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Ice Cores and SeaRISE--What We Do (and Don't) Know

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ABSTRACT. Ice-core analyses are needed in SeaRISE to learn what the West Antarctic ice sheet and other marine ice sheets were like in the past, what climate changes led to their present states, and how they behave. The major results of interest to SeaRISE from previous ice-core analyses in West Antarctica are that the end of the last ice age caused temperature and accumulation-rate increases in inland regions, leading to ice-sheet thickening followed by thinning to the present.

INTRODUCTION.

The goal of SeaRISE is to make predictions of marine-ice-sheet changes and any contribution they may make to changes in global sea level, based on scenarios of the future produced by other global-change research. For example, a scenario might be "If humans burn all of the known oil and coal reserves on the planet over the next 100 years and raise the global temperature 5°C", and the prediction might be "then the West Antarctic ice sheet (WAIS) will collapse, raising global sea level 6 m in 200 a". Different scenarios might lead to different predictions.

The predictions must be made with models--simplified representations of the real marine-ice-sheet system. All models, whether numerical, analytic, or analog, share certain features. These features include a description or representation of how the system being modeled behaves (governing equations in a numerical model), what the system is like at some time (initial conditions), and what is done to it by things outside (boundary conditions).

In addition, for a model to be useful, it must be tested or validated. For earth processes that occur over decades or centuries, the only practical way to test a model is against data from the past: if the model can start from some former conditions and, using past forcings, predict the present state (or if the model can "retrodict" past states), then we can place some confidence in its ability to predict the future.

In building the SeaRISE model(s), ice cores are critical for learning ice-sheet history for validation, and play an important role in learning the governing equations and even the initial conditions. Isotopic, trace-element and total-gas analyses of ice cores contribute primarily to the history, and physical-property measurements provide information on the governing equations. In addition, the borehole left by core recovery is valuable for extracting history and initial conditions from ice temperatures, learning governing equations from borehole deformation and logging, and gaining access to the bed for further studies.

Here, I briefly review what is known, and not known, about marine ice sheets (primarily the WAIS) based on ice-core measurements. I provide some literature citations to guide the interested reader, but make no attempt to present a fully referenced review. In order, I discuss ice-core contributions to learning ice-sheet history (global record, surface temperature, accumulation rate, surface elevation, and other information), current conditions, and governing equations, and I conclude with an editorial on what further studies are needed for SeaRISE.

ICE-SHEET HISTORY.

THE GLOBAL RECORD.

An ice core records both local and global events. Of these, local events are more valuable to SeaRISE, but the global signal may be of greater interest to the broader scientific community.

Probably the most notable result from ice-core research has been the demonstration that the ice-age atmosphere contained less CO_2 and CH_4, and more dust, than the modern atmosphere (Thompson and others, 1981; Lorius and others, 1989; Chappellaz and others, 1990). Reduction of these greenhouse gases, and increase of dust, would have tended to cool the planet, contributing to the low ice-age temperatures (Harvey, 1988). The changes may have been startlingly rapid, although ice flow has caused at least some of the changes to appear more rapid than they actually were (Staffelbach and others, 1988).

These and similar results place significant constraints on global climate models, and provide important clues to the past operation of the global climate system. To the extent that marine ice sheets are embedded in the world climate, these results are useful in determining ice-sheet history. However, they are unlikely to be of direct, primary importance to SeaRISE.

TEMPERATURE.

Of greater interest to SeaRISE are ice-core analyses for more local effects: near-surface temperature, net snow accumulation, and ice-surface elevation. These require measurement of isotopes, total gas content, physical properties, and annually varying properties in ice cores, combined with borehole measurements, atmospheric and oceanic modeling, and ice-flow modeling.

For temperatures, the most useful information for SeaRISE is the history of the near-surface snow temperature (the 10-m temperature), not the air temperature. The rate of ice deformation is known to increase exponentially with the ice temperature, which changes over climatic cycles. The rate at which changes in 10-m temperature propagate into the ice and affect its flow is readily calculated if the flow pattern is known, so the 10-m temperature provides a useful boundary condition for ice-flow modeling. The 10-m temperature may differ from the mean annual near-surface air temperature by a few tenths of a degree to a few degrees, however, depending on radiative conditions, wind pumping, rates of sublimation, and other factors that are difficult to model or measure, and that may have changed over time (Paterson, 1981, ch. 10). (Much is known about these processes, but we cannot calculate the difference between air and snow temperatures from first principles, nor can we estimate how it has changed with time.)

The only direct sensor of past 10-m temperatures is the physical temperature of the ice today, and of underlying bedrock if the ice is frozen to its bed. Any change in snow temperature moves downward into the ice sheet by conduction and ice flow. If the ice flow, heat sources, and thermal properties of ice are known, then the steady-state temperature profile in ice can be calculated. Deviations of measured temperatures from the steady-state profile represent climatic changes, with the age of the changes increasing with the depth of the deviations in the ice.

Because of thermal diffusion, the deviation of measured temperature from steady state caused by a surface-temperature change decreases in amplitude but spreads over a greater depth range with increasing age, and eventually becomes too small to measure or overlaps perturbations of similar age and ceases to be recognizable. Thus, whereas seasonal cycles of the last year or two are evident in temperature records from the upper few meters, events thousands of years old had to persist for hundreds of years to be recognizable in temperature records today.

Measurement of ice temperatures to 0.01°C or better is possible once a hole exists and has equilibrated with the surrounding ice, and such accuracy is needed for paleoclimatic interpretation. Such accuracies are difficult to achieve from thermistor strings frozen into holes melted with hot water, owing to thermistor drift and freeze-in pressure. Hot-water drilling combined with antifreeze and casing could be developed to allow temperature measurements without core drilling (G. Clow, private communication, 1990), but the only proven method today for accurate, deep measurements is to utilize a hole left by ice-core drilling.

Most earlier workers have tested whether the ice-temperature record is consistent with some assumed history of near-surface snow temperatures, by using that assumed history to drive a forward model of ice temperature and asking whether this explained most of the variance between observed and steady-state model profiles (within unspecified but rather broad error limits; e.g. Johnsen, 1977). The assumed history may be derived from isotope records or a variety of regional proxy records and may be adjusted in a more-or-less *ad hoc* method to optimize the fit between model and observed profiles (e.g. Fig. A; Alley and Koci, 1990). A better approach is to use inverse theory, a technique that allows determination of that surface-temperature history (or histories?) most consistent with the observed temperature profile, and that is independent of other proxy records (MacAyeal and others, in review).

The major results of borehole temperature analyses to date have been to demonstrate the basic correctness of isotopic or other proxy records (e.g. Dahl-Jensen and Johnsen, 1986), and to provide information on ice-flow patterns (Bolzan, 1985). It certainly should be possible to obtain better results from new temperature records and from re-analysis of previously measured temperatures, using the more-sophisticated analysis tools now under development. The major drawbacks in use of modern ice temperatures to infer past 10-m temperatures are: 1) ice dynamics and past accumulation rates must be known to calculate temperature profiles for comparison with measured data; 2) the method lacks high resolution for old events; and 3) analyses to date have not been optimized.

An independent method to infer past temperatures, and the method that has been used most widely, is to calculate the temperatures from measurements of oxygen- or hydrogen-isotope ratios in ice cores.

The isotopic composition of snowfall depends on: the composition of the source waters; the degree of equilibrium and nonequilibrium isotopic fractionation during evaporation from the source; the amount of water-vapor removal and the temperature history of that removal prior to formation of the precipitation of interest; and the condensation temperature of the precipitation of interest. If the source and atmospheric-path terms are known, then the isotopic composition gives the condensation temperature for the snowfall. Assuming further that the condensation temperature changes in the same way over time as the near-surface air temperature, the isotopic composition is a proxy for the air temperature (Robin and Johnsen, 1983). Finally, if snow drifting is limited, if sublimation following deposition does not change the isotopic composition (except perhaps by diffusional smoothing of annual signals), if the seasonality of precipitation does not change over time, and if the source of the ice in a core can be determined from ice-flow calculations (Fig. B), then the isotopic composition of the ice in a core provides the near-surface air temperature when and where that ice was deposited, which can be taken as a proxy for the 10-m temperature (with the further uncertainties about offset between snow and air temperatures, as discussed above).

The major advantages of this method of paleothermometry over ice-temperature measurements are that the isotopes have much higher resolution (isotopic diffusion does smooth the record with time, but this smoothing is some orders of magnitude slower than the smoothing of the physical temperature record), and that the isotope records from central regions of ice sheets are nearly independent of ice flow, unlike physical temperature records. The major disadvantage is the indirect nature of the measurement: source compositions, evaporation fractionations, condensation histories, offsets between condensation and near-surface air temperatures, diagenetic changes, and offsets between near-surface air temperatures and 10-m temperatures all must be known to provide the glaciologically significant boundary condition.

The situation is not quite so bleak as this may sound. Temperature histories inferred from isotopes are broadly consistent with those from ice temperatures. Isotopic analyses of marine fossils constrain changes of the isotopic composition of oceanic source regions (although with some uncertainty introduced by oceanic temperature changes; Shackelton, 1983). Combined analysis of hydrogen and oxygen isotopes in ice cores provides information on isotopic fractionation during evaporation from the oceans (Jouzel and others, 1982). Atmospheric modeling, including modeling with general circulation models (GCMs), helps estimate isotopic changes prior to the snowfall of interest (Jouzel and others, 1987).

Modern studies show a good correlation between mean annual isotopic composition of snowfall and mean annual near-surface air temperature at a site (Lorius, 1983). This correlation generally is used to convert isotopic compositions to paleotemperatures, thus avoiding the need to calculate path effects from first principles. However, one study comparing isotopic compositions to instrumental records and site temperatures in the Antarctic Peninsula found that the correlation of site temperature with isotopic composition at the present time has a significantly different slope than the correlation between temperature and isotopic composition over time at a site (Peel and others, 1988). Also, if a given isotopic composition is found in Marie Byrd Land in West Antarctica, it indicates a temperature about 10°C warmer than if found in most of East Antarctica (Fig. C, D; Lorius, 1983). The reasons for this difference are not understood fully, so we cannot tell whether it has been maintained over time. These difficulties may not affect interpretation of East Antarctic records, but are a concern in West Antarctica.

If we wish to obtain paleotemperatures from isotopes with accuracy at the 1 or 2°C level, then there is a clear need for improved coupled ocean-atmosphere-sea ice modeling with explicit treatment of oxygen and hydrogen isotopes to learn how climatic factors have affected the isotope-air temperature relation. Further surveys of near-surface isotopic compositions in West Antarctica would be useful. Detailed studies of the processes causing isotopic fractionation are needed. Also, the relation between near-surface air temperature and 10-m snow temperature must be investigated systematically.

Despite the numerous weaknesses, isotopes are generally accepted as the primary paleothermometer for ice sheets, and have provided much of our present knowledge of past temperatures. They indicate that, during the height of the last ice age, surface temperatures in central regions of East and West Antarctica were 5 to 10°C colder than today (Fig. E; Jouzel and others, 1989), with perhaps 10 to 15°C cooling in Greenland (Dansgaard and Oeschger, 1989), and that warming to near modern values occurred over roughly 5000 years in the range of 15,000 to 10,000 years ago. Since this deglacial warming, temperature oscillations have been in the 1 to 2°C range if averaged over more than a few years. Very good isotopic time series are available from a few places (e.g. Mosley-Thompson and others, 1990; Grootes and others, 1990), although the regional coherence and even the climatologic or meteorologic significance of the observed post-glacial oscillations are questionable (Benoist and others, 1982; Grootes and others, 1990).

ACCUMULATION RATE.

The second major forcing on ice sheets is net accumulation (or ablation). For the grounded part of the West Antarctic ice sheet, and for most of the East Antarctic and Greenland ice sheets, this is dominated by the surface snowfall. Analysis of ice cores is the only reasonably direct way we have to learn past accumulation, although climate modeling can provide important information and should be pursued.

Net snowfall can be determined from ice cores by measuring annual layer thicknesses and correcting for ice flow, by using flow models to estimate accumulation rates needed to produce observed ice-temperature profiles or surface-elevation histories, or by measuring dilution of some trace constituent with known deposition or production rate. Of these, the methods combining flow modeling with ice-core analyses have been the most practical, but improved resolution from dilution techniques would allow us to learn accumulation independent of ice flow and to use the other techniques to determine ice-flow patterns better.

Many parameters that can be measured in ice cores vary with an annual signal, including various trace chemicals, physical properties, dust, and isotopes (Hammer, 1989). Layer thinning eventually renders all of these methods unusable in ice, and diffusion affects some of them and limits their utility. For isotopes and hydrogen peroxide, an accumulation rate of about 20 cm/a ice or more is needed to allow annual signals to survive diffusion in snow and firn and reach the ice, where diffusion is slower and the signal can be recognized for hundreds or thousands of years (Johnsen and Robin, 1983).

Annual layers have been counted continuously from the surface for several thousand years in Greenland (Dansgaard and Oeschger, 1989), for about 2000 years in East Antarctica (Mosley-Thompson and Thompson, 1982 for 911 years; A.J. Gow, pers. comm., 1990 for 2000 years), and for several hundred years in West Antarctica (Gow, 1968), and older and deeper annual layers have been recognized, although the tedious job of measuring all of the annual layers down to these deeper levels has not been done. Microparticles offer the greatest opportunity for recognizing annual layers in deep and old ice, because they are negligibly affected by diffusion.

The annual-layer thickness provides the accumulation rate only if corrected for density and ice flow. Density is measured easily, and ice flow is relatively well known in the upper half of an ice sheet, especially away from ice divides, so that past accumulation rates can be learned quite accurately for recent times if accumulation rates are high enough to produce annual layers. However, ice flow complicates absolute determination of accumulation rates from deeper ice.

Ice-flow modeling combined with ice-core data can contribute to learning past accumulation rates in other ways. The temperature-depth relation in ice depends on the accumulation rate as well as the surface temperature, and changes in accumulation rate can be estimated by jointly determining the accumulation-rate and surface-temperature histories needed in models to match observed temperature profiles (e.g. Dahl-Jensen and Johnsen, 1986). Similarly, accumulation rate helps control ice-sheet thickness and surface elevation. Past accumulation rates can be estimated by determining the history needed in ice-flow models to match changes in ice-sheet elevation estimated from total-gas-content measurements (see below).

The other direct method of determining accumulation rates from ice-core measurements is to measure dilutions. This technique is in its infancy, but could prove quite useful if developed successfully. The basic idea is that certain trace constituents are deposited on the ice sheet or formed near the surface of the ice sheet at a rate independent of time. If the annual accumulation of snow is large, then the concentration of the trace constituent is low, but if the annual accumulation of snow is small, then the trace constituent is more concentrated. One such constituent is ^{10}Be, which is a cosmogenic isotope with a short atmospheric residence time. With certain exceptions, it is believed that production and deposition rates have been nearly constant over time, allowing changes in snow accumulation to be inferred from concentration changes of ^{10}Be measured in ice cores (Raisbeck and others, 1987).

An indirect method to estimate past accumulation rates is used more often than any of these more-direct methods. At present, net snow accumulation on the East Antarctic ice sheet is inversely correlated with mean annual temperature, in the way that would be expected if accumulation were controlled by the saturation vapor pressure of the air. If it is assumed that saturation vapor pressure of precipitating air masses is correlated with near-surface air temperature and controls mean accumulation rate, then the accumulation rate can be calculated from the surface temperature, which is estimated from isotopes or ice temperatures (Robin, 1977).

Such an approach obviously has certain drawbacks. For example, in the modern world snowfall increases with increasing temperature in central East Antarctica on a mean annual basis, but most snowfall in Antarctica occurs in the winter when it is cold (Bromwich, 1988). In West Antarctica even the mean-annual correlation does not work well in places--Byrd Station is about 1.5°C colder than the Siple Coast, but has almost twice the net accumulation of the Siple Coast (probably owing in part to the anomalously low elevation of the Siple Coast; Robin, 1983; Alley and Bentley, 1988; Whillans and Bindschadler, 1988). Indeed, the variability in accumulation rate for a given temperature in Antarctica approaches an order of magnitude, even if the blue-ice ablation zones are excluded (Robin, 1977). Clearly, physically based atmospheric modeling is needed to test or replace this simple correlation model.

Despite all of the problems of the different methods of inferring past accumulation rates, they do achieve some degree of internal consistency. The temperature-accumulation relation and ^{10}Be produce similar estimates of accumulation during the last ice age in central East Antarctica (Jouzel and others, 1989). In south-central Greenland, model attempts to match the measured physical temperature of the ice require a glacial-interglacial accumulation change similar to that estimated from ^{10}Be, coupled with a temperature change similar to that indicated by isotopes (Dansgaard and Oeschger, 1989). For West Antarctica, changes in surface elevation inferred from measurements of total gas content on the Byrd core indicate a deglacial thickening such as would be produced by a significant increase in accumulation rate (Whillans, 1981; Raynaud and Whillans, 1982), similar to that estimated from layer thicknesses measured by electrical conductivity (Hammer and others, 1985) and estimated from temperature change. In each case, accumulation apparently increased significantly at the end of the last ice age (by about a factor of two, with large error bars). Following this deglacial transition, accumulation seems to have shown relatively small oscillations around a Holocene mean (e.g. Clausen and others, 1988).

Because the most commonly used technique for estimating accumulation rates on glacial-interglacial time scales is to estimate temperature from isotopes, and to estimate accumulation from temperature, we are left with less confidence in our knowledge of accumulation rates than in our knowledge of temperatures. One alternative approach would be to develop good atmospheric models for the ice sheet. If we can determine surface elevations and surface temperatures from ice cores, and use these to constrain atmospheric models, then the models can (in principle) estimate snowfall rates. A second approach would be to accelerate continuing work on the dilution techniques.

A third approach would be to improve the coupling of ice-flow models and ice-core observations. The ice temperature, ice thickness, and form and thickness of layers in the ice depend on the way the ice flows, and on the past surface temperature and accumulation rate (among other factors). A fully coupled model, if used with appropriate inverse techniques, should be able to reveal those combinations of temperature history, accumulation history, and ice-flow parameters that best match the modern state of the ice sheet. The hope is that the result of such a model, constrained by independent data, would yield a unique ice-sheet history. Care must be taken to avoid circularity in such an exercise: we seek paleoclimatic data to test ice-sheet models, but if we use all the data in developing the models, they no longer are testable. Thus, some subset of the data should be held out from model development for testing.

All of these approaches for learning past accumulation rates more accurately merit study. Of them, development of improved models for polar precipitation, and development of coupled ice-flow/ice-core models, seem especially suited for an interdisciplinary program such as SeaRISE.

ICE-SURFACE ELEVATION.

The third major element of ice-sheet history that can be derived from ice-core analysis is surface elevation, and from surface elevation, ice-sheet thickness. Total gas content of the ice is the best measure here, although isotopes also are useful.

In the modern world, the volume of air bubbles per unit volume of ice trapped by the transformation of snow to ice without melting is observed to be a weak function of mean annual temperature and appears to be independent of accumulation rate. From the perfect gas law and this known relation between temperature and bubble volume, the number of air molecules trapped in unit volume of ice yields the air pressure at pore close-off. Because air pressure decreases with increasing elevation, the calculated air pressure can be used to infer past elevations of pore close-off. Further, ice-surface elevations are believed to undergo larger and more-rapid changes than any changes in the depth of pore close-off below the surface, so changes in total gas content are used to estimate changes in surface elevation (Raynaud, 1983).

The method is subject to some uncertainty, of course. The total gas content varies between closely spaced samples, probably indicating seasonal or other changes in the snow density (for example, close-off of a high-density layer may effectively seal a deeper layer with lower density in which pores are still open internally; Raynaud and Whillans, 1982). The measurements must be corrected for ice flow. The relation between air pressure and elevation can be changed meteorologically (e.g. by a change from a consistently low-pressure to a consistently high-pressure region). Elevations and atmospheric pressures are referenced to sea level, which moves up and down significantly on glacial-interglacial time scales. In addition, we lack a physical understanding of the process of pore close-off and the observed volume-temperature relation, so we lack a strong basis for assuming that the relation has not changed over time, especially when extrapolating the relation to ice-age conditions of temperature and accumulation rate that do not exist on earth today.

The data obtained by this method are, in general, consistent with inferences from ice-flow modeling, glacial geology, and other ice-core techniques. For example, data from Camp Century in Greenland show that old ice originated at a higher altitude than the modern surface at this near-coastal site (Budd and Young, 1983). Glacial-geological evidence from along the Greenlandic coast shows that the ice sheet was advanced during the last ice age, and ice-flow modeling shows that such an advance probably would cause inland ice deposited at a higher elevation to flow through Camp Century, which now is a local center of ice outflow. In East Antarctica, data from Vostok indicate that the ice sheet was thicker during interglacials than during glacials (Raynaud and others, 1987). This is expected from ice-flow modeling, if reconstructed surface-temperature and accumulation-rate histories are accurate (Whillans, 1981).

Preliminary results from Vostok seemed to show that the ice sheet thickened more rapidly during deglaciation than the reconstructed accumulation rate (Raynaud and others, 1987), and this has been used in informal discussions (e.g. at the first SeaRISE workshop) to argue that total-gas measurements are not reliable. However, the data and interpretations are not yet published, and were not finished at the time they were presented. Moreover, the reconstructed ice-age conditions of low temperature and accumulation rate at Vostok are not duplicated on earth today, requiring an extrapolation rather than interpolation of the modern temperature-volume relation, and introducing greater uncertainty. Finally, there are real questions about how the pressure field over Antarctica varied during the last ice age, questions that should be addressed by atmospheric (or polar environmental) modeling. Pending further data, it does appear that total-gas-content measurements provide useful information on past ice-surface elevations, but further research directed to solving ice-dynamical problems clearly is warranted.

An independent approach is based on analysis of isotopic data (Grootes and Stuiver, 1986; 1987). After correction for source and path effects and ice flow, an isotopic shift in an ice core can be interpreted as a temperature change at a site. That temperature change can be decomposed into two components: a climatic change and an elevation change. If the climatic change can be inferred in some manner (e.g. by measuring the isotopic shift at a location where models or other data show that elevation changes were small, and assuming regional coherence of the climatic change), then subtracting the climatic from the total temperature change yields the change linked to elevation. By assuming a lapse rate, this elevation-linked change yields the elevation change at a site.

Clearly, there are a number of assumptions here. Particularly troubling is the observation that the relation between modern surface temperatures and elevations in West Antarctica does not follow the expected atmospheric lapse rate: Byrd Station is about 1200 m higher (Drewry, 1983), but only 1.5°C colder, than the Upstream B camp on ice stream B. (Again, the low elevation of Upstream B and the Siple Coast in general may contribute to this discrepancy.) However, the results of the method agree fairly well with results from total-gas measurements (Grootes and Stuiver, 1986; 1987).

For the West Antarctic ice sheet, detailed, multi-parameter analyses of surface elevation are available from Byrd Station only. There, it appears that the onset of deglaciation caused thickening of one to a few hundred meters (Fig. F), followed by similar thinning to the present. This is consistent with ice-flow models based on the inferred changes in accumulation rate and temperature: warming and accumulation-rate increase at the end of the ice age would have caused immediate thickening from the extra snowfall, with delayed thinning as the warmth penetrated to deeper ice and increased flow rates (Whillans, 1981; Raynaud and Whillans, 1982; Jenssen, 1983; Budd and Young, 1983; Grootes and Stuiver, 1986; 1987).

OTHER DATA.

Ice cores can provide other useful historical data. For example, if basal freeze-on has been occurring, it will be evident through a thick layer of dirty basal ice (as at Byrd; Gow and others, 1979) or of frozen-on seawater (as at J9 on the Ross Ice Shelf; Zotikov and others, 1980). The physical properties of the ice record the stress state and cumulative strain of the ice (or at least the minimum value of the cumulative strain; Alley, 1988).

Many further examples could be given. However, most would be relatively peripheral to SeaRISE, and are omitted here. Notice, however, that this in no way implies that other areas are not important science, but only that they are not central to SeaRISE goals.

SUMMARY.

For central West Antarctica, then, ice-core data and other evidence centered around Byrd Station tell a broadly consistent story. Ice-age temperatures were colder (by several degrees), accumulation was less (by perhaps a factor of two), and surface elevation was similar to modern. The end of the ice age increased temperature and accumulation rather rapidly to modern values (between perhaps 15,000 and 10,000 years ago), which caused thickening of one to a few hundred meters, followed by thinning to the present. Exact timing of events is not known accurately, however, and uncertainties in magnitudes are large. Since the end of the deglaciation, temperatures and accumulation rates probably have fluctuated little about Holocene means.

INITIAL CONDITIONS.

Ice cores will provide, and have provided, some useful information on initial conditions for SeaRISE modeling of the WAIS. This information is mostly related to mass balance.

The modern mass balance (i.e., whether the ice sheet is growing or shrinking) can be calculated if we know the ice input from snowfall and ice output from ice flow and basal melting. The input can be measured in several ways. The most successful has been collection of shallow ice cores (<20 m) with hand augers, measurement of gross beta activity to locate known reference horizons from atmospheric atomic-bomb testing (e.g. A.D. 1955), and measurement of core density; the accumulation rate averaged over the period between the reference horizon and the coring then can be calculated directly (e.g. Whillans and Bindschadler, 1988). Technically, this is ice-core analysis, although the work usually is carried out by ice dynamicists.

The ice outflow depends on the surface velocity field and the velocity-depth profile. The surface velocity field is measured by surface surveying tied to satellite tracking (TRANSIT or GPS) or through repeat imaging of specific sites (satellite or airborne photogrammetry)(SCP, 1988). The velocity-depth profile can be calculated from flow models, but should be measured for improved accuracy. Measurement is easiest through repeat inclinometry in boreholes left by ice-core recovery (Fig. G). Data from such inclinometry from the Byrd Station borehole have been used by Whillans (1977) to estimate that the ice sheet upglacier of Byrd now is thinning by a few centimeters per year.

GOVERNING EQUATIONS.

An ice-flow model requires governing equations for ice deformation, basal sliding, and subglacial deformation, linking the ice velocity or strain rate to the driving stress and other factors. For basal sliding and subglacial deformation, ice-core drilling provides access holes to the bed. Ice-core access holes are more expensive and slower to drill than hot-water holes, but remain open much longer for extended experiments, and thus are complementary to hot-water holes. (Antifreeze techniques for hot-water holes are under development, however; H. Engelhardt, pers. comm., 1990.)

Ice-core drilling is especially useful for learning the governing equations of ice deformation. The rate of ice deformation in any direction is linearly proportional to the stress deviator in that direction and to the square of the effective stress, increases exponentially with the absolute temperature, and also varies with ice fabric and dislocation substructure (Paterson, 1981, ch. 3). Ice-core drilling provides access holes in which temperature can be measured, and also allows studies to measure the ice fabric.

An ice crystal is about two or more orders of magnitude "softer" parallel to its basal plane than perpendicular to it (that is, a given shear stress oriented parallel to the basal plane will produce a shear strain rate more than 100 times larger than that produced by a shear stress of the same magnitude oriented perpendicularly to the basal plane). Deformation of a polycrystalline ice sample causes the pattern of c-axes (normals to the basal planes) to change, which changes the hardness of the ice and the deformation rate (Budd and Jacka, 1989). In natural ice, thin-section analyses supplemented by ultrasonic measurements on core samples, borehole sonic logging, and surface seismic surveys, show that a shear stress causes c-axes to cluster so that the ice is soft to that shear stress, that a small normal stress causes the ice to harden to that stress, and that a large normal stress causes the ice to soften to that stress (Alley, 1988; Budd and Jacka, 1989). In addition, softening to a given stress causes hardening to some other stresses; anisotropy of c-axis fabrics causes the ice hardness to be a tensor property. In polycrystalline ice, the difference between hard and soft to a given stress can be as large as an order of magnitude or more, as demonstrated by laboratory deformation tests on ice-core samples and by analysis of data from repeated borehole inclinometry (Budd and Jacka, 1989). Ice-flow models show that the behavior of modern ice sheets cannot be explained adequately without including effects such as these (Van der Veen and Whillans, in press).

The major results directly applicable to SeaRISE thus far are that ice subjected primarily to basal shear in inland ice (and especially in ice-age ice inland) develops a strong vertical maximum of c-axes, which softens the ice to simple shear by roughly three times (e.g. Fig. G; Dahl-Jensen and Gundestrup, 1987). Ice subjected to strong normal stresses on ice shelves forms a soft c-axis girdle pattern around the normal stresses (Budd and Jacka, 1989); seismic results suggest that this extends into ice streams as well (Blankenship and others, 1989). We can infer that as ice flows from inland into ice streams or ice shelves, it must reorganize its fabric, passing through a hard configuration (Alley, in prep.); however, this has not been tested directly. The picture in inland ice is complicated slightly by the deepest ice of the Byrd core (Gow and Williamson, 1976), which would be soft to a vertical normal stress but not to a basal shear stress; this probably is related to the unusual topographic position of Byrd near the top of a subglacial mountain.

The temperature profiles from Byrd and elsewhere in coreholes, and from Upstream B, Crary Ice Rise, and other West Antarctic sites in hot-water holes (e.g. Engelhardt and others, 1990), are useful results for flow modeling. Also, the repeat inclinometry of the Byrd hole (Hansen and others, 1989) shows how the ice deforms, which helps constrain the flow law there.

An additional useful result from ice-core analyses is the identification of marker horizons (Budd and others, 1989). An event such as a large volcanic eruption, or a spike in

^{10}Be deposition, can be identified in several ice cores (Fig. E). The depth of the event depends on accumulation rates and ice flow. Any flow model must be able to reproduce the relative depths of an event in different cores. (Internal layers detected with radar serve a similar purpose; Whillans, 1976.)

SUMMARY--AND AN EDITORIAL.

Ice-core analysis is a widely useful discipline. It provides the only direct measurement of past atmospheric compositions, and thus is a centerpiece of global-change studies. It also provides important data on local atmospheric conditions, and on behavior of the ice sheet.

Because SeaRISE studies marine ice sheets that are embedded in the global environment, the global implications of ice-core analyses are of interest to us. However, to optimize return from investment, SeaRISE needs to emphasize certain aspects of ice-core analysis. The major contributions of ice coring to SeaRISE would come (and have come) from learning the history of the ice sheet (near-surface temperatures, net surface accumulation rate, and surface elevation), the present state of the ice sheet (temperature and mass balance) and the way it flows (ice hardness). Ice cores also provide important, long-lasting access holes to the bed for ice-dynamical and geological studies.

A representative (but not exhaustive) list of the major experiments needed on ice cores to achieve these contributions includes: isotopic analyses (for past temperatures, surface elevations and accumulation rates); detailed measurements of annually varying components such as microparticles, chemistry, isotopes and visible strata (for accumulation rates); studies of cosmogenic nuclides such as ^{10}Be (for accumulation rates); total-gas-content measurements (for surface elevations); and thin-section and ice-deformation studies (for ice hardness). Related experiments in the corehole include: thermometry (ice hardness, past temperature); repeat inclinometry (mass balance, ice hardness); and sonic logging (ice hardness); as well as subglacial studies for other purposes.

For local signals, core locations are critical to answering major questions. In West Antarctica, the past configuration of the ice sheet is of special interest: Did the ice sheet collapse during the previous interglacial? Did its margins thicken and its grounding line advance during the most recent glacial? Is it collapsing now? Discovery of West Antarctic ice from the previous interglacial would show that the ice sheet existed then. The best chance of finding such ice is in regions without basal melting, as indicated by ice-flow models. If such ice is absent, measurement of geothermal flux in subglacial bedrock combined with ice-flow modeling can be used to determine whether it melted off recently or whether it was not present because of ice-sheet collapse. Inland ice is only weakly sensitive to changes at the ice-sheet margins (by analogy to Alley and Whillans, 1984), so data from marginal regions are needed to learn what has happened at the margins during deglaciation. For example, total-gas and isotopic data from a core only a few hundred meters long on ridge BC on the Siple Coast would reveal whether the ice-age ice sheet there was significantly thicker than today (Graham, 1988).

In the U.S., ice-core analysis often is equated with ice-core geochemistry. In the case of the Greenland Ice Sheet Project 2 (GISP2) drilling, borehole measurements, ice-core geophysics and surface glaciology were discouraged purposefully, to concentrate on ice-core geochemistry. This was entirely appropriate, given the goals of the project, its limited funding, and the possibility of adding-on studies during or after coring. However, to optimize the value of future ice coring to SeaRISE, it is imperative that GISP2 not be used as a model. Rather, glaciological as well as geochemical aspects should figure in site selection, and the geochemical studies must be coupled to a full suite of "glaciological" measurements: total gas content for surface elevations, ice-core physical properties, borehole logging for sonic velocity, temperature, and inclination change, surface surveys of accumulation, temperature, and ice movement, surface geophysical surveys, etc. This would ensure that information on both global and local conditions would be obtained.

In addition, there are a number of opportunities for interdisciplinary research that would greatly improve the scientific return from ice-core analyses. The major openings seem to be in the fields of atmospheric modeling and ice-flow modeling. Models should be formulated to track features that can be measured in ice cores (e.g. isotopes for atmospheric modeling, ice temperatures and the depth of marker horizons for ice flow). Joint analysis of ice-core, atmospheric and ice-flow data then can improve the interpretation of the ice cores and of the dynamics of the entire system.

Few truly solid results have been obtained from ice-core studies that are directly relevant to SeaRISE. However, no major effort has been mounted to obtain such results for more than twenty years. The many techniques available to us provide considerable confidence that a well-developed ice-coring program would provide much information absolutely essential to the success of SeaRISE.

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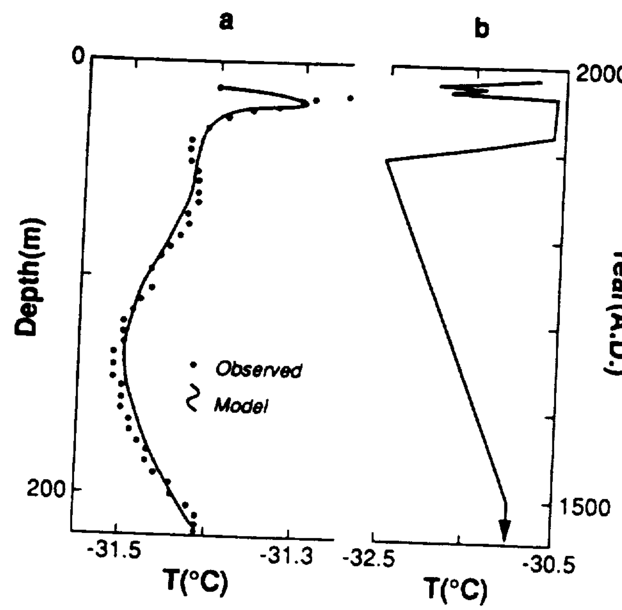

Fig. 1a. Temperature-depth profile, in °C. Data (·) measured with first thermistor (second thermistor was consistently 0.035°C colder), and model results (smooth curve). Seasonal effects from the two years prior to drilling are evident between 12 and 15 m: they are large (but not plotted) shallower than 12 m. 1b. Best-fit history of surface temperature.

Fig. A. Temperature data and interpretation, from GISP2, Greenland (Alley and Koci, 1990).

Figure 5.53. Oxygen isotopic ratios for Byrd Station (after Epstein *et al.*, 1970) compared with curve calculated using flowline model.

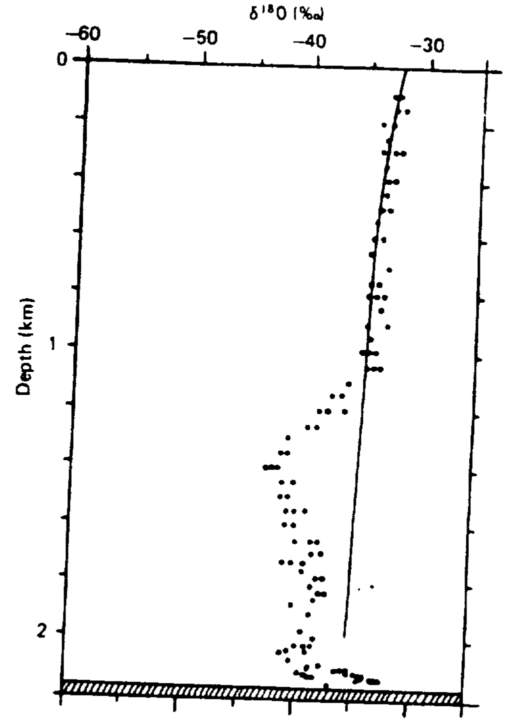

Fig. B. Isotopic data and correction for flow, Byrd Station, Antarctica (Budd and Young, 1983).

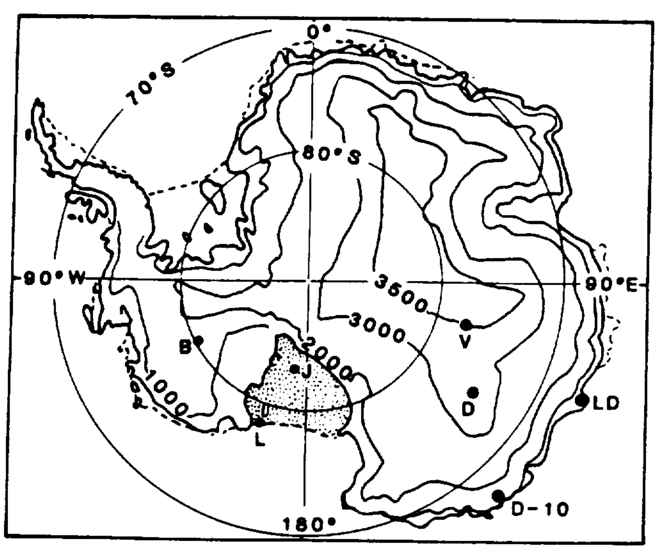

FIG. 1 Localities of Antarctic cores discussed. B: Byrd Station; D: Dome C; D-10: Terre Adélie, core D-10; L: Little America V; LD: Law Dome, cores BHF, BHC-1, BHC-2; J: J-9; V: Vostok.

Fig. C. Locations of Antarctic cores (Grootes and Stuiver, 1987).

Figure 3.11. Mean near-surface δ values plotted against mean air temperature at ground level. Key to symbols shown on diagram. δ values are from Lorius *et al.* (1969, Table 1) and from the following: Dansgaard (unpublished?) for D45, D59, D80; Gonfiantini (1965) for Roi Baudouin; Epstein, Sharp & Goddard (1963) for Little America, Wilkes S2; Dansgaard *et al.* (1973) for Horlick Mountains, Halley Bay and Byrd; Vilenskiy *et al.* (1970) for Mirny; Vilenskiy *et al.* (1974) for Molodezhnaya and Vostok $\delta^{18}O$ values; Epstein & Sharp (1967) for Eights; Aldaz & Deutsch (1967) for South Pole; Picciotto (1967) for Plateau Station; Picciotto *et al.* (1968) for Pole of Relative Inaccessibility; measurements by Lorius (unpublished) on samples from Wolmarans from SA9 and Peel from HB3. Line 1 for inland stations from Lorius & Merlivat (1977), line 2 for coastal stations from data in Budd & Morgan (1973). Points in Area 3 are from Marie Byrd Land and Area 4 from coastal stations. Some temperature values are from Bentley *et al.* (1964).

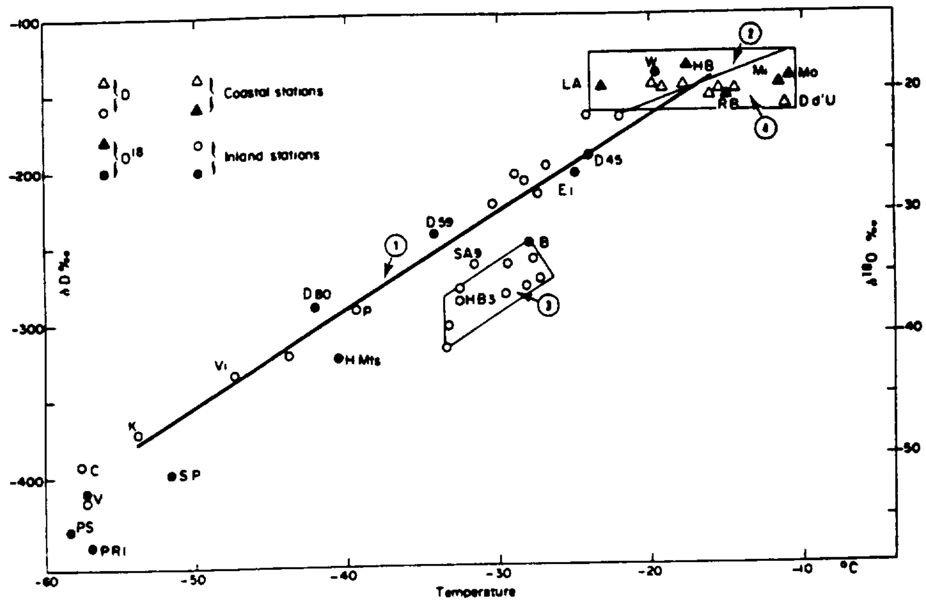

Fig. D. Variation of surface isotopic values in Antarctica (Lorius, 1983). Note anomalous W. Antarctic data.

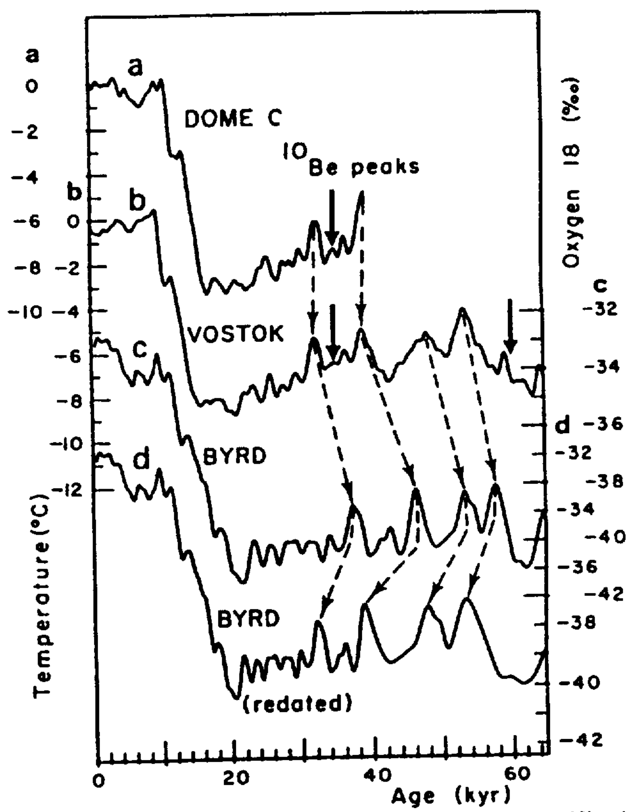

Fig. E. Antarctic-core isotopes and interpreted temperatures, with core correlations based on ^{10}Be spikes (Jouzel and others, 1989).

FIG. 3. Antarctic ice core climatic records over the last 65,000 yr. Dome C (a) and Vostok (b): left scale in °C expressed as a difference with respect to current surface temperature value. Solid arrows indicate observed ^{10}Be peaks. Byrd (c): right scale in $\delta^{18}O$‰ (from Johnsen *et al.*, 1972); time scale from H. B. Clausen and C. U. Hammer (personal communication, 1988). The dotted arrows suggest a possible correspondence of climatic events in the three cores. Curve d, the Byrd record, uses modified time scale obtained using these events as time markers.

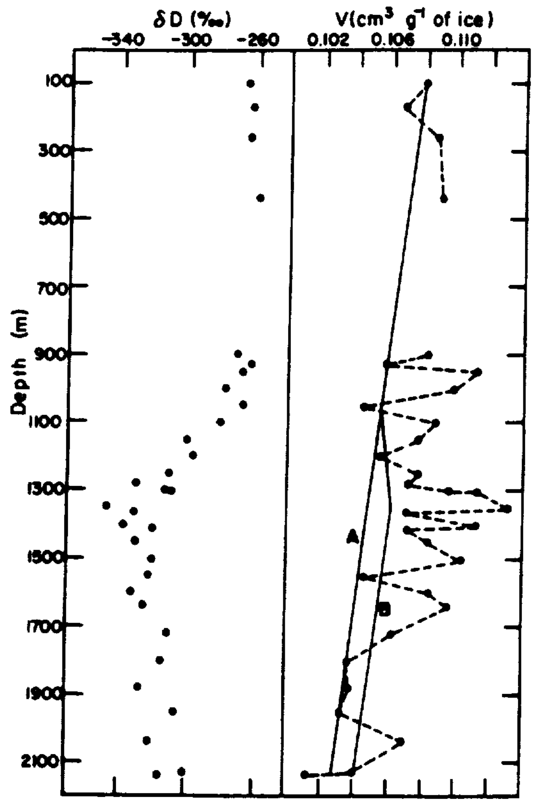

Fig.3. Byrd station. Variations of the mean total gas content V and isotope ratio δ with depth. Mean V values measured at different depth levels are compared with the change of V to be expected by ice flow according to the present flow pattern under constant temperature (line A) and corrected for colder temperatures during the ice age (curve B).

Fig. F. Total-gas-content data at Byrd Station, showing decrease at end of ice age probably caused by ice thickening (Raynaud and Whillans, 1982).

Fig. G. Borehole and core data from Dye 3, Greenland (Dahl-Jensen and Gundestrup, 1987). Borehole inclinometry shows enhanced softness of ice-age ice (which has high dust content and small crystals).

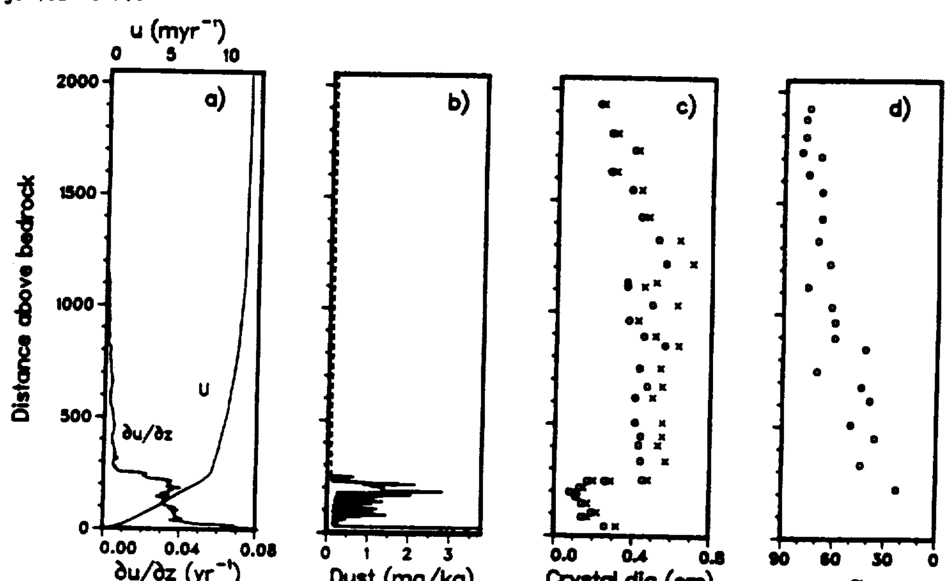

FIG. 2 (a) Profiles of horizontal velocities and shear strain rate derived from the inclinometer surveys of the Dye 3 borehole (Fig. 2a). The ∂u/∂z profile can be divided into three zones: the bottom 25 m of silty ice with very high deformation rates; the immediately overlying 230 m of ice, deposited during the last glaciation; and the uppermost 1780 m of Holocene ice, with lower deformation rates. (b) Dust concentration, the bulk concentration of insoluble microparticles in the ice. (Hammer et al., 1985). In the silty ice the dust concentration is > 20 mg/kg. (c) Crystal diameter (Herron et al., 1985). The circles represent vertical dimensions and the x's represent the horizontal dimensions. (d) Fabric parameter α, the half-apex angles of the cones containing 90% of the c-axis (Herron et al., 1985).