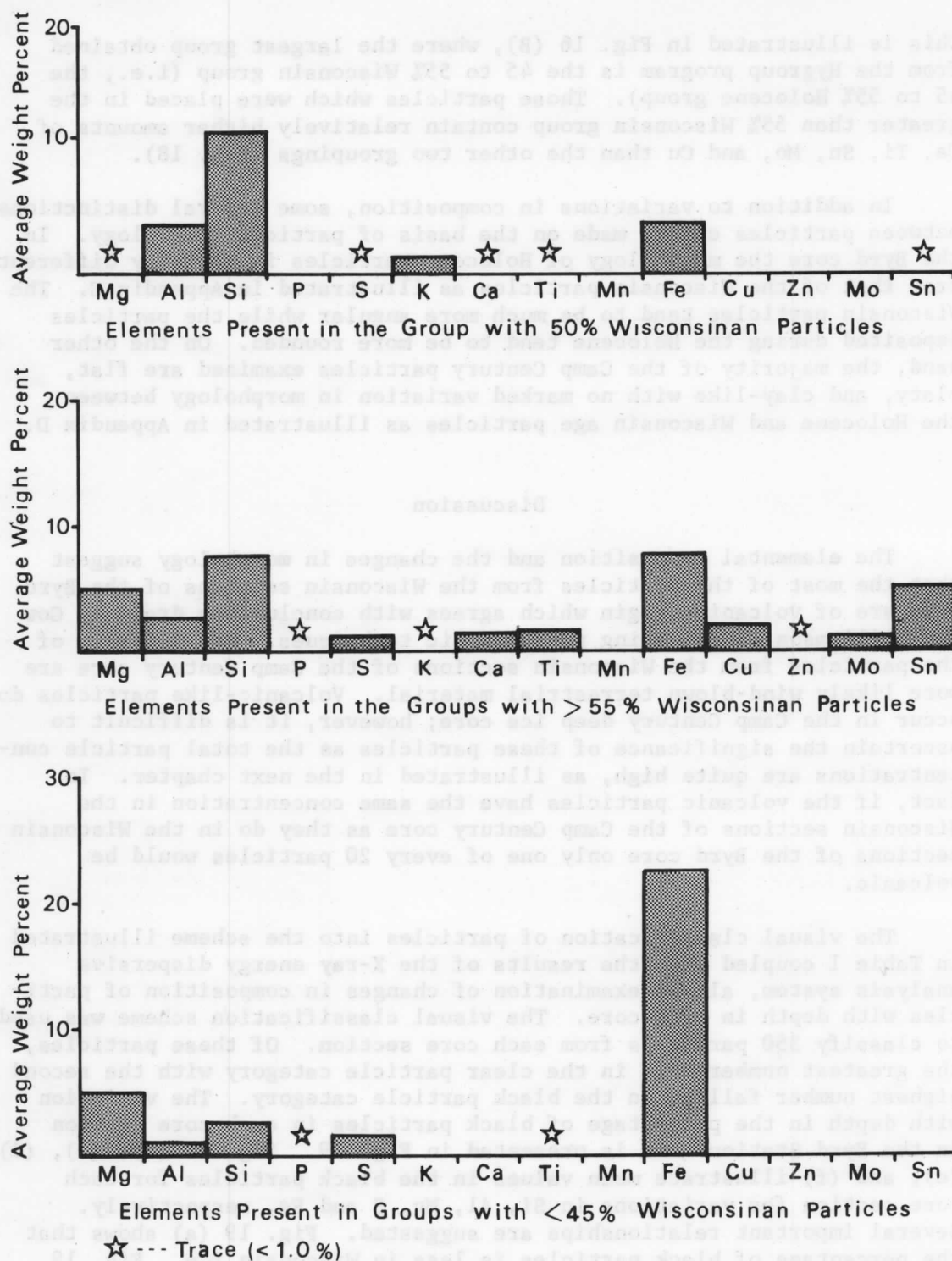


Figure 18. The groups of Fig. 16 (B) are broken into relative weight percentages of the elements present in those particles placed in groups composed of less than 45% Wisconsin, greater than 45% Wisconsin, and between 45 and 55% Wisconsin particles.

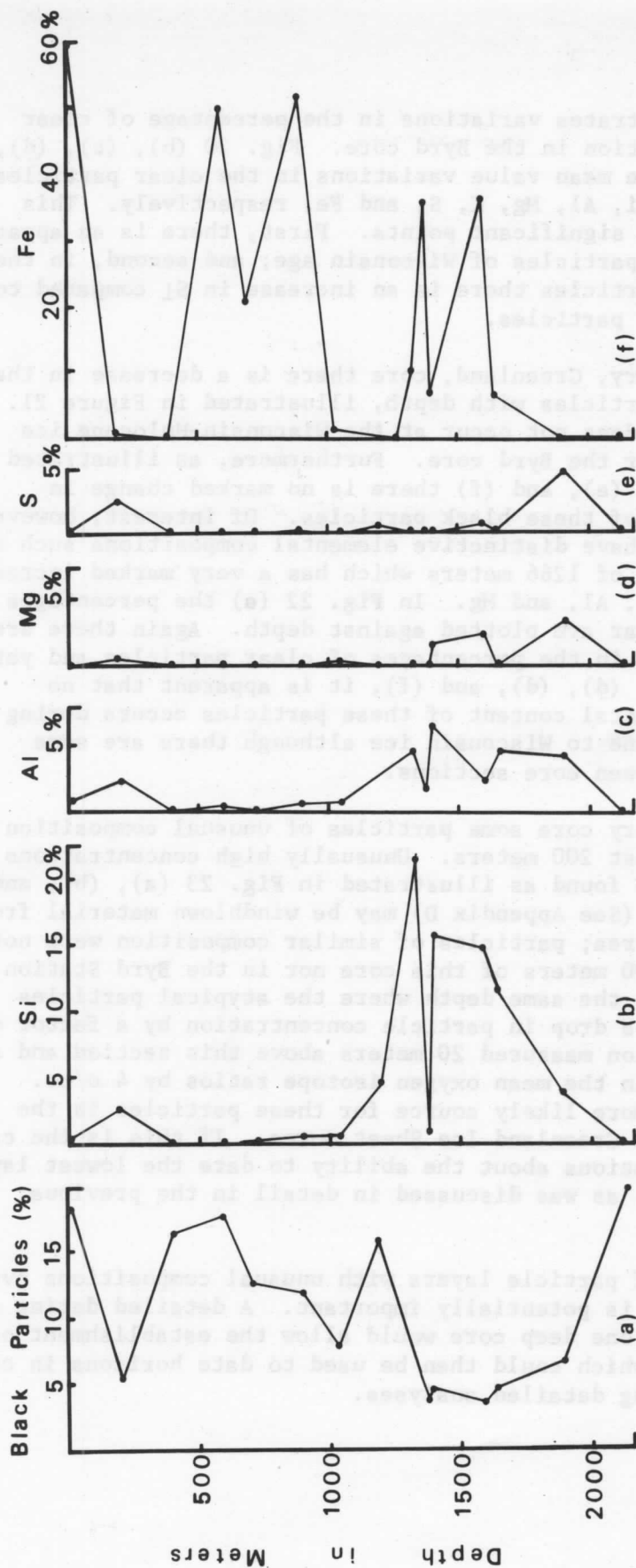


Figure 19. The variation with depth in the percentages of black particles in each core section in the Byrd core is presented in profile (a). In profiles (b), (c), (d), (e), and (f) the mean values for Si, Al, Mg, S, and Fe are plotted for black particles in each core section.

Fig. 20 (a) illustrates variations in the percentage of clear particles per core section in the Byrd core. Fig. 20 (b), (c), (d), (e), and (f) illustrate mean value variations in the clear particles for each section for Si, Al, Mg, K, S, and Fe, respectively. This figure illustrates two significant points. First, there is an apparent increase in the clear particles of Wisconsin age; and second, in these clear Wisconsin age particles there is an increase in Si compared to the Holocene age clear particles.

In the Camp Century, Greenland, core there is a decrease in the percentage of black particles with depth, illustrated in Figure 21. However, the decrease does not occur at the Wisconsin-Holocene ice boundary as it does for the Byrd core. Furthermore, as illustrated by Fig. 21 (b), (c), (d), (e), and (f) there is no marked change in elemental composition of these black particles. Of interest, however, is that some sections have distinctive elemental compositions such as the section at a depth of 1266 meters which has a very marked increase in the abundances of S, Al, and Mg. In Fig. 22 (a) the percentages of particles that are clear are plotted against depth. Again there are some marked variations in the percentages of clear particles and yet from Fig. 22 (b), (c), (d), (d), and (f), it is apparent that no marked change in elemental content of these particles occurs during the transition from Holocene to Wisconsin ice although there are some marked variations between core sections.

In the Camp Century core some particles of unusual composition were found in the lowest 200 meters. Unusually high concentrations of Sn, Cu, and Mo were found as illustrated in Fig. 23 (a), (b), and (c). These particles (See Appendix D) may be windblown material from some atypical source area; particles of similar composition were not found in the upper 1200 meters of this core nor in the Byrd Station core. In addition, at the same depth where the atypical particles first appear there is a drop in particle concentration by a factor of 2 from the concentration measured 20 meters above this section and a corresponding change in the mean oxygen isotope ratios by 4 o/oo. Another, and perhaps more likely source for these particles is the bedrock over which the Greenland Ice Sheet moves. If this is the case, it raises serious questions about the ability to date the lowest layers of the Greenland core, as was discussed in detail in the previous chapter.

The occurrence of particle layers with unusual compositions over short depth intervals is potentially important. A detailed dating and elemental analysis of one deep core would allow the establishment of time marker horizons which could then be used to date horizons in other ice cores without doing detailed analyses.

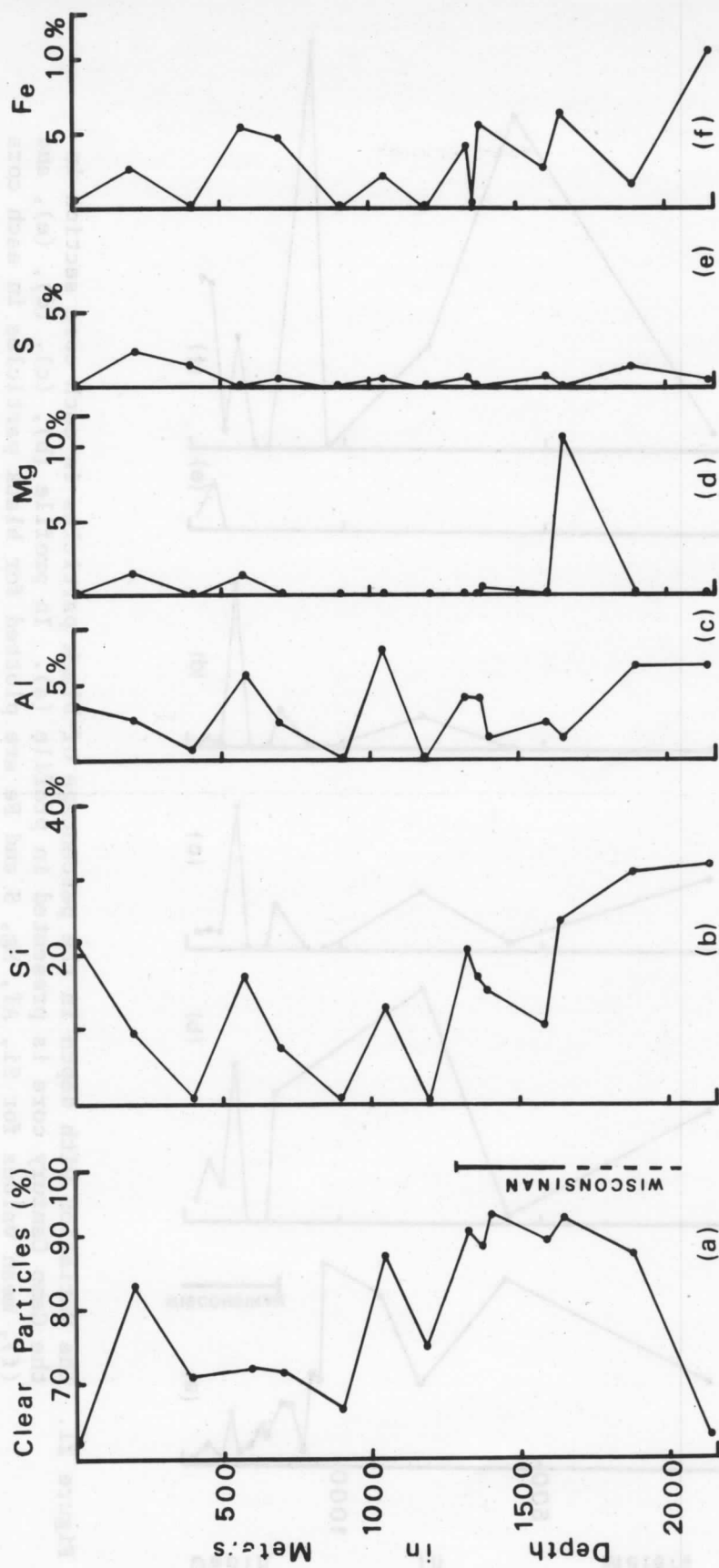


Figure 20. The variation with depth in the percentage of clear particles in each core section in the Byrd core is presented in profile (a). In profile (b), (c), (d), (e), and (f) mean values for Si, Al, Mg, S and Fe are plotted in clear particles for each core section.

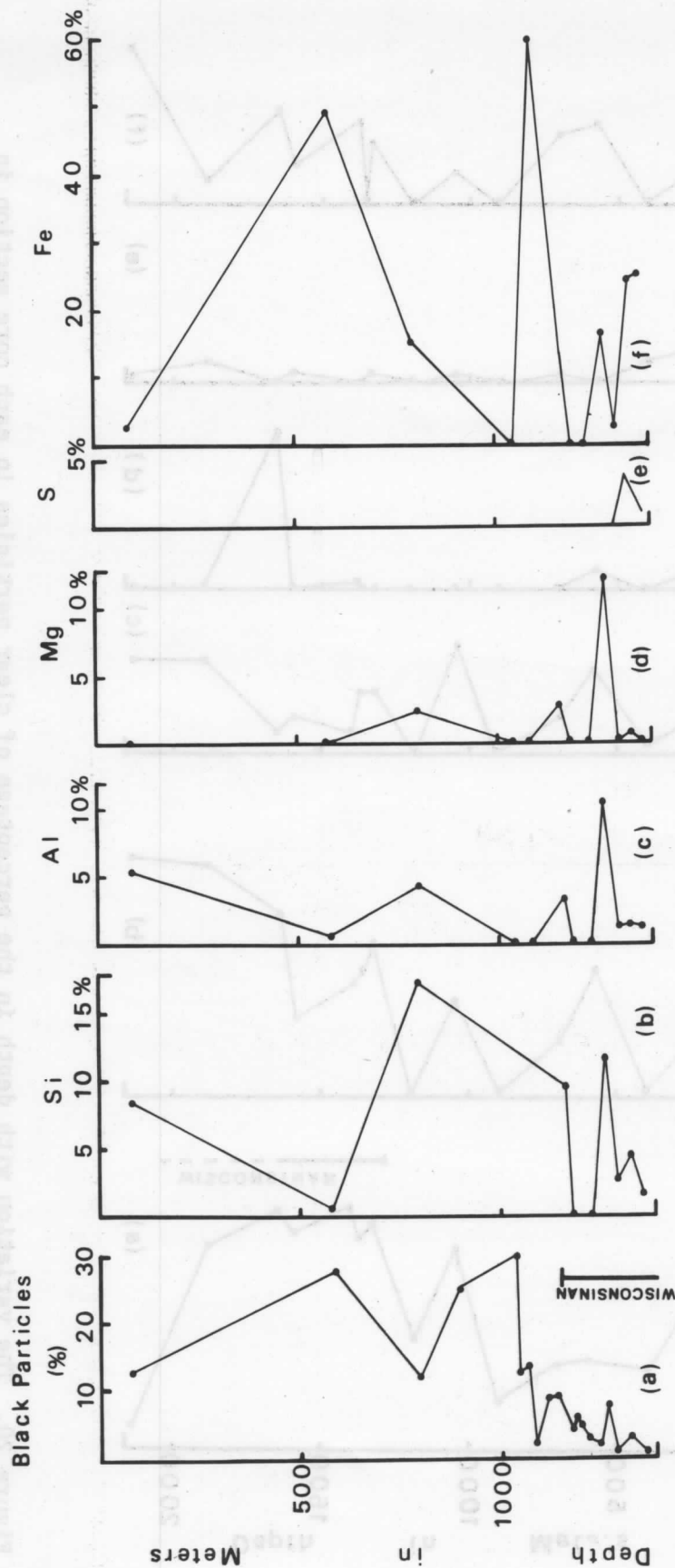


Figure 21. The variation with depth in the percentage of black particles in each core section in the Camp Century core is presented in profile (a). In profile (b), (c), (d), (e), and (f), mean values for Si, Al, Mg, S and Fe are plotted for black particles in each core section.

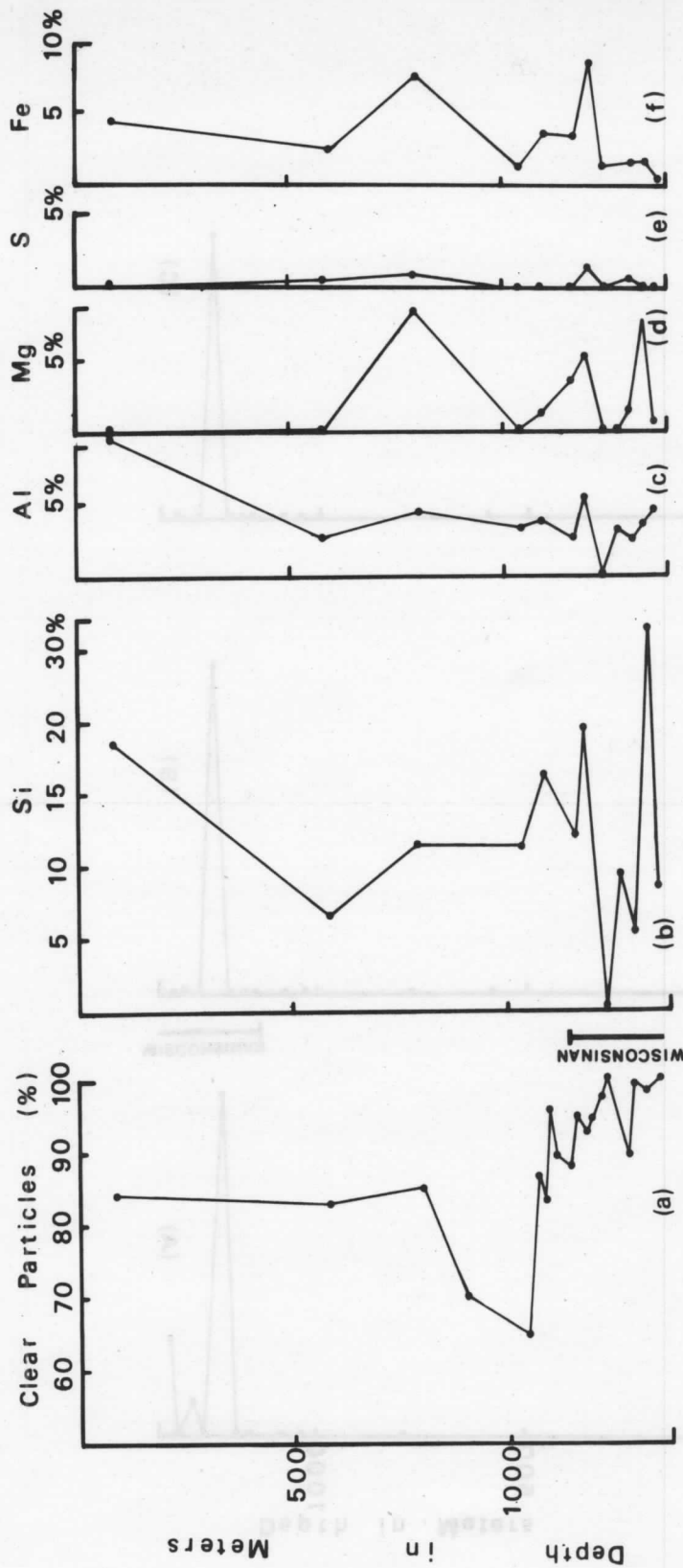


Figure 22. The variation with depth in the percentages of clear particles in each core section in the Camp Century core is presented in profile (a). In profile (b), (c), (d), (e), and (f) mean values for Si, Al, Mg, S, and Fe are plotted in the clear particles for each section.

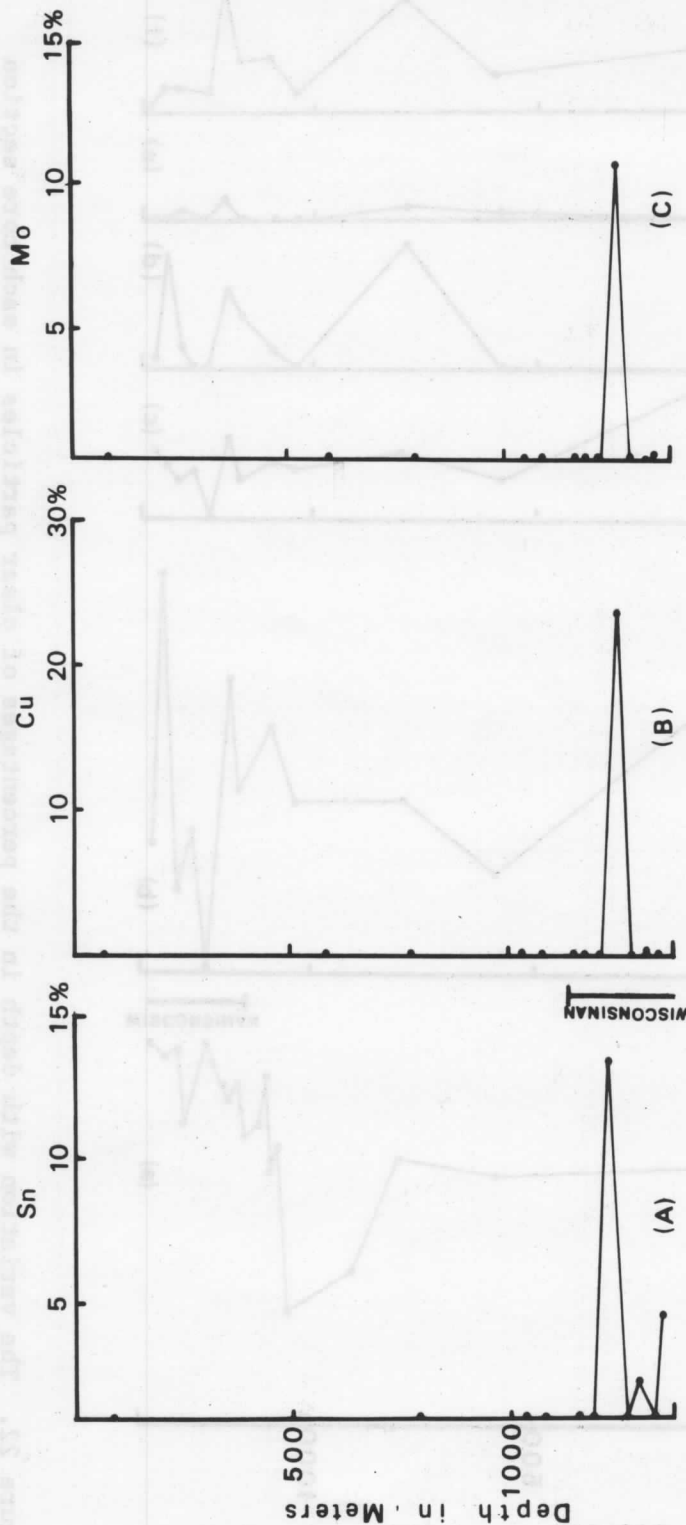


Figure 23. Profiles (A), (B), and (C) illustrate variations in concentrations of Sn, Cu, and Mo in the Camp Century core. The occurrence of such particle layers with unusual compositions could serve as marker horizons.

## THE RELATIONSHIPS AMONG DUST, OXYGEN ISOTOPES AND CLIMATE

### Microparticle Variations Over Millennial Time Intervals

The mean concentration and size distribution of particles in each core section can be compared to determine the variations over millennial time intervals (Thompson, 1975b). In Fig. 24 (a), (b), and 25 (a), (b), 26 (a), (b), and 27 (a), (b) the microparticle and oxygen isotope variations with depth in both ice cores are compared (See Table 2 and 3 for data). In the Byrd Station ice core the major variations of microparticle content (Fig. 24a and 25a) correspond closely with the major variations of the  $\delta^{18}O$  values (Figs. 24b and 25b) as determined by Johnsen and others (1972). The largest microparticle concentrations occur in those sections of core with the lowest  $\delta$  values. The same general relationship is found for the Camp Century ice core (Figs. 26 (a), (b) and 27 (a), (b)). The large increase in microparticles within the two cores from different hemispheres and within sections of comparable age is remarkable and strongly suggests either a higher global atmospheric particle content or a decrease in the snow accumulation rate or both during the Wisconsin. The lack of a tropical ice core containing a climatic record over the last 30,000 years prevents establishing this as a "fact". In both deep ice core particle profiles the last large peak in the microparticle concentrations occur before what has traditionally been interpreted as the end of the glacial stage as defined by the  $\delta$  profile.

### Dust, Oxygen Isotopes and Climate

Atmospheric particles may be divided into three main categories according to radius as suggested by Junge (1963):

Aitken particles	less than $0.1\mu m$
Large particles	$0.1$ to $1\mu m$
Giant particles	greater than $1\mu m$

Junge (1963) found that the predominant size of natural tropospheric aerosols ranged in diameter between  $0.1$  and  $20\mu m$  and that particles with diameters less than  $1.0\mu m$  have the longest residence time; therefore, they are most likely to have the greatest effect upon incoming radiation and thus on world temperature. So as to restrict the analyses to those particles capable of being globally distributed, the larger particles have been disregarded in the following discussion and attention is focused upon the smaller diameter particles:  $0.65\mu m$  to  $0.82\mu m$  from the cleanest 10 percent of samples cut from each section. Presumably these samples consist of the snow accumulated during periods of reduced concentrations of locally derived material. Figs. 28 and 30 illustrate

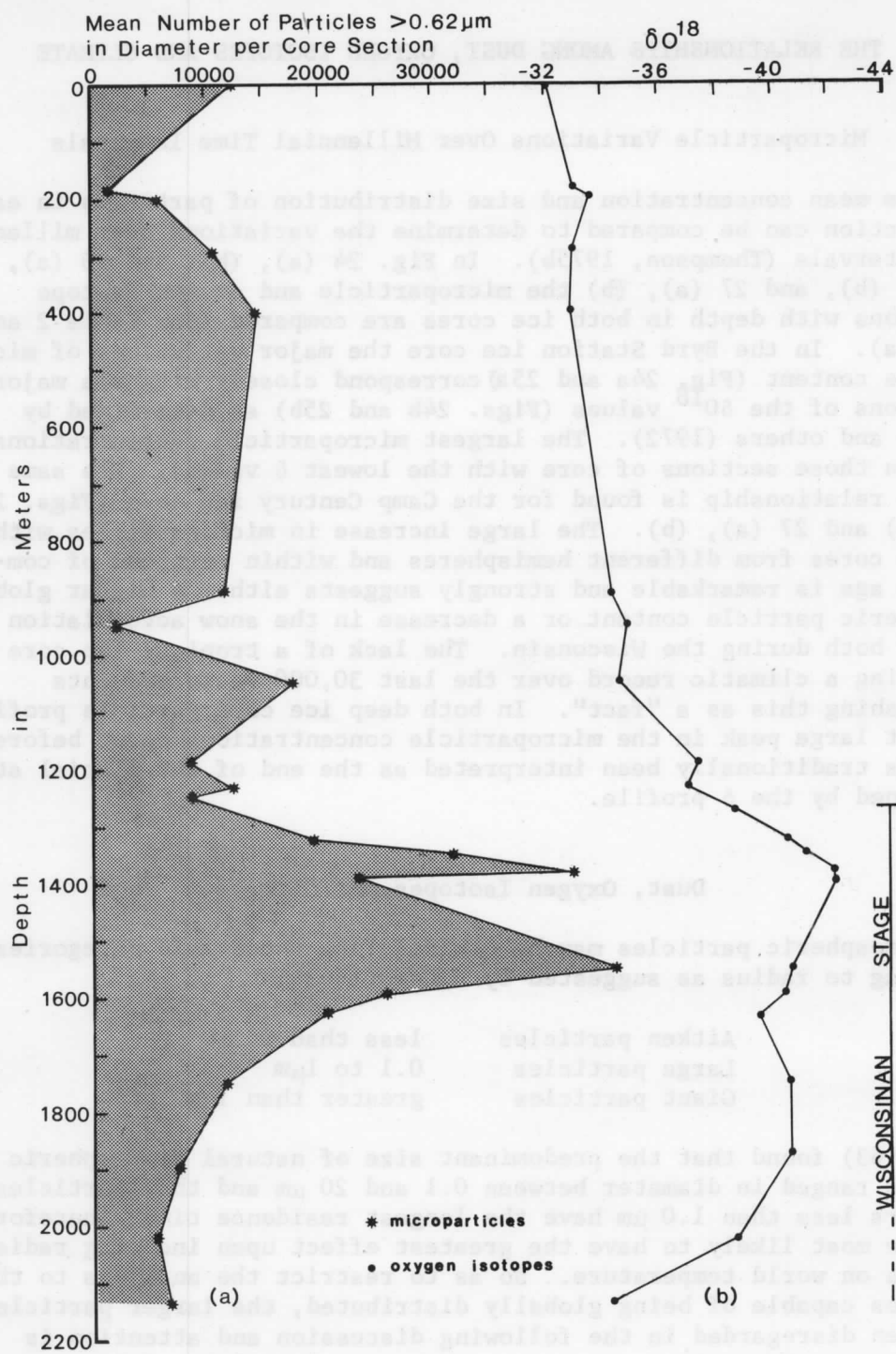


Figure 24. Data from the Byrd Station, Antarctica, deep ice core. (a) Profile obtained by plotting the mean number of particles greater than 0.62 μm in diameter per core section. (b) Profile of mean δO¹⁸ values (Epstein and others, 1970) from the core sections analyzed in Figure 24 (a).

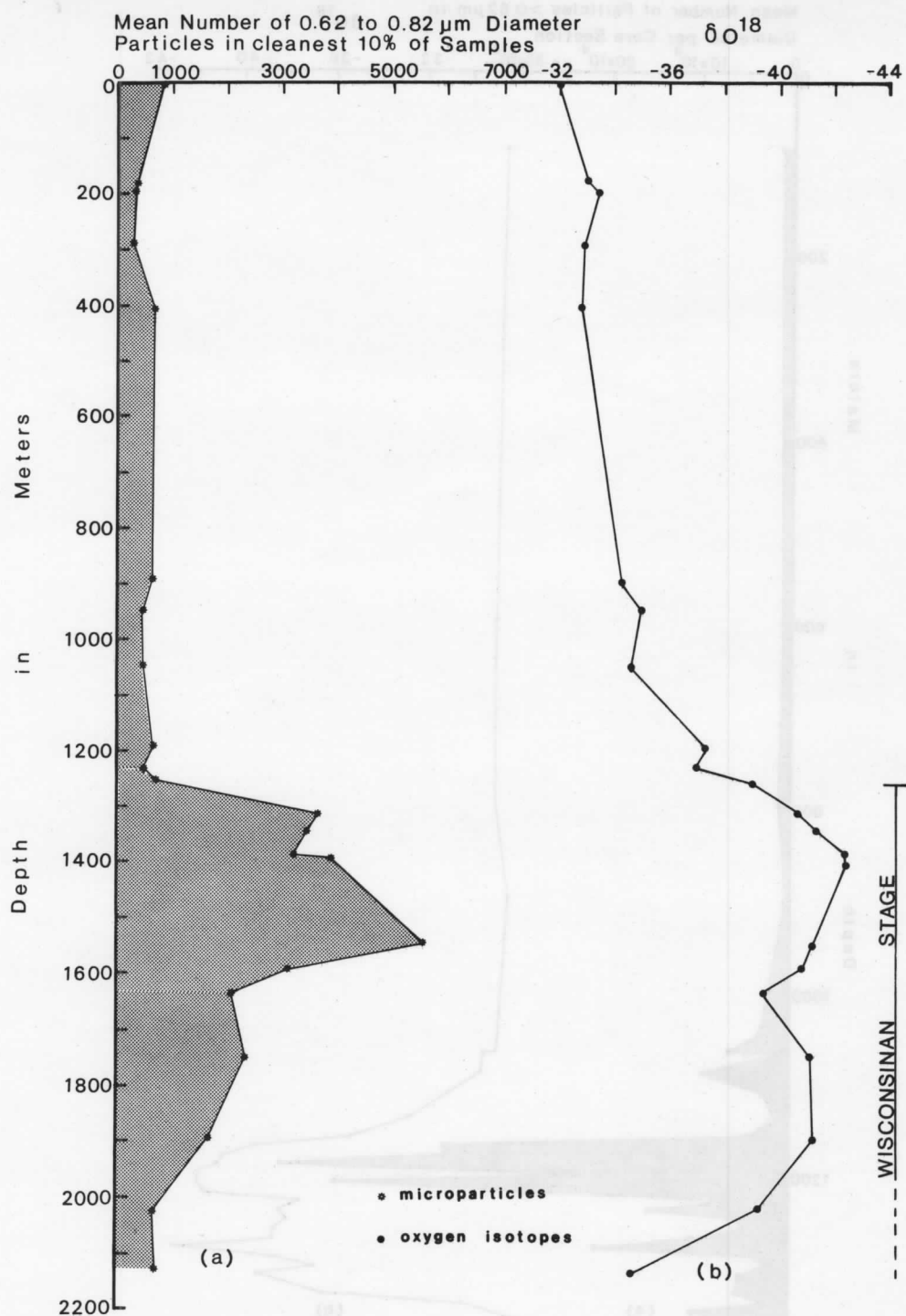


Figure 25. Data from the Byrd Station, Antarctica, deep ice core. (a) Profile obtained by plotting the mean number of 0.62 to 0.82  $\mu\text{m}$  diameter particles in the cleanest 10% of each core section. (b) Profile of mean  $\delta\text{O}^{18}$  values (Epstein and others, 1970) from the core sections analyzed in Figure 25 (a).

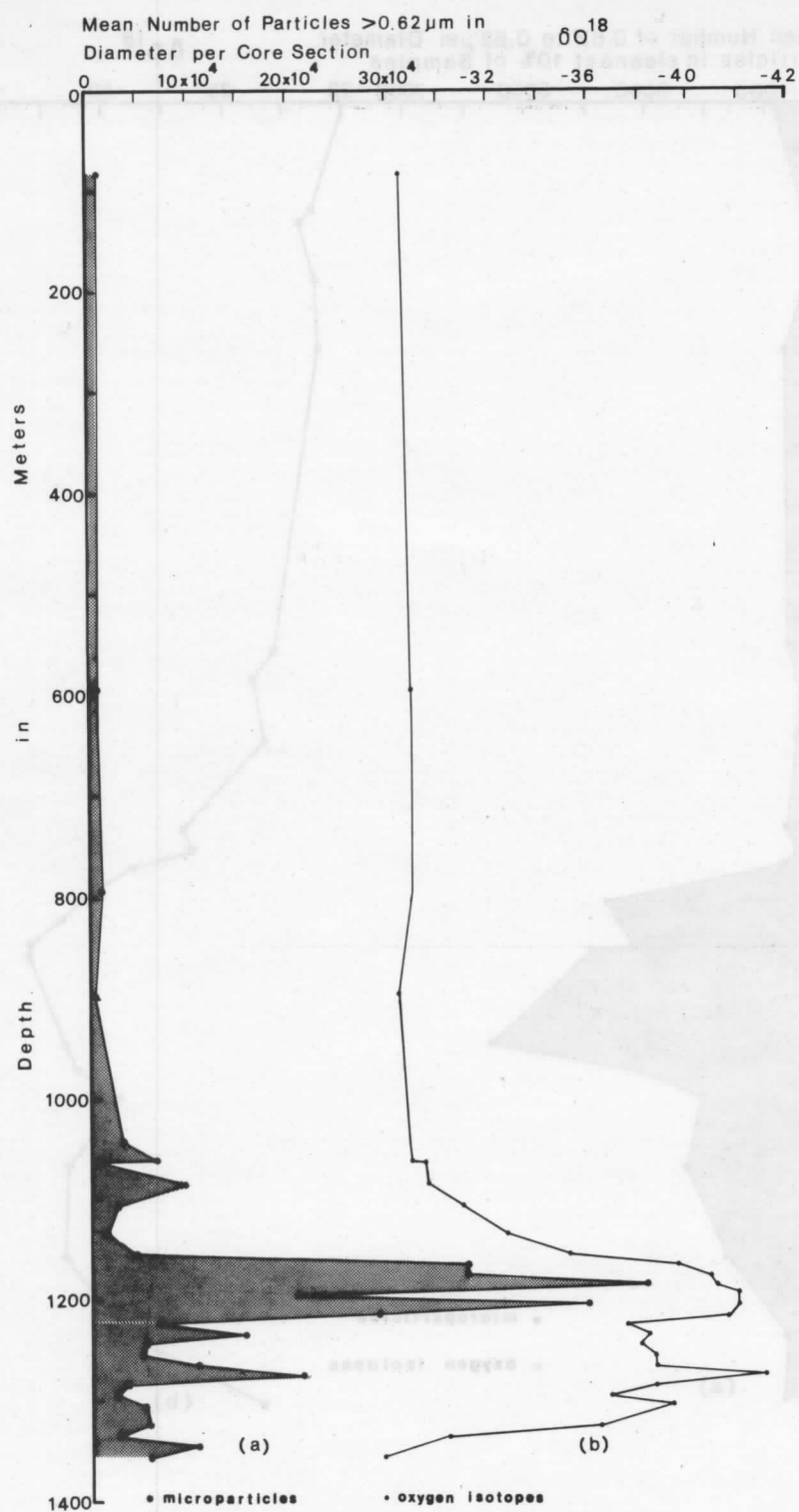


Figure 26. Data from the Camp Century, Greenland, deep ice core. (a) Profile obtained by plotting the mean number of particles greater than  $0.62\mu\text{m}$  in diameter per core section. (b) Profile of mean  $\delta\text{O}^{18}$  values (Dansgaard and others, 1971) from the core sections analyzed in Figure 26 (a).

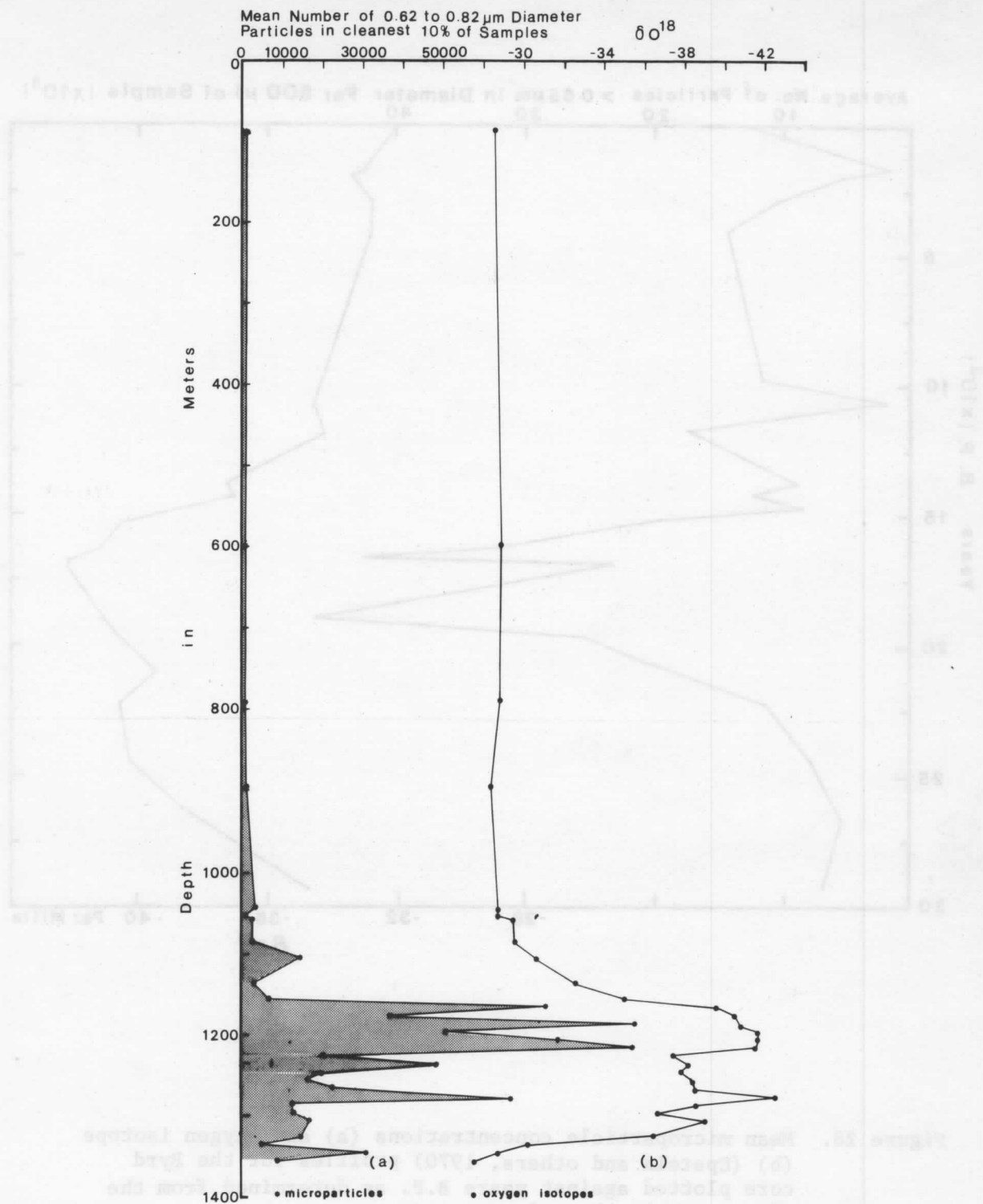


Figure 27. Data from the Camp Century, Greenland, deep ice core. (a) Profile obtained by plotting the mean number of 0.65 to 0.82  $\mu\text{m}$  diameter particles in the cleanest 10% of each core section. (b) Profile of mean  $\delta\text{O}^{18}$  values (Dansgaard and others, 1971) from the core sections analyzed in Figure 27 (a).

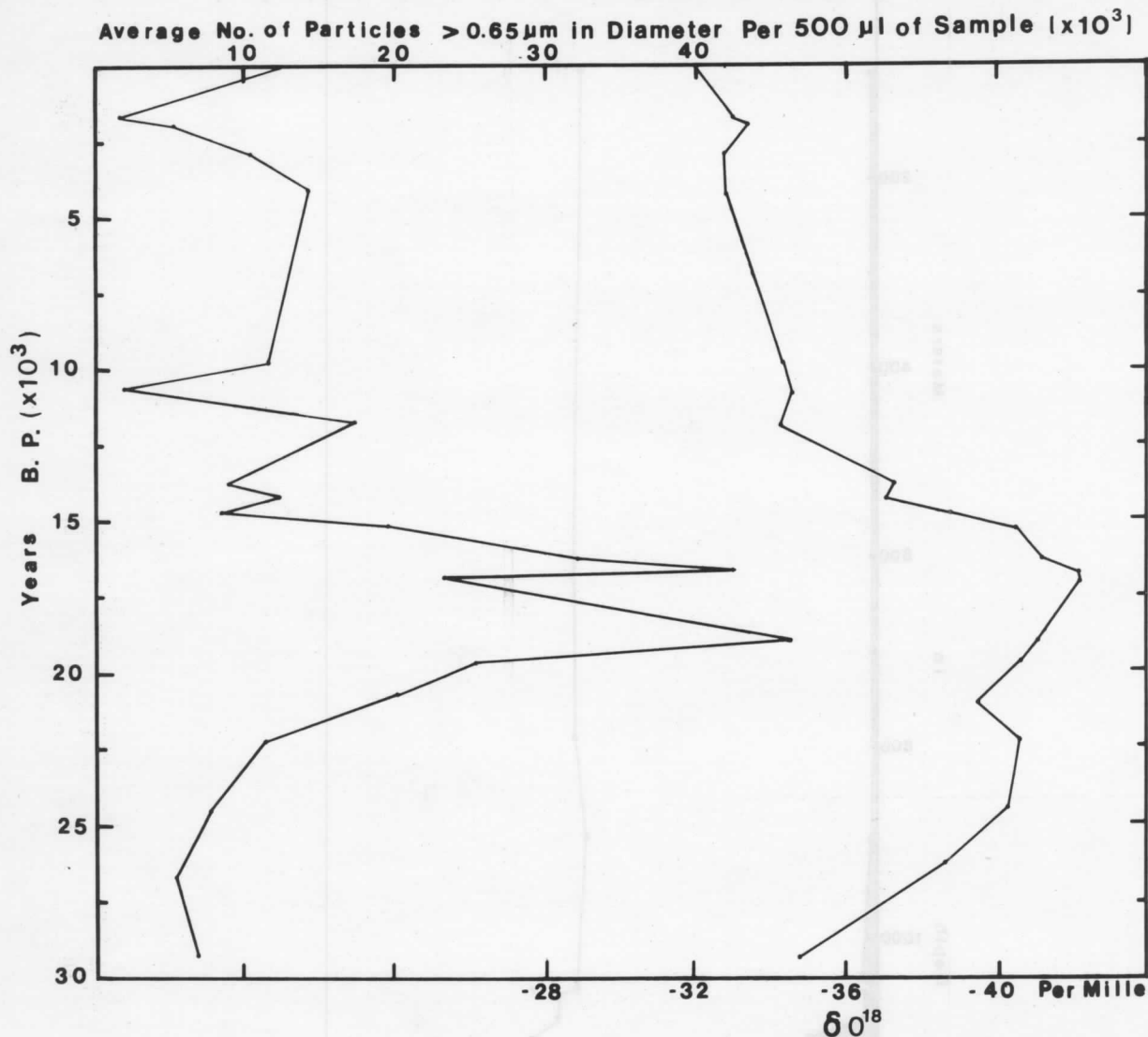


Figure 28. Mean microparticle concentrations (a) and oxygen isotope (b) (Epstein and others, 1970) profiles for the Byrd core plotted against years B.P. as determined from the microparticle chronology.

the relationship of total particle concentrations and oxygen isotopes over the past 30,000 years for the Byrd Station and Camp Century cores, respectively, while Figs. 29 and 31 show the same relationship for the small diameter particles. The microparticle chronology is used for the Byrd core and the oxygen isotope chronology is used for the Camp Century core for reasons outlined previously.

The comparison of microparticle variations and  $\delta$  values yields a relationship over millennial time intervals; increased concentration of particles are correlated with more negative values. It is evident from Figs. 29 and 31 that the non-soluble particle content of the snow deposited on the continents of Greenland and Antarctica during the Wisconsin Stage was 100 times and 4 times greater than the mean Holocene values.

It is possible to differentiate those Byrd core particles deposited in the Wisconsin Glacial Stage from those deposited in the Holocene based on the elemental composition and morphology as discussed previously. Particles deposited in Wisconsin ice tend to be angular and contain relatively more Si and K than the more rounded Holocene particles (See Plates III through VI in Appendix C).

For the Camp Century core particles there is no marked stratigraphic change in the elemental composition of the particles or in the morphology between Holocene and Wisconsin ice as previously discussed. The majority of the Camp Century core particles are flat plate-like, with Wisconsin particles containing greater concentrations of Ca, Ti, Sn, Mo, and Cu (See Plates VII through X in Appendix D).

Consideration of the elemental and morphological data from these cores as presented earlier indicate that the majority of the particles from the Wisconsin sections of the Byrd core are most probably volcanic in origin whereas most of the particles in the Camp Century core are more likely to be wind-blown terrestrial material. The number of volcanic particles present in the Wisconsin sections of both deep ice cores is important in that they may be a causative factor of climatic change. The importance of the volcanic fraction of the particles present in the Wisconsin sections of the Camp Century core is difficult to ascertain as the total particle concentrations are quite high, as was illustrated earlier. If volcanic particles had the same abundance in the Camp Century core as they do in the Byrd core only one of twenty would be volcanic.

#### Discussion

The nature of atmospheric circulation patterns during the last glaciation is a subject of considerable debate but one of importance in explaining variations in dust deposition on the polar ice sheets. Wind directions during the last glacial phase in central Europe, deduced from

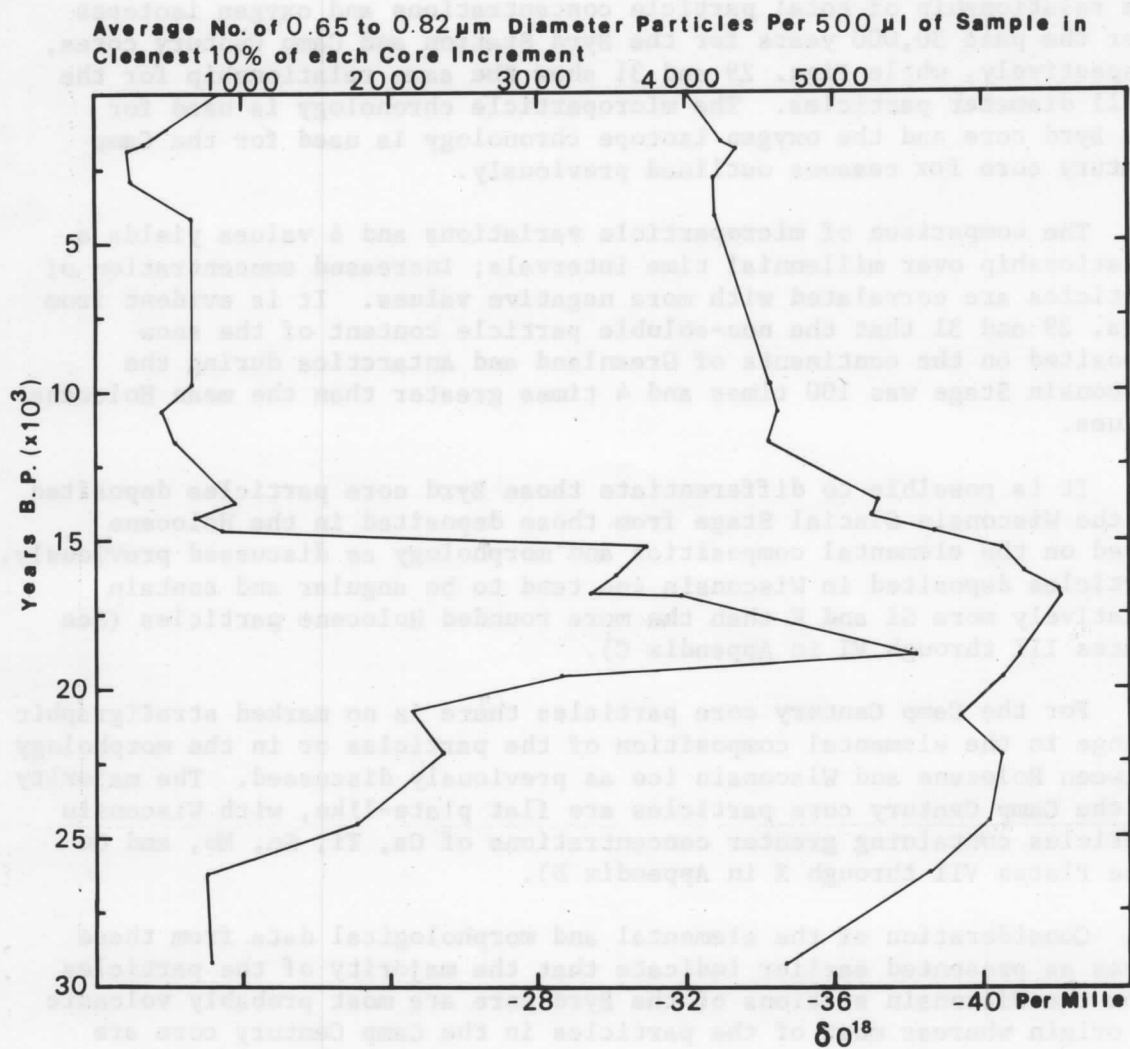


Figure 29. Average number of 0.65 and 0.82  $\mu\text{m}$  diameter particles in the cleanest 10% of each core section (a) and oxygen isotope (b) (Epstein and others, 1970) profiles for the Byrd core plotted against years B.P. as determined from the microparticle chronology.

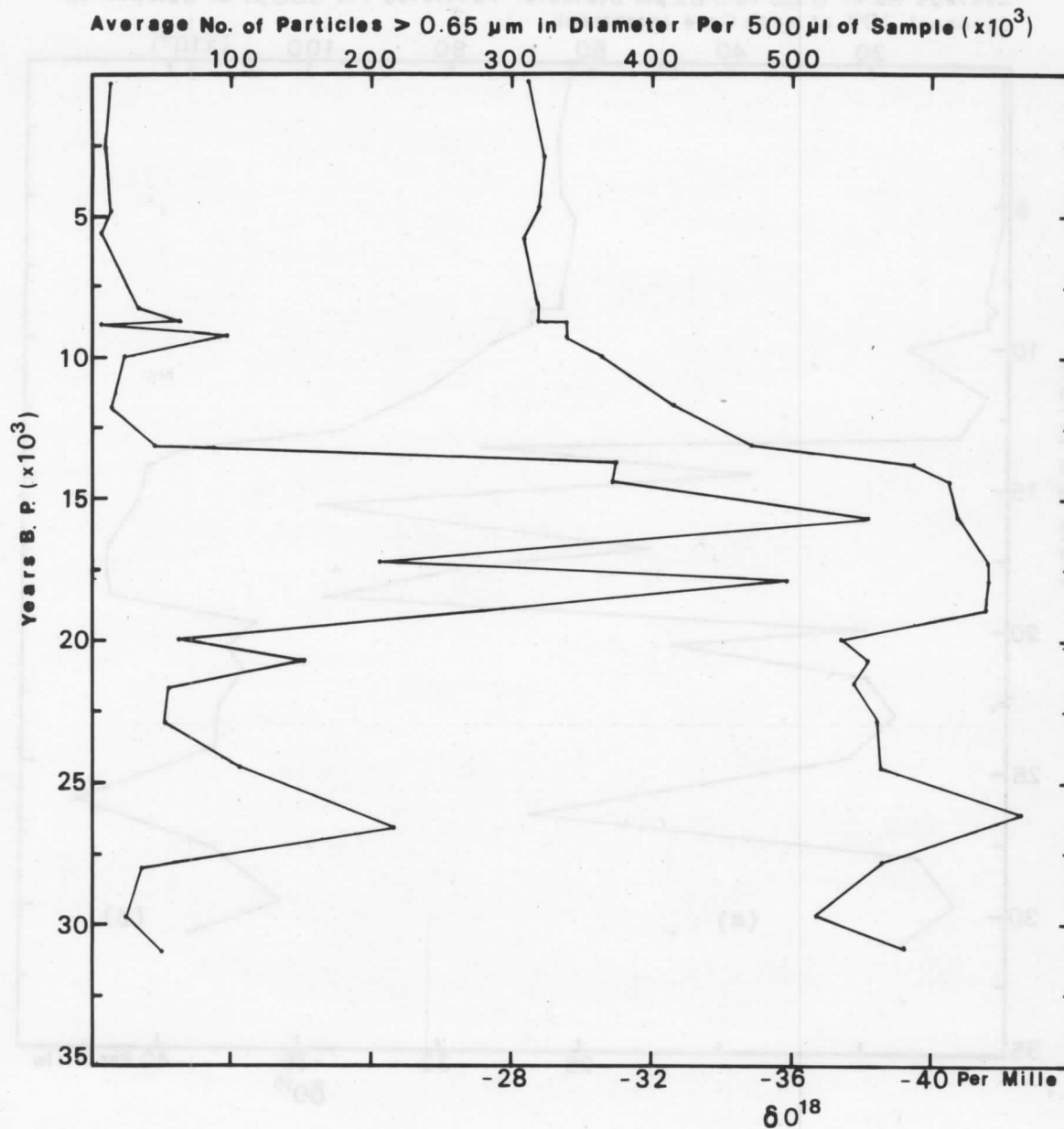


Figure 30. Mean microparticle concentrations (a) and oxygen isotope (b) (Dansgaard and others, 1971) for the Camp Century core plotted against years B.P. as determined by the oxygen isotope chronology.

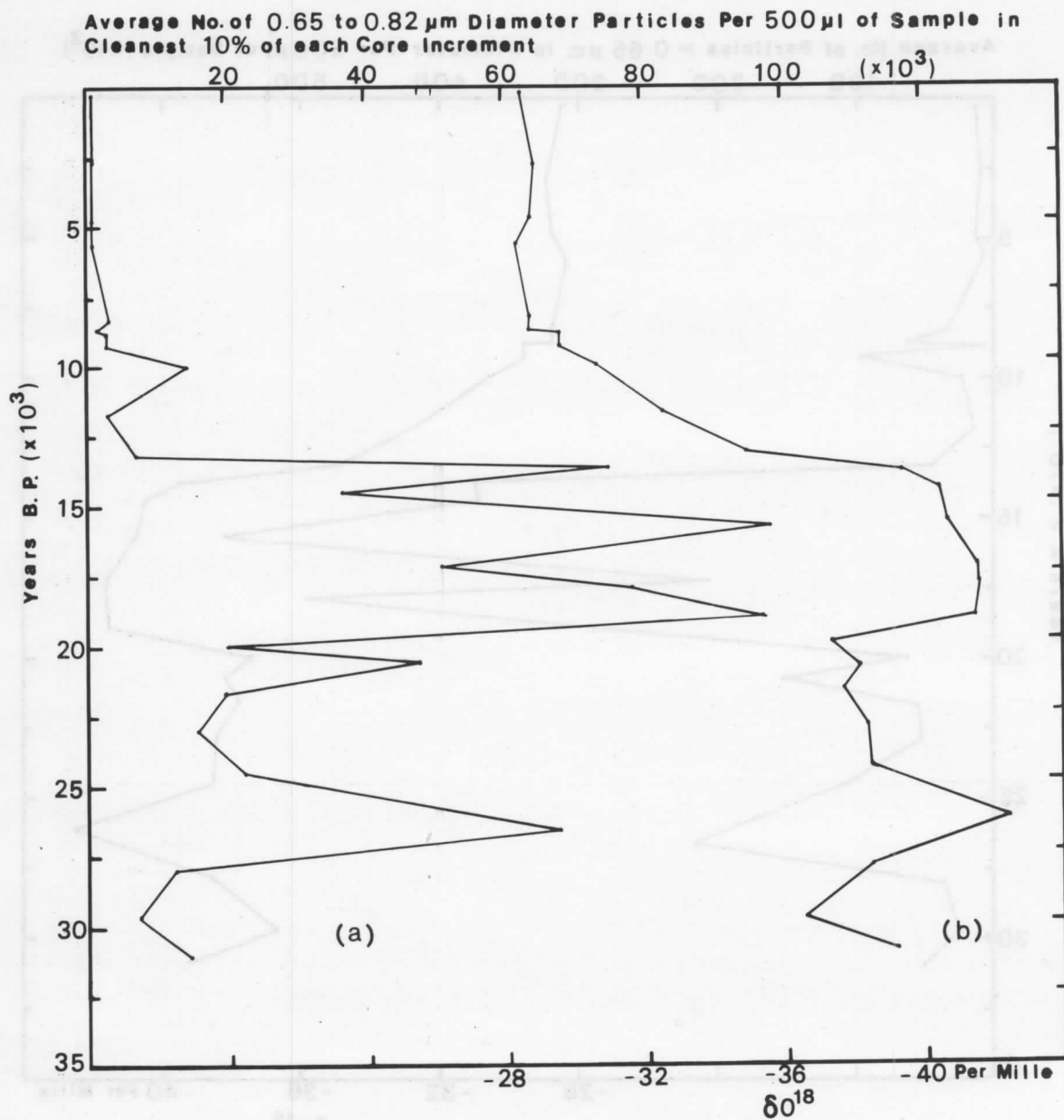


Figure 31. Average number of 0.65 to 0.82  $\mu\text{m}$  diameter particles in the cleanest 10% of each core section (a) and oxygen isotope (b) (Dansgaard and others, 1971) for the Camp Century core plotted against years B.P. as determined from the oxygen isotope chronology.

the structure of migrating dunes formed at that time, indicate that during the last glacial stage in Europe the prevailing wind direction was the same as at the present time (Frenzel, 1973). The radiation and reflection conditions associated with snow and ice cover today suggest that anticyclone circulation systems should have been associated much of the time with areas of ice and snow cover during glacial conditions. Maarleveld and van der Schans (1961) carried out careful inclination measurements on the direction of gradient dip of sand horizons which were deposited by wind during the peak of the last cold period. During the period of maximum ice cover they found the prevailing wind in the Netherlands blow out of the north. With the retreat of the ice the wind backed to northwest and finally to the west southwest.

In North America the distribution of loess indicates a dominant westerly wind. Indeed, Goldthwait (1968) has indicated that in Ohio the areas of deepest loess, over 50 inches, are located just east or southeast of main river courses and thus indicate a prevailing westerly wind. In general, for the middle latitudes of North America, one third as much loess occurs on the west sides of the valleys as on the east, which indicates wind direction changes comparable to those accompanying the passing of high and low pressure weather systems such as we have at present.

Since the areas of maximum thickness of loess occur southeast of the broader expanses of the valleys Fry and Willman (1973) thought the loess was deposited mainly during the fall and winter when northwesterly winds dominated. Also in the fall less flooding of the outwash surfaces in the valleys permitted more drying and more wind erosions of silt deposits.

Daugherty (1968) suggested that the atmospheric circulation which prevailed over North America during glacial stages was similar to a strong present-day winter circulation. The prominent circulation features that influenced North American climates according to Daugherty were an extensive anticyclonic circulation over the ice sheet, a stronger mid-latitude westerly flow, well-developed low-pressure systems in the North Pacific and North Atlantic, and a strengthening of the cellular component of circulation. Indeed, one might expect that the dominant surface wind directions were similar to those indicated by the arrows in Fig. 32. The strong meridional temperature gradient and the more intense atmospheric circulation might have increased the frequency of mid-latitude cyclonic activity and storminess, which would have directly affected the quantity and the distance dust would have been carried.

Fig. 33 is a sketch (Flint, 1971) showing compiled glacial invasions of the territory between the Appalachians and the Mississippi River during Wisconsin time. Interesting relations between ice advance at

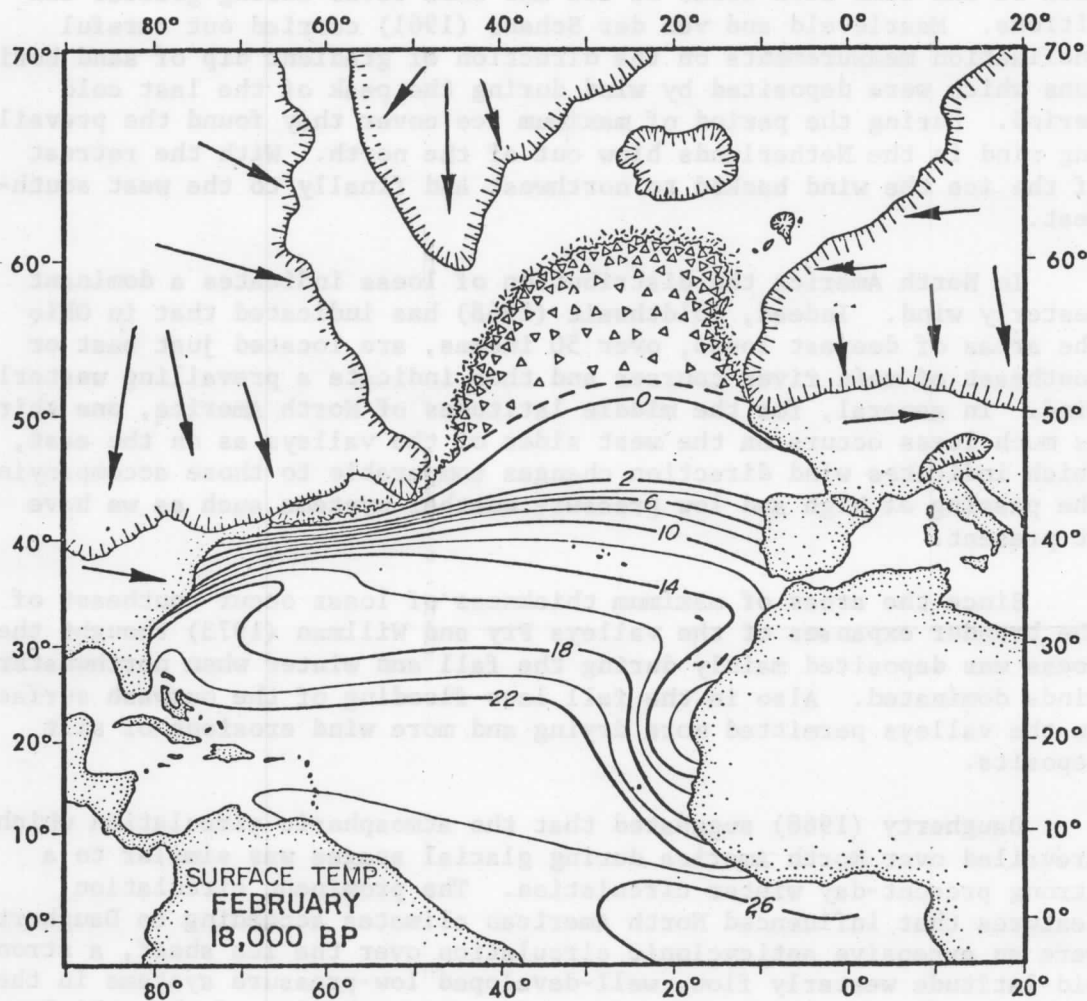


Figure 32. Surface water isotherm map of February, 18,000 B.P. The dashed isotherms are interpretative. The major continental ice masses are delineated by hatched borders, the permanent pack ice by granulate borders. Glacial shoreline is drawn to a sea level of -85 meters from present (McIntyre, 1975). The arrows indicate dominant surface wind direction inferred from data presented in this report.

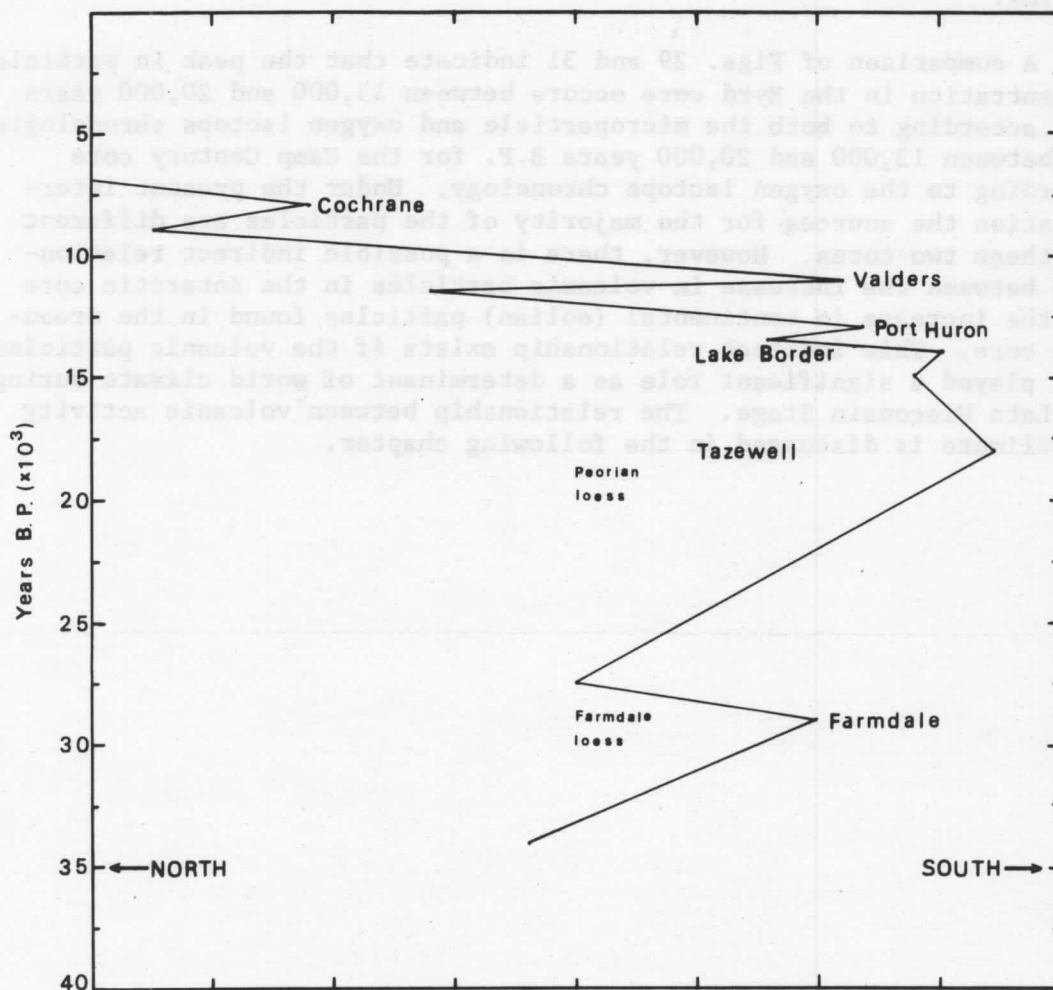


Figure 33. Sketch showing glacial invasions of the territory between the Appalachians and the Mississippi River during the Late Wisconsin Stage. Glacial sediments and loess are labelled. Pattern becomes poorly controlled toward the base and is partially conjectural (compiled by Flint, 1971).

the ice margin, and variations in dust deposition on the Greenland Ice Sheet can be seen by comparing Figs. 31 and 33. It appears that the times (oxygen isotope chronology) of greatest particle deposition of the Ice Sheet are correlated with times of ice advance at the ice margins.

A comparison of Figs. 29 and 31 indicate that the peak in particle concentration in the Byrd core occurs between 13,000 and 20,000 years B.P. according to both the microparticle and oxygen isotops chronologies, and between 13,000 and 20,000 years B.P. for the Camp Century core according to the oxygen isotope chronology. Under the present interpretation the sources for the majority of the particles are different for these two cores. However, there is a possible indirect relationship between the increase in volcanic particles in the Antarctic core and the increase in continental (eolian) particles found in the Greenland core. This indirect relationship exists if the volcanic particles have played a significant role as a determinant of world climate during the Late Wisconsin Stage. The relationship between volcanic activity and climate is discussed in the following chapter.

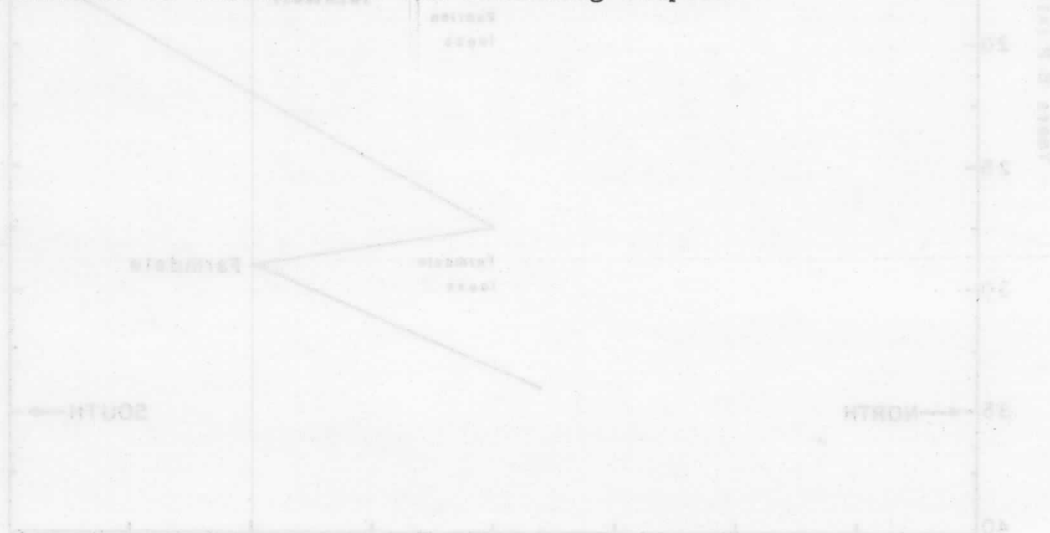


Figure 33. Sketch showing glacial invasions of the territory between the Appalachians and the Mississippi River during the Late Wisconsin Stage. Glacial sediments and loess are labeled. Pattern becomes poorly controlled toward the base and is partially conjectural (compiled by Viles, 1971).

# IS THERE A RELATIONSHIP BETWEEN VOLCANIC ACTIVITY AND CLIMATIC CHANGE?

## General

The Earth undergoes what appear to be major periodic changes in climate for which there must be a cause or combination of causes. The only readily available clues to past climatic change are those recorded on and in the land surfaces, the ocean floors, and particularly in the polar ice sheets.

Certain initial conditions are probably prerequisites for glaciations, such as a favorable distribution of land and ocean bodies. The land and ocean distribution during the Pleistocene, and at present, is sufficient for glaciation, but probably not the optimum for glaciation, since there was a lack of extensive glaciation in the northeast part of Siberia during the last glacial period. It also must be pointed out that the lack of glaciation in any given geological period does not mean that the earth has not undergone significant changes in climate. From the changes in successive floras and faunas in the geological record it is clear that along with relatively slow changes rather abrupt changes have occurred repeatedly. These may well be indicative of major changes in environment brought about by climatic changes. Many theories have been proposed to account for climatic change. In this chapter the relationship between explosive volcanic activity and climatic change is discussed.

## Theory

The concept that increases in atmospheric dustiness from volcanic activity may affect world climate is not new. P. and E. Sarasin (1901) suggested that the lowered surface level temperatures essential to glaciation were due to reduced surface insolation caused by the absorption of solar radiation in high level volcanic dust clouds. This concept was further developed by Humphreys (1929) and Wexler (1953) among others. More recently workers have argued that the effect upon surface temperatures of a dust veil when it is distributed in the stratosphere, is theoretically one of cooling (Dyer and Hicks, 1968; Lamb, 1970; Mitchell, 1961, 1971; Newell, 1970).

When a volcanic dust veil is present in the atmosphere, direct radiation received at the earth's surface decreases with the mean elevation of the sun. This means that with a given optical mass for the atmosphere direct radiation decreases to a greater extent in higher latitudes than in lower latitudes, and more strongly in winter than in summer. The effect is even greater for total solar radiation (direct plus diffuse, Budyko, 1974). Thus stronger zonal circulation patterns

due to greater meridional temperature gradients are expected in the presence of an extensive volcanic dust veil.

A dense layer of dust particles can cause a net local heating or cooling depending upon such factors as their height above the surface, their composition and size distribution as well as the value of the albedo of the underlying surface. A study of aerosols in the air over the South Pole Station (Shaw, 1975) indicated that aerosols over central Antarctica may cause a net heating of the earth's atmosphere on the order of nearly 0.1°C. The effects of such atmospheric heating on surface temperatures have not been determined.

Today greater amounts of precipitation occur during the summer months than in the winter in both the Northern and Southern Hemispheres (Lamb, 1972). This condition is due mainly to the ability of warm air to carry more moisture than cold air. Thus, when explaining the initiation of glaciation it is necessary to consider the paradox that large volumes of water from oceans were transported to the land to become ice sheets during a period when mean world temperature was lower than today. One approach to this problem is to assume a constant total solar influx, modulated by changes in the dustiness or turbidity of the atmosphere produced by the varying intensity of explosive volcanic eruptions. This condition could produce a climatic regime marked by periods of extreme cold, followed by relatively warmer intervals during times of low volcanic activity. Thus to produce a unit drop in mean surface temperature it would take a greater magnitude explosive eruption in the initial stages of glaciation when the volume of warm water is large than in the waning stages when the oceans are reduced in volume and mean temperature. A sequence of events which lead into and out of a glacial stage are postulated below.

Intense explosive-type volcanism associated with orogenic uplifting would initiate high latitude surface cooling. The atmospheric circulation systems would be most intense in winter, produced by a strong latitudinal temperature gradient. The initial effect of the oceans would be to work as a buffer providing warmth and moisture, and producing snowfall at high latitudes and elevation. The summer temperatures would be lower so that each year the preceeding year's snowfall would last longer into the summer months thus, increasing the effective albedo of the earth. This general cooling would be broken by periods of reduced volcanic activity during which time the land temperatures in the high latitudes and altitudes undergo some recovery, but the cooler oceans permit only slow recovery. If this interval is followed closely by renewed explosive type volcanic activity the process is repeated. Following each succeeding explosive volcanic event the mean world temperature is reduced, resulting in a lowering of the world snowline. The lower snowline would lead to further cooling by increasing the surface albedo. Thus, the glacial stage would

deepen. The dampening effect of the oceans on mean world temperature would diminish as more oceanic water is tied up in the ice sheets and the surface area of the oceans decreases. The coldest and driest climatic conditions occur at this time. The glacial stage would end with a final flurry of volcanic activity releasing pressures within the earth's mantle and initiating a period of low volcanic activity and glacial retreat. According to this theory the climatic optimum would occur soon after entering the interglacial.

#### Supportive Evidence

There have been events which demonstrate the effects volcanic dust in the atmosphere can have upon climate. The significantly cooler years of 1783-85, followed the great explosion of Asama, Japan, in 1783. "The year without a summer," 1816, followed the eruption of Tomboro, in the Little Sunda Islands, which ejected so much ash that "for three days there was darkness at a distance of 300 miles" (Humphreys, 1929). In more recent work reductions of direct radiation received at the earth's surface have been measured following the 1912 eruption up Mount Katmai in Alaska (Budyko, 1974) and following the Mount Agung eruption on Bali in 1963 (Dyer and Hicks, 1968).

The microparticle studies on the Byrd Station and the Camp Century deep ice cores presented in the previous chapters illustrate that the particle concentration in what traditionally has been interpreted as Wisconsin ice is much higher than in Holocene ice. The small (0.65 to 0.82  $\mu\text{m}$  diameter) microparticle content of ice deposited during the Wisconsin glacial stage in the Byrd Station core is four times, and that in the Camp Century core is one hundred times as great as Holocene concentrations (Fig. 25 and 27). In both cores the increase is gradual throughout most of the Wisconsin ice and reaches a maximum at the close of the Wisconsin stage corresponding to the maximum decrease in paleotemperatures as indicated by stable oxygen isotope ratios. In the Byrd core this microparticle increase is suspected to be the result of an increase in volcanic particles while in the Camp Century core the increase in particles is most likely due to an increase in wind velocities or aridity or both. In Fig. 27 the transition from the Wisconsin high particle concentrations to the Holocene low concentrations occurs in less than 100 years. This suggests that the conditions characterizing the climate of the Wisconsin glacial stage changed quite suddenly. Volcanic activity is one of the few processes known to fluctuate dramatically over a short period of time and which is capable of causing climatic change.

The possibility that the onset of glacial stages are associated with volcanism was initially rejected because volcanic ash layers were lacking in deep sea cores. As early as 1929, Hymphreys argued that deposits of volcanic dust in ocean sediments would not be expected since

his calculations showed that an annual emission of dust necessary to decrease the direct solar radiation by 20 percent would, after 100,000 years, accumulate only to a layer 0.02 inches thick.

However, recent literature indicates that there is evidence in deep sea cores of volcanism before the onset of glacial stages (Menard, 1971; Kennett and Huddleston, 1972). More recently Kennett and Thunell (1975) found evidence of a global increase in Quaternary explosive volcanism. In sediment cores obtained by the Deep Sea Drilling Project volcanic ash layers and zones were shown to occur at a much higher frequency in the Quaternary. If it is assumed that the stratigraphic interpretation and dating of the majority of these cores is correct, then the increased Quaternary volcanism coincides approximately with that episode of the Cenozoic marked by major and rapidly fluctuating climatic change. These data indicate an increase in worldwide volcanism greater than the 40 percent within several thousand years which was calculated by Chappel (1973) as necessary for glaciation. It is interesting to compare the deterioration of climate in central Europe (Woldstedt, 1954) with the increase in volcanism as determined from the deep sea cores over the past 20 million years. Such a comparison shows that the decline in central European temperatures was associated with a gradual increase in volcanism.

### Cycles

Since glaciations appear to be cyclic, in order to use volcanism as a cause, a mechanism must exist whereby there is a periodicity in explosive volcanic activity. Fuchs and Paterson (1947) were two of the early workers to argue the relationship of rhythms in volcanism and orogeny to climate. Damon (1971) employed a statistical treatment of the late Mesozoic and Cenozoic magmatic events coupled with analysis of paleogeographic data to demonstrate that orogenesis is periodic with a wavelength of  $36 \pm 11$  million years. He also suggested that epeirogenic uplift of continents is a periodic phenomenon with a wavelength of  $250 \times 10^6$  years and is related to continental orogeny. These two cycles could be related to glacial periods and glacial epochs respectively. Many other authors have suggested a connection between volcanism, orogeny, and plate tectonics, (Hays and others, 1972; Stille, 1924; Brookfield, 1971). Naturally, orogeny and uplift may not always be associated with explosive volcanism, in which case the climatic effect would only be regional.

### Discussion

Although the concept of volcanic activity as a possible cause of climatic change is centuries old, techniques for determining relationship between volcanism and climatic change are just now becoming

available. Therefore, the question as to whether there is a relationship between volcanic activity and climatic change can be answered within the next few decades. Microparticle and stable oxygen isotope studies on deep ice cores through the East Antarctic ice sheet will aid in establishing the relationship between variations in microparticle concentration and climate, as these cores may contain a climatic record of several glacial stages. The establishment of the cause of large scale climatic changes is one of the most important tasks facing modern man.

#### Suggestions for Future Work

One of the key problems is to determine the relationships between the particle concentration and size distribution in the air above the snow surface and in the snow deposited on the surface. When these relationships are known, the particle concentrations and size distributions measured in the deep ice cores can better be used to predict the particle load of the paleoatmosphere.

The potential uses of microparticle data are both numerous and varied. The data can be used to establish an independent chronology for deep ice cores as illustrated in previous chapters. Microparticles have the distinct advantage of not being subject to the degree of diffusion at depth in deep ice cores. The data can be used to reconstruct the depositional history of volcanic cosmic and continental (eolian) particles on snow and ice fields. The microparticle composition and morphology has the potential of being used not only to differentiate particles deposited on the ice sheet but also to separate these particles from those entrained into the ice sheet through bottom processes. It is probable that microparticle data from glacial ice over millennial time intervals can be used to determine background levels by which the variations in natural and anthropogenic particle deposition over recent years can be evaluated.

In order to make full use of microparticle data, one continuous core should be drilled from each of the major ice sheets and ice caps of the world. These cores should come from locations which provide the greatest potential for recovering a complete, long-term stratigraphic record. The cores should be analyzed in detail. For only with a detailed record can these potentials be fully realized and goals fulfilled. The core records would be used as a reference for all future investigations.

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## CONCLUSIONS

In this investigation it has been shown that in both the Camp Century, Greenland, and the Byrd Station, Antarctica, deep ice cores the highest concentrations of particles occur where oxygen isotope ratios exhibit the greatest negative values. In those sections of core which have traditionally been interpreted as being of Wisconsin age ( $> 10,000$  years B.P.) the concentration of small particles ( $0.65$  to  $0.82 \mu\text{m}$  in diameter) was up to 100 times greater than the mean Holocene ( $< 10,000$  years B.P.) concentrations in the Camp Century core and more than four times greater than mean Holocene concentrations for the Byrd core. This large increase in particle concentration may be due to an increase in global atmospheric particle concentration or due to a reduction in annual snow accumulation or both. The elemental composition and morphology for particles of Wisconsin age in the Byrd core suggest that the majority of the particles are volcanic in origin while in the Camp Century core most of the particles are probably terrestrially derived. The large increase in volcanic particles in the Byrd core suggests the possibility of greater volcanism during the Wisconsin, at least, in Antarctica. On the other hand, the large increase in terrestrially derived particles in the Camp Century core suggests a more intense atmospheric circulation during the Wisconsin glacial stage.

One of the important results of this work has been the observation of annual microparticle variations which provide a means of dating ice cores. The ages of the bottom ice calculated from microparticle concentration variations, assuming an annual cycle, are much less for both cores than those ages determined from a combination of the oxygen isotope and flow theory calculations. Based on preliminary microparticles studies the age of the bottom ice at Byrd Station is estimated to be 30,000 years old compared with the greater than 84,000 year logarithmic age calculation by Johnsen and others (1972). The age calculation for the bottom ice for the Camp Century core based on the microparticle variation interpretations is 12,000 years old, compared with the greater than 120,000 year logarithmic age calculated by Johnsen and others (1972). It is significant, however, that in the Byrd core the microparticle and logarithmic ages are in essential agreement for the last 20,000 years and in the Camp Century core for the last 6,200 years. The disparity in ages occurs in the lower 300 meters of both cores with the greatest disparity in the lowest 100 meters of each core. In fact, 100,000 years of the total logarithmic age estimate for the Camp Century core is represented by the lowest 100 meters of that core.

In both deep cores there is the possibility of a break in the stratigraphic record. There is evidence for this in the Camp Century core. The composition of particles in the lower 200 meters of the Camp Century core are atypical as they contain substantial amounts of Sn and Mo. Particles of this same combination of elements are not found in the

overlying ice in the Camp Century core nor in the entire Byrd core. In addition, at the same depth in the Camp Century core where the atypical particles first appear the particle concentration drops by a factor of 2 with respect to the concentration measured 20 meters above this section while the mean oxygen isotope ratios change by 4 o/oo. Thus, it may be argued (if sources of contamination are ruled out) that the source for these particles is the bedrock over which the Greenland Ice Sheet moves. If the particles below 1200 meters depth are from the bottom then the microparticle and oxygen isotope stratigraphies below this depth can not be meaningfully interpreted.

The microparticle data from these deep ice cores offer renewed support for volcanic activity as a viable climatic determinant during the Wisconsin glaciation. However, volcanic activity may play a significant causative role in the climate of a glacial period, only when the prerequisite of appropriate land and ocean distribution exists. The large thermal inertia of the oceans, as well as orogenic and epeirogenic movements, probably play equally important roles in producing glacial climates.

## REFERENCES

- Bader, H., W.L. Hamilton and P.L. Brown. 1965. Measurements of natural particulate fallout onto high polar ice sheets: Part I: Laboratory techniques and first results. U.S. Army Material Command, Cold Regions Research and Engineering Laboratory, Research Report No. 139, 86.
- Brookfield, M. 1971. Periodicity of orogeny. *Earth and Planetary Science Letters*, 12, 419-424.
- Bryson, R.A. 1968. All other factors being constant. . . Weatherwise, 12, 56-61.
- Budyko, M.I. 1974. *Climate and Life*. (English ed. D.H. Miller). Academic Press, New York, 309 pp.
- Bull, C. 1971. Snow accumulation in Antarctica. In Quam, L.O., and H.D. Porter, ed. *Research in the Antarctic*, American Association for the Advancement of Science, Washington, D.C., 367-421.
- Cameron, R.L. 1971. Glaciological studies at Byrd Station, Antarctica; 1963-1965. In Crary, A. ed., *Antarctic Snow and Ice Studies II*, Antarctic Research Series, 16, American Geophysical Union, Washington, D.C., 317-332.
- Castaing, R. 1963. X-ray microprobe techniques. In Pattee, H.H., Cosslett, V.E., and A. Engstrom, ed. *X-ray Optics and X-ray Microanalysis*. Academic Press, N.Y., 263-277.
- Chappel, J. 1973. Astronomical theory of climatic change: status and problem. *Journal of Quaternary Research*, 3, 221-236.
- Charlson, R.J. and M.J. Pilat. 1969. Climate: The influence of aerosols. *Journal of Applied Meteorology*, 8, 1001-1002.
- Colby, J.W. 1969. Quantitative microprobe analysis of thin insulating films. *Advances in X-ray Analysis*, 11, 287-305.
- Colby, J.W. 1971. Magic IV - A Computer program for quantitative microprobe analysis. Bell Laboratories, Inc., publication, 20 pp.
- Coulter Counter TA II Manual. 1975. Coulter Electronics, Inc., Healeah, Florida.
- Craig, H. 1961. Standard for reporting concentrations of Deuterium and oxygen-18 in natural waters. *Science*, 133, 1833-1834.

- Crozaz, G. and C.C. Langway, Jr. 1966. Dating Greenland firn-ice cores with Pb<sup>210</sup>. *Earth and Planetary Science Letters*, 1, 194.
- Damon, P.E. 1971. The relationship between late Cenozoic volcanism and tectonism and orogenic-epeirogenic periodicity. In Turekian, K.K., ed. *The Late Cenozoic Glacial Ages*, 15-35.
- Dansgaard, W. 1964. Stable isotopes in precipitation. *Tellus XVI* (X); 3, 436-468.
- Dansgaard, W., S.J. Johnsen, H.G. Clausen and C.C. Langway, Jr. 1971. Climatic record revealed by Camp Century ice core. Turekian K.K., ed. *Late Cenozoic Glacial Ages*. Yale University Press, New Haven, Conn., 37-56.
- Daugherty, H.E. 1968. Quaternary climatology of North America with emphasis on the State of Illinois, In Bergstrom, R.E., ed. *Quaternary of Illinois*, 61-69.
- Denton, G.H. and W. Karlen. 1973. Holocene climatic variations-- their pattern and possible cause. *Journal of Quaternary Research*. 3(2), 155-205.
- Dyer, A.J. and B.B. Hicks. 1968. Global spread of volcanic dust from the Bali eruption of 1963. *Quarterly Journal of the Royal Meteorological Society*. 94, 545-554.
- Epstein, S., R.P. Sharp and A.J. Gow. 1970. Antarctic Ice Sheet: stable isotope analyses of Byrd Station cores and interhemispheric climatic implications. *Science*, 168, 1570-1572.
- Faegri, K. and J. Iverson. 1964. Textbook of pollen analysis. Hafner Publishing Co., N.Y., 237 pp.
- Flint, R.F. 1971. *Glacial and Quaternary Geology*. John Wiley and Sons, Inc., N.Y., 251-266.
- Frenzel, B. 1973. Climatic Fluctuations of the Ice Ages. (Trans. from 1967 German, Nairn, A.E.M.). Case Western Reserve University, Cleveland, 230-333.
- Frye, C.J. and H.B. Willman. 1973. Wisconsin climatic history interpreted from Lake Michigan Lobe deposits and soils. Geological Society of America memoir 136. In Black, R.F., Goldthwait, R.P., and Willman, H.B., ed. *The Wisconsin Stage*, 135-152.
- Fuchs, V.E. and T. Paterson. 1947. The relation of volcanicity and orogeny to climatic change. *Geological Magazine*, 84, 325-333.

- Goldthwait, R.P. 1968. Two loesses in central southwest Ohio. In Bergstrom, R.E., ed. Quaternary of Illinois, 41-47.
- Gow, A.J., H.T. Ueda and D.E. Garfield. 1968. Antarctic ice sheet. Preliminary results of first core hole to bedrock. Science, 161 (3845), 1011-1013.
- Gow, A.J. and T. Williamson. 1971. Volcanic ash in the Antarctic ice sheet and its possible climatic implications. Earth and Planetary Science Letters, 13, 210-218.
- Gudmandsen, P. 1973. Radioglaciology soundings at proposed drill sites. Laboratory of Electromagnetic Theory, The Technical University of Denmark Lyngby, D185, 20.
- Hamilton, W.L. 1967. Measurement of natural particulate fallout onto high polar ice sheets; Part II: Antarctic and Greenland cores: U.S. Army Material Command, Cold Regions Research and Engineering Laboratory Research Report 139, 39 pp.
- Hamilton, W.L. 1969. Microparticle deposition on polar ice sheets. Ohio State University Research Foundation, Institute of Polar Studies, Report No. 29. 77 pp.
- Hamilton, W.L. and C.C. Langway, Jr. 1967. A correlation of microparticle concentrations with oxygen isotope ratios in 700-year old Greenland ice. Earth and Planetary Science Letters, 3, 363-366.
- Hays, J.D., H. Cook, G. Jenkins, W. Orr, R. Goll, F. Cook, D. Milow, and J. Fuller. 1972. In Initial Reports of the Deep Sea Drilling Project, Government Printing Office, Washington, D.C., 909-931.
- Hodge, P.W., F.W. Wright and C.C. Langway, Jr. 1964. Studies of particles for extraterrestrial origin, 3; Analyses of dust particles from polar ice deposits. Journal of Geophysical Research, 69, 2919-2931.
- Hughes, T. 1970. Convection in the Antarctic ice sheet leading to a surge of the ice sheet and possibly to a new ice age. Science, 170(3958), 630-633.
- Hughes, T. 1972. Thermal convection in polar ice sheets related to the various empirical flow laws of ice. Geophysical Journal of the Royal Astronomical Society, 27, 215-229.
- Hughes, T. 1973. Is the West Antarctic ice sheet disintegrating? Journal of Geophysical Research, 78(33), 7884-7910.
- Humphreys, W.J. 1929. Physics of the Air, McGraw-Hill, N.Y., 552-615.

- Johansen, S.J., W. Dansgaard, H.B. Clausen and C.C. Langway, Jr. 1972. Oxygen isotope profiles through the Antarctic and Greenland ice sheets. *Nature* 235(5339) 429-434; Corrigendum, 236(5344), 249.
- Junge, C.E. 1963. Air Chemistry and Radioactivity. Academic Press, N.Y., 111-201.
- Kennett, J. and P. Huddleston. 1972. Late Pleistocene paleoclimatology, foraminiferal biostratigraphy and tephrochronology, Western Gulf of Mexico. *Journal of Quaternary Research*, 2, 38-69.
- Kennett, J. and R. Thunell. 1975. Global increase in Quaternary explosive volcanism. *Science*, 187(4176), 497-503.
- Lamb, H.H. 1970. Volcanic dust in the atmosphere with a chronology and assessment of its meteorological significance. *Transactions of the Royal Society*, 266, 425-533.
- Lamb, H.H. 1972. Climate: Present, Past and Future. Volume 1: Fundamentals and Climate Now. Methuen and Co., LTD., London, 613 pp.
- Langway, C.C., Jr. 1963. Sampling for extraterrestrial dust on the Greenland Ice Sheet. *Assemblee Generale de Berkeley, Assoc. Internat. D'Hydrol. Scientifique*, 186-198.
- Langway, C.C., Jr. 1970. Stratigraphic analysis of a deep ice core from Greenland. *The Geological Society of America, Special paper* 125, 186 pp.
- Langway, C.C., Jr. and U.B. Marvin. 1964. Some characteristics of black spherules. *New York Acad. Sci., Conf. on Cosmic Dust*, 119, 205-223.
- Maarleveld, G.C. and R.P.H.P. van der Schans. 1961. De Dekzand-morfologie van de Gelderse Vallei. *Tijdsch. Koninkl. Nederl. Aardrijkskund. Genootsch.*, 78, 22-34.
- Marshall, E.W. 1959. Stratigraphic use of particulates in polar ice caps. *Geological Society of America Bulletin*, 70, 1643.
- Marshall, E.W. 1962. The stratigraphic distribution of particulate matter in the firn at Byrd Station, Antarctica. *In Antarctic Research, Amer. Geophys. Union, Geophys. Monogr.*, 7, 185-196.
- Mason, B.J. 1962. Clouds, Rain and Rainmaking. At the University Press, Cambridge, 145 pp.
- Mattern, C.R., F.S. Brackett and B.J. Olson. 1957. Determination of number and size of particles by electrical gating: blood cells. *Journal of Applied Meteorology*, 10, 56-70.

- McCormick, R.A. and J.H. Ludwig. 1967. Climate modification by atmospheric aerosols. *Science*, 156, 1358-1359.
- McIntyre, A. 1975. Thermal and oceanic structures of the Atlantic through a Glacial-Interglacial cycle. WMO/IAMAP Symposium on Long-Term Climatic Fluctuations, 421, 75-80.
- Menard, H.W. 1971. The late Cenozoic history of the Pacific and Indian Ocean Basins. In Turekian, K.K., ed. Late Cenozoic Glacial Ages, 1-14.
- Mitchell, J.M., Jr. 1961. Recent secular changes of global temperature. *Annals of the New York Academy of Science*, 95, 235-250.
- Mitchell, J.M., Jr. 1971. The effect of atmospheric aerosols on climate with special reference to temperatures near the Earth's surface. *Journal of Applied Meteorology*, 10, 703-714.
- Mitchell, M.J., Jr. 1975. What can we say about future trends in our climate? In Kopec, R.J., ed. Atmospheric Quality and Climatic Change, 160-167.
- Mock, S.J. 1968. Snow accumulation studies on the Thule Peninsula, Greenland. *Journal of Glaciology*, 7(49), 59-76.
- Mörner, N.A. 1972. Timescale and ice accumulation during the last 125,000 years as indicated by the Greenland  $O^{18}$  curve. *Geological Magazine*, 109(1), 17-24.
- Moseley, A.C.J. 1913. The high frequency spectra of the elements. *Philosophical Magazine*, 26, 1024-1034.
- Neumann, J. and A. Cohen. 1972. Climatic effects of aerosol layers in relation to solar radiation. *Journal of Applied Meteorology*, 11, 651-657.
- Newell, R.E. 1970. Stratospheric temperature change from Mount Agung volcanic eruption of 1963. *Journal of Atmospheric Science*, 27, 977-978.
- Nordenskiöld, A.E. 1886. Grönland. Seine Eiswüsten in Innern und seine Ostküste. Schilderung der Zweiten Dickson'schen Expedition ausgeführt im Jahre 1883. Leipzig, F.A. Brockhaus, 505 pp.
- Nye, J.F. 1959. The motion of ice sheets and glaciers. *Journal of Glaciology* 13, 493-507.
- Park, C.F., Jr. and R.A. MacDiarmid. 1970. Ore Deposits. W.H. Freeman and Company, San Francisco, 174-183.

- Pollack, J.B., B. Toon, C. Sagan, A. Summers, B. Baldwin and W.V. Camp. 1976. Volcanic explosions and climatic change: A theoretical assessment. *Journal of Geophysical Research*, 81(6), 1071-1083.
- Putnins, P. 1970. The climate of Greenland, In Orvig, S., ed. *World Survey of Climatology*, 3-128.
- Rasool, S.I. and S.H. Schneider. 1971. Atmospheric carbon dioxide and aerosols: effects of a large increase on global climate. *Science*, 173, 138-141.
- Reed, S.J.B. 1973. Principles of X-ray generation and quantitative analysis with the electron probe. In Andersen, C.A., ed. *Microprobe Analysis*, John Wiley and Sons, Inc., N.Y., 53-81.
- Robin, G., S. Evans and J. Bailey. 1969. Interpretation of radio echo sounding in polar ice sheets. *Philosophical Transactions of the Royal Society of London, Mathematical and Physical Sciences*. 265 (1166), 437-505.
- Rubin, M.J. and W.S. Weyant. 1965. Antarctic Meteorology. In Hatherton, T., ed. *Antarctica*. Praeger, N.Y., 375-401.
- Sarasin, P. and E. Sarasin. 1901. *Verhandlungen der Naturforschenden Gesellschaft in Basel*, 13, 603 pp.
- Schwerdtfeger, W. 1970. The climate of the Antarctic. In Orvig, S., ed. *World Survey of Climatology*, 253-355.
- Shaw, G.E. 1975. The Polar aerosol-climatic implications. In Warburton, J.A., ed. *Polar Meteorology Report of Workshop in Reno, Nevada*, 1975, 94-98.
- Sood, S.K. and M.R. Jackson. 1970. Scavenging by snow and ice crystals. In Engelmann, R.J. and W.G.N. Slinn, ed. *Precipitation Scavenging*, U.S. Atomic Energy Commission, Washington, D.C., 121-136.
- Stille, H., 1924. *Grundfragen der Verleichenden Tektonik*. (Borntraeger, Berlin), 443 pp.
- Suess, H.E. 1970. The three causes of secular C<sup>14</sup> fluctuations, their amplitudes and time constants. In Olsson, I.V., ed. *Radiocarbon Variations and Absolute Chronology*. Wiley Interscience, N.Y., 595-604.
- Taylor, L.D. and J. Gliozzi. 1964. Distribution of particulate matter in a firn core from Eights Station, Antarctica. In *Antarctic Research Series*, American Geophysical Union, Washington, D.C., 2, 267-277.

- Thomas, R.H. 1976. Thickening of the Ross Ice Shelf and the equilibrium state of the West Antarctica Ice Sheet. *Nature*, 259(5540), 180-182.
- Thompson, L.G. 1973. Analysis of the concentration of microparticles in an ice core from Byrd Station, Antarctica. Ohio State University Research Foundation, Institute of Polar Studies Report 46, 77 pp.
- Thompson, L.G. 1975a. Microparticle Research at the Institute of Polar Studies. *Antarctic Journal of the U.S.* 10(6), 312-313.
- Thompson, L.G. 1975b. Variations in microparticle concentration, size distribution, and elemental composition found in the Camp Century, Greenland and the Byrd Station, Antarctica, deep ice cores. Presented IUGG 16th General Assembly, International Symposium on Isotopes and Impurities in Snow and Ice, Grenoble, France, 17 p.
- Thompson, L.G., W.L. Hamilton and C. Bull. 1975. Climatological implications of microparticle concentrations in the ice core from Byrd Station, Western Antarctica. *Journal of Glaciology*, 14(72), 433-444.
- Thompson, L.G. and W. Dansgaard. 1975. Oxygen isotope and microparticle studies of snow samples from the Quelccaya Ice Cap, Peru. *Antarctic Journal of the U.S.* X(1), 24-26.
- Toon, O.B. and J.B. Pollack. 1976. A global average model of atmospheric aerosols for radiative transfer calculations. *Journal of Applied Meteorology*, 15(3), 225-246.
- Ueda, H.T. and D.E. Garfield. 1968. Deep-Core drilling program at Byrd Station. *Antarctic Journal of the U.S.* III(4), 111-112.
- U.S. Air Force. 1960. Handbook of Geophysics. The Macmillan Company, N.Y., 1-12.
- Ward, J.H., Jr. 1963. Hierarchical grouping to optimize an objective function. *Journal of American Statistical Assoc.*, 58(301), 236-243.
- Ward, J.H., Jr. and M.E. Hook. 1963. Application of an hierarchical grouping procedure to a problem of grouping profiles. *Education and Psychological Measurement*, 23(1), 69-80.
- Wexler, H. 1953. Radiation balance of the earth as a factor in climatic change. In Shapley, H., ed., *Climatic Change*. Harvard University Press, 73-107.
- Whillans, I.M. 1973. State of Equilibrium of the West Antarctic Inland Ice Sheet. *Science*, 182(4111), 476-479.

- Whillans, I.M. 1975. The surface mass-balance of Marie Byrd Land, Antarctica data interpretation. Institute of Polar Studies Report No. 51, 86 pp.
- Windon, H.L. 1969. Atmospheric dust records in permanent snowfields. Implications to marine sedimentation. Geological Society of America Bulletin, 80, 761-782.
- Woldstedt, P. 1954. Die Klimaturve der Tertiars und Quartars in Mitteleuropa, Eiszeitalter und Gegenwart, 415, 5-9.
- Wright, C.S. and R.F. Priestley. 1922. Glaciology, British (Terra Nova) Antarctic Expedition, 1910-1913. Harrison and Sons, London, Chapter 7, 581.
- Wright, F.W., P.W. Hodge and C.C. Langway, Jr. 1963. Studies of particles for extraterrestrial origin, 1. Chemical analyses of 118 particles. Journal of Geophysical Research, 68(19), 5575-5587.