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Microwave Remote Sensing of Snowpack Properties

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Microwave Remote Sensing of Snowpack Properties

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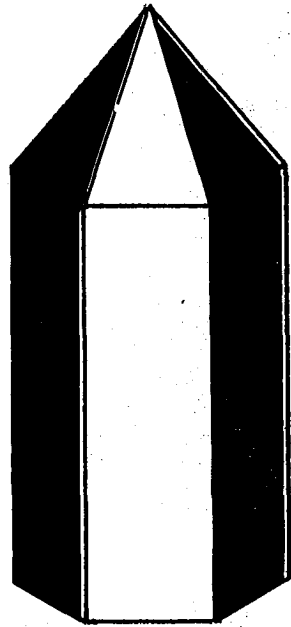Goddard Space Flight Center

NASA

FOREWORD

The use of remote sensing in snowpack studies has initially focussed on the applications of visible and infrared data for determining snow covered area. The techniques for interpreting snow cover from satellite data and using this information in snowmelt runoff prediction were developed and tested in a NASA Applications Systems Verification and Transfer (ASVT) project. Definite improvements in runoff prediction were realized using the satellite snow cover data, and the results were documented in NASA Conference Publication CP-2116. The results of this ASVT indicate the strong need for some additional remote sensing capabilities in surveying snowpack properties. Although visible and infrared data can be used very effectively for measuring snow cover extent in clear weather, an all-weather remote sensing capability should be developed to provide reliable snow cover data through clouds. Secondly, a remote sensing capability should be developed for the basin-wide measurement of snow water equivalent. The combination of snow area and water equivalent yields the snow water volume stored on the watershed.

Such additional capabilities will most likely require an expanded development of remote sensing techniques in the microwave region of the electromagnetic spectrum. Early studies with microwave data have indicated the strong potential for improving snowpack characterization in all-weather situations. Much microwave snow research has been performed, but an organized exchange of research results, technical discussions, and interactions between conventional snow hydrologists and microwave scientists has been lacking. As a result NASA organized a Workshop on the Microwave Remote Sensing of Snowpack Properties, May 20-22, 1980 in Fort Collins, Colorado to bring together the major microwave and snow investigators for a presentation of research results and a series of informal discussions. Fourteen scientific papers were presented at this Workshop over a three day period interspersed with numerous discussion sessions. Informal demonstrations of related remote sensing and conventional capabilities were also presented. The final discussion at the Workshop on May 22 focussed on the initial development of a microwave snow research plan to assist NASA in formulating their continuing research program. The 14 scientific papers and the transcription of the final discussion session are published in this document.

Session chairmen for the Workshop were M. Martinelli, Jr., Rocky Mountain Forest and Range Experiment Station, Fort Collins, Colorado; A. Wankiewicz, National Hydrology Research Institute, Ottawa, Canada; D. J. Angelakos, University of California, Berkeley, California; and W. O. Willis, USDA/SEA/AR, Fort Collins, Colorado. E. B. Jones, Resource Consultants, Inc., Fort Collins, Colorado was the Program Coordinator for the Workshop. In addition, Resource Consultants, Inc. provided technical support for the Workshop. Valuable assistance in the preparation of this document for publication and in editorial review was provided by E. B. Jones and J. Sjogren, Resource Consultants, Inc., R. Peterson, General Electric Company, Beltsville, Maryland, and B. Hartline, Goddard Space Flight Center

The fourteen papers and one discussion session in these proceedings are in the same order as presented at the Workshop. In order to expedite the publication of the proceedings, papers were prepared in camera ready format. Each author assumes full responsibility for the content of their paper.

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WORKSHOP
ON THE MICROWAVE
REMOTE SENSING OF
SNOWPACK PROPERTIES

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EVOLUTION OF SNOW SENSORS AND THE POTENTIAL OF

MICROWAVE DEVICES IN OPERATIONAL TELEMETRY NETWORKS

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ABSTRACT

Devices to measure snow depth and water content have advanced a long way since the efforts of Dr. Church in the early 1900's. Today both manually collected snow-course data and telemetered information from SNOTEL sites throughout the western United States are being used to make forecasts of streamflow. Although these conventional methods have proven highly reliable over the years, they still exhibit some shortcomings. Active and passive microwave remote sensing systems recently developed and tested present the potential to eliminate some negative aspects associated with other types of sensors currently in use. Because of their ability to "see" through the snowpack they offer a unique opportunity to improve snowpack measurement techniques. As yet neither the active nor passive system has been developed sufficiently to supplant existing methods in operational telemetry networks. Active microwave sensors are the most promising for the near future and are being used in research and development programs to study snow pillow performance in the SNOTEL system.

INTRODUCTION

The USDA Soil Conservation Service (SCS) coordinates the cooperative snow survey and water supply forecasting program in the western United States, with the exception of California, where the program is conducted by the State Department of Water Resources. The snow survey program serves the needs of a wide variety of water users who receive 65-80 percent of their water supply from mountain snows. Manual measurements from over 1,700 snow courses throughout the west have historically provided information on snow depth and water content once a month from January through June. These measurements serve as indices of watershed snowpack and are incorporated into various forecasting methods to predict snowmelt runoff.

An automated snow telemetry data collection system (SNOTEL) has been pioneered by the SCS to collect hydrometeorological data from over 500 remote sites throughout the mountains of the West. SNOTEL retrieves information twice daily on snowpack water equivalent, total precipitation and ambient air temperature. It uses a relatively new communication system which relies upon meteorite trails for transmission of data. Overall system specifications and design criteria are well documented by Barton and Burke (Ref. 1).

A primary objective of the SNOTEL operating system is to accurately measure precipitation -- both rain and snow -- which falls on the watershed. To accomplish this, each remote site has stainless steel snow pillows and a 30.5 cm (12 inch) diameter storage precipitation gage. These sensors are the

result of an evolutionary process of remote snow sensor development which began with the concept of a rubber pressure pillow to weigh the snowpack reported by Beaumont and Freeman in the early 1960's (Ref. 2). A great deal of designing, testing and evaluating has brought us to the present level of operation. However, the present sensors do not always provide accurate data under all conditions. SCS is constantly striving to improve current sensor designs and looking into other instruments to monitor mountain snowpack more accurately and efficiently. A brief review of the sequence of major technological advancements will help us bring into perspective the state of the art of snow measurement.

SNOW SENSOR DEVELOPMENT

Beginning with Dr. J. E. Church's design of the Mt. Rose snow sampler in the early 1900's the practice of snow measurement changed little until 1960, except for minor modifications. Early in the 1960's a demand for more frequent and timely data became apparent to improve management of limited water resources. As a result, the SCS undertook a program to develop operational snow sensors which could provide continuous measurements of the snowpack water content during both accumulation and depletion. The snow pressure pillow was an outgrowth of that program. Many other methods of measuring snowpack have been investigated since that time.

PRECIPITATION GAGES

Precipitation gages in a variety of forms have been used to measure snowfall. The basic method by which this is accomplished is by using an antifreeze solution to melt the snow when it falls into the gage. The gage then weighs the liquid content. The gages have significant disadvantages: A tendency to catch less than the actual amount of precipitation which falls in the form of snow; where gages have orifices smaller than 30.5 cm (12. in) in diameter there is greater chance for capping in intense snowfall events; and they are unable to provide information on the rate of snowpack melt -- an important factor in predicting runoff hydrographs during the spring. These inadequacies make the exclusive use of precipitation gages as snow sensors undesirable. Securing accurate precipitation data is hindered by winds in excess of 1 m/sec and cannot completely be compensated for by wind shields (Ref. 3).

Recently I conducted a study of SNOTEL data sites in Colorado and New Mexico. Monthly precipitation gage catch and snow pillow catch were compared at 51 SNOTEL sites for January-March during 1976-1980. The study showed that at 45 percent of the sites the pillow recorded a greater catch than registered by the precipitation gage. Significant differences were observed in locations of heavy snowfalls.

SNOW TUBE SAMPLES

Although coring snow samplers similar to the federal sampler have been used as the standard for over 50 years, they do not measure snowpack water content accurately in all conditions. Work and Brown (Ref. 4 and 5) have shown the federal sampler overweighs actual water equivalent in an increasing amount--ranging from 2 to 12 percent--with increased density of the snow.

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Recent work by Farnes et al.(Ref. 6) suggests the overmeasurement of the federal sampler may be related to the amount of snow water equivalent, cutter diameter as well as density. However, because the federal sampler produces consistent results in highly diversified snowpack environments, it remains the mainstay of the manual measurement program. High personnel and travel costs associated with large numbers of measurements and the destruction of the sampling site when repetitive measurements are taken at the same location are major disadvantages of this type device.

NUCLEAR TECHNIQUES

Nuclear techniques which rely on the attenuation of radiation by the snowpack to determine water equivalent provide reliable measurements of snow water content under most conditions. Snow sensors which fall into this category can be divided into single point devices such as the Radioactive Snow Gage (RSG) developed by Idaho Industrial Instruments (Ref. 7) and large areal integrating techniques instituted by the National Weather Service (Ref. 8). The single point method is attractive because it allows an investigator to discern information about the density of specific layers within the pack as well as the overall pack depth and water content. One disadvantage of the RSG system is the tendency for under measurement in late spring when melt holes develop around the support poles. The gamma ray survey technique used by the National Weather Service is based upon differences in the background radiation level between when the ground is snow covered and when it is snow free. Measurements are taken from an aircraft along pre-defined transects. Corrections must be made for both soil moisture variations and atmospheric radon concentrations. This technique appears to produce reliable mean areal snow water content estimates for the shallow snowpacks found in the relatively flat topography of the Upper Midwest. It does not appear to have general application in the mountains at the present time.

MICROWAVE-ACTIVE AND PASSIVE

Remote sensing techniques employing microwave devices for determining properties of snowpack including water content fall into two systems--active and passive. Passive systems relate snow depth and water content to relative microwave brightness temperatures. This is generally accomplished from scanning radiometers aboard aircraft or satellites. Hall (Ref. 9) demonstrated that decreasing microwave brightness temperatures at wave lengths from 0.8 cm to 21.0 cm were generally positively correlated with increasing snow depths in two study sites in Colorado. However, soil moisture conditions and free water in the pack exerted strong influence and markedly affected this relationship. To a lesser degree changes in crystalline structure of the pack also affect the accuracy of this method. The multifrequency approach advocated by Hall helps some in resolving problems associated with free water in the pack, but the technology is not sufficiently developed to consider this a viable snow sensor for operational purposes at the present time. Successful development would provide a much needed tool for estimating total volumetric snowpack accumulation on a basin-wide basis. It also has potential for measurement of near surface soil moisture conditions which would help predict runoff rates from melting snow.

Active microwave radar sensors such as the FM-CW system described by Boyne and Ellerbruch (Ref. 10) not only provide snowpack water equivalent measurements to +5 percent accuracy but also give information on snowpack stratigraphy; this device relates amplitude response of microwave signals to snow depth and water equivalent. The devices used in studies by Boyne show promise as remote snow sensors, but additional development and testing are needed to overcome the present instrument's inability to provide an accurate measure of snowpack water content once the snowpack becomes isothermal with free water and when prominent stratigraphy vanishes. A potential solution to this dilemma may be to lower the sweep frequencies from the 8-12 gigahertz range to the 2-12 gigahertz range in the FM-CW system. It is not known how this will affect the sensor's ability to penetrate deep packs. The low power requirements give this type system definite potential as a remote snow-sensor once it is proven operational throughout the entire snow season and more specifically during the melt phase. It is not affected by bridging (which plagues snow pillow systems) and lag of precipitation registration. Its performance in a remote mode with data being transmitted via telemetry has not yet been established.

SNOW PILLOWS

Snow pillows are the most common continuous snow sensor in the field today. Their ability to accurately sense snowpack water content has been demonstrated thoroughly in studies by Brown and Bartee (Ref. 6 and 11). Snow pillows fall into two categories: steel tanks and hypalon type bladders. These are filled with antifreeze solution and they sense the water content of the snow by acting as a hydraulic weighing platform for the column of snow that accumulates on it. The weight of the snowpack is converted to a measure of water content by one of two methods. One method is to record the change in hydraulic head on the pillow by means of a stilling well and float system attached to a continuous recorder; the other method is to convert the hydraulic pressure to an analog signal via a pressure transducer which is either recorded onsite or transformed to a digital value for transmission where it is converted to a water equivalent value. Over 2,000 pillow years of record have shown that the snow pillow when properly installed and maintained is a highly reliable sensing system in most conditions. However, certain operating characteristics of snow pillows detract from any claim that they are the ultimate snow sensor.

One would like to have a snow sensor which responds instantaneously to snowfall (loading) or melt (unloading). Cumulative experience with snow pillows has revealed that such is not always what occurs. In some cases, bridging caused by ice layering in the pack prohibits transmittal of full snowpack weight to the pillow, resulting in under measurement. This tendency appears to be less prevalent early in the season and increases up to the time of isothermal conditions. Ice layers can produce a registration lag in snowfall events, recording the full effect of a storm over a several day period after storm activity has ceased. Highly variable temperature regimes inside shelters housing recording instruments often lead to fluctuations in sensor readings which make it hard to interpret water equivalent values. This condition is encountered more frequently at lower latitudes and during high sun angles of spring. Rain on snow events are sometimes difficult to

evaluate when some or all of the rainfall migrates through the pack. This phenomenon occurs most often in ripe snowpacks or in packs which are relatively warm.

Pillows provide very little information on the internal structure of the pack or the rate of metamorphism which constantly occurs. In addition a steadily accumulating amount of evidence developed from pillow dig-outs in California, Idaho, and Colorado show pillows overweigh as much as 20 percent in some circumstances. Reasons for the error are not understood.

SNOTEL SYSTEM STATUS

Since the start of SNOTEL in 1975, a tremendous amount of knowledge has been gained in installation and operation of a huge state of the art remote data collection network. Some problems have been solved, but many remain. With virtually all 511 sites installed and over 250 of these presently reporting on telemetry, a critical evaluation of the system's performance is underway. The meteor burst communication system has met most expectations and the remote sensors have performed well for the most part. However, questions remain unanswered: whether snow pillows are fulfilling all the needs of users who require accurate measurements of snowpack water equivalent in highly diverse snow accumulation environments; under what conditions they function well and under what conditions readings are unacceptable? A five year cooperative research effort by SCS and Colorado State University has been initiated to help answer these and other questions.

This research is aimed at answering questions on the absolute accuracy of snow pillow measurements as well as the relative accuracy compared to such devices as the federal sampler, glacial sampler, snow pit profiling, and FM-CW microwave radar system. Procedures involve extensive laboratory and field tests in Colorado and other western states to assure that a wide range of snowpack conditions is sampled. Investigations with an FM-CW microwave radar system will be an integral part of this program. Results will be evaluated and recommendations made for improvements to the existing system. Recommendations may involve development and deployment of new sensing systems to complement the present SNOTEL data array. However, new devices must be superior to snow pillows or be able to offer an element of knowledge about the snowpack not currently provided by snow pillows. Demonstrating superiority to pillow systems may be difficult, as evidenced by the figures 1-4 which show the performance of the Upper San Juan SNOTEL site in southwestern Colorado for the past 4 years.

Figure 1 shows the relationship between visual manometer readings and the average of four control samples taken with a standard federal sampler near the pillow. It also reflects an increasing over measurement by the pillow in relation to the control samples. However, when the pillow manometer readings are corrected for a fluid specific gravity of 0.92, the relationship improves markedly, as shown in figure 2. The simple correlation coefficient between the control samples and the pillow readings is 0.996 with an average error of 2.96 cm over a range of 140 cm. The regression line in figure 2 is close to a perfect 1:1 relationship, a desired result for direct comparability with historical snow-course records. Figure 3 shows the rela-

tionship between telemetered snow water equivalent and manually recorded pillow manometer readings adjusted for specific gravity. The relationship is extremely consistent exhibiting a simple correlation coefficient of 0.999 and an average error of only 0.71 cm. About a 3 percent over measurement error is indicated by the telemetered values compared to manometer readings. This may be attributed to either a systematic bias in the pressure transducer or in the transceiver or both. Since the control samples were taken with a federal sampler and overweighing is characteristic of this method, it is instructive to compare the adjusted pillow manometer readings to the control samples corrected for over measurement, according to the work of Brown (Ref. 5). When this is done for the Upper San Juan data (shown in Figure 4) the pillow, in relation to the federal sampler, has a tendency to overweigh at increasing depths and densities.

Analysis of these data from this single site leads to the following conclusions: 1) snow pillow data for an 2.44m x 3.05m (8 ft. x 10 ft.) array of metal pillows compares closely with federal sampler control measurements, 2) the 2.44m x 3.05m (8 ft. x 10 ft.) pillow array overweighs in relatively the same proportion as the federal sampler over measures, 3) historical snow course records at the site obtained with a federal sampler can be compared directly with telemetered SNOTEL data, 4) pillow measurements of water equivalent are accurately transmitted by the SNOTEL system at this site, and 5) the observation that pillows catch more than precipitation gages in heavy snowfall areas may be partially related to pillow over measurement characteristics.

Because of their long-term consistency the federal sampler and snow pillow will remain the standard for judging other sensors. Until a clearly superior device becomes available they will be used in most operational programs and serve as ground truth for future research and development programs.

SUMMARY

Demand for high quality snowpack and precipitation data in the mountains of the West continues to accelerate. The SNOTEL system has proven its ability to provide continuous daily measurements of snowpack water equivalent and total precipitation over a wide range of snowpack environment and operating conditions. Snow pillows, the standard snow sensors in the SNOTEL system, have produced over 2,000 pillow years of data. These data establish the pillow system senses water equivalent values, which are generally quite close to manual measurements taken by a federal sampler and are thus directly comparable to long term historical records available from snow courses. Pillow systems are relatively simple to operate and maintain and have produced a large volume of usable data.

Snow pillows do not work in all snowpack conditions all of the time; they are prone to measurement error induced by ice lenses, melt crusts, and wind crusts within the pack. Registration lag sometimes results in pillows showing snowpack accumulations days after a storm has ceased.

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Remote sensing devices which measure both snow depth and water content of the snowpack possess the potential to provide information which could resolve some problems inherent in snow pillows. Active microwave systems appear to hold the most promise for the immediate future. Such systems have demonstrated an ability in certain circumstances to accurately sense snow depth and water equivalent in a research mode. If these systems can be developed into operational tools which will function throughout all of the snowpack conditions during accumulation and melt they will have passed the first major hurdle toward deployment in data acquisition networks.

Because of their unique ability to "see" through the snowpack, active microwave radar devices such as the FM-CW system offer a tool to research reasons for specific pillow system erratic behavior. It will also be useful in documenting performance of field tests of various snow pillow configurations. Substantial development and testing are anticipated before the FM-CW system will stand alone as a snow sensor in an automated telemetry network such as SNOTEL.

Until several problems are resolved, passive microwave systems which relate snow depth and water content to equivalent brightness temperatures appear to have only limited application as an operational snow hydrology tool in the mountains for the forseeable future. Sensitivity to free water in the snowpack and to changes in snowpack crystalline structure through temperature metamorphosis and the effects of soil moisture are all concerns. When these are resolved passive microwave inventories of basin snowpack water equivalent may be extremely valuable as supplementary data to existing sources.

Due to the need for accurate and timely data on snowpack water content, the SCS looks forward to new developments in technology in the field of remote sensing to augment the SNOTEL system. SCS has a commitment to research in this area and will provide assistance wherever possible to efforts aimed at improving measurement techniques.

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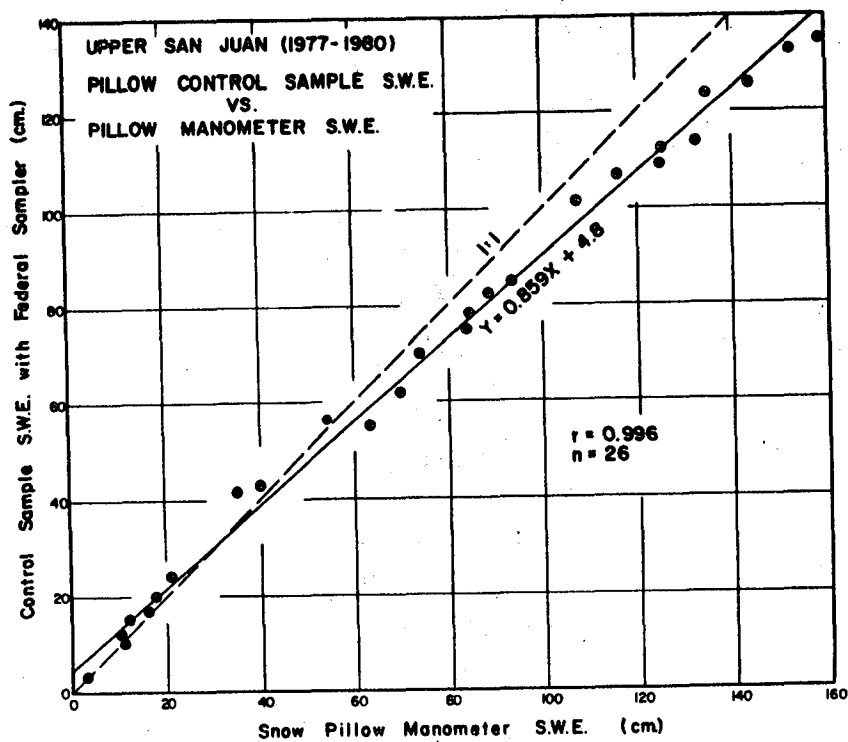
Figure 1 -- Comparison of control sample snow water equivalent (S.W.E.) taken with a Federal sampler and S.W.E. sensed by a 2.44m x 3.05m pillow array at Upper San Juan SNOTEL site for the 1977-80 period.
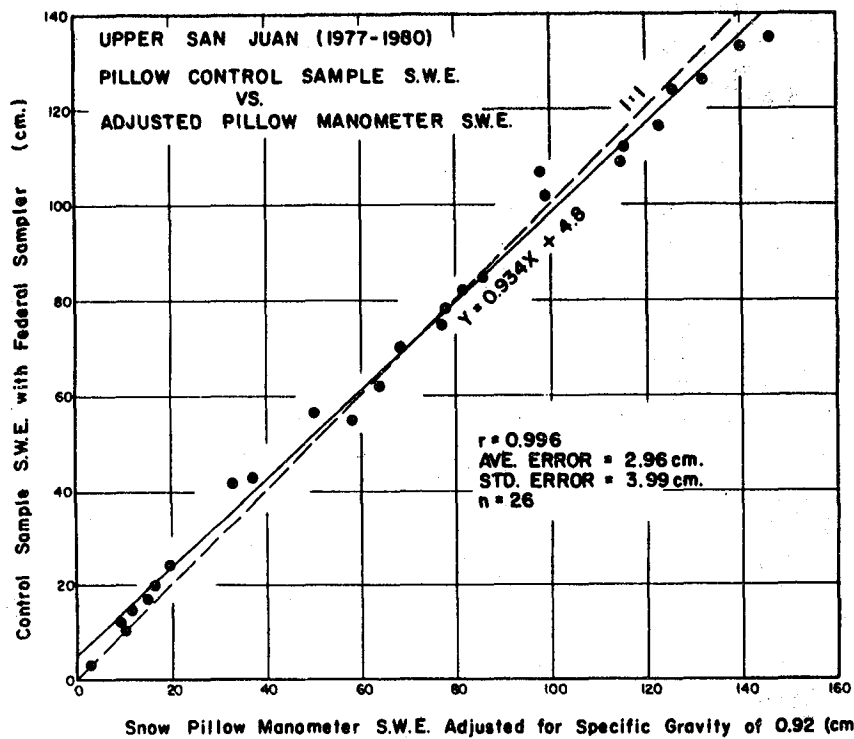
Figure 2 -- Comparison of control sample snow water equivalent (S.W.E.) and S.W.E. measured by the pillow manometer adjusted for a fluid specific gravity of 0.92.

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Figure 3 -- Comparison of snow pillow manometer snow water equivalent (S.W.E.) values transmitted by the SNOTEL system.
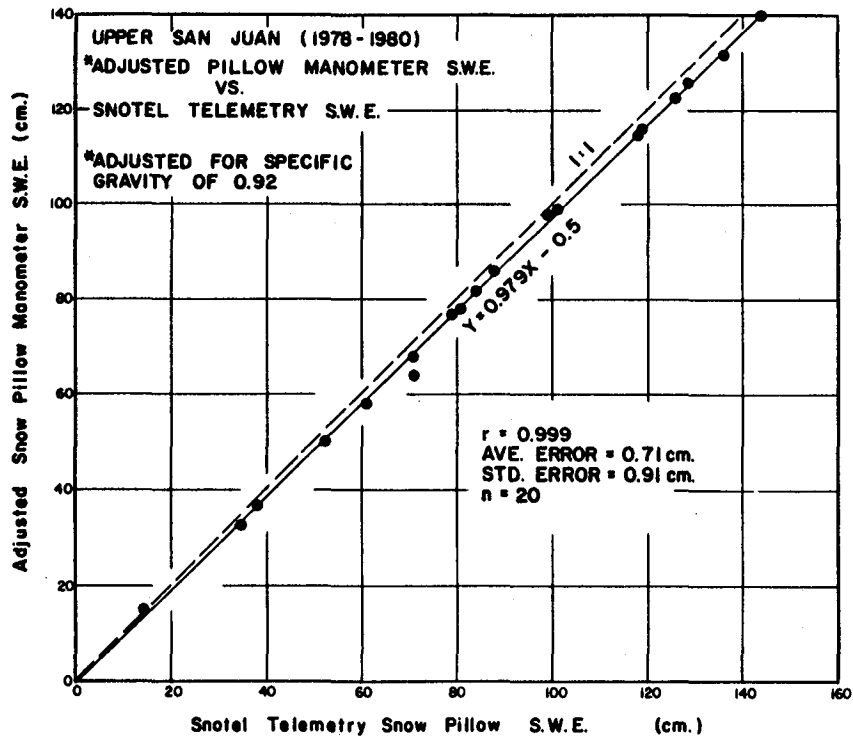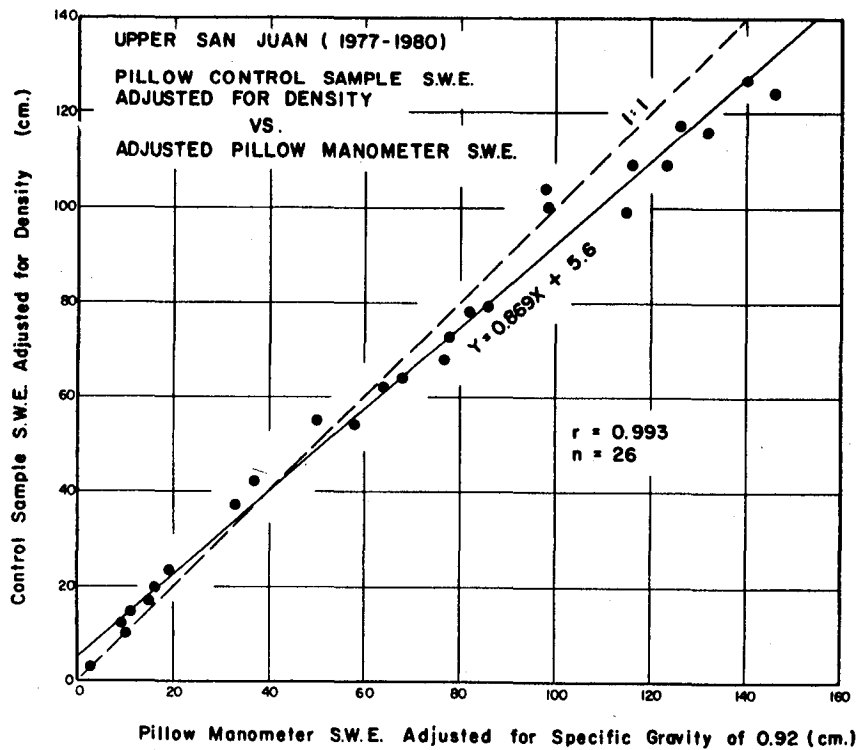

Figure 4 -- Comparison of control sample snow water equivalent (S.W.E.) adjusted for over-measurement error and pillow manometer S.W.E. adjusted for a fluid specific gravity of 0.92.

CALIFORNIA'S TRANSITION FROM CONVENTIONAL SNOWPACK

MEASUREMENTS TO A DEVELOPING REMOTE SENSING

CAPABILITY FOR WATER SUPPLY FORECASTING

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ABSTRACT

California's traditional Snow Survey Program and water supply forecasting procedures are described, and a review is made of current activities and program direction on such matters as the growing statewide network of automatic snow sensors; restrictions on the gathering hydrometeorological data in areas designated as wilderness; the use of satellite communications, which both provides a flexible network without mountaintop repeaters and satisfies the need for unobtrusiveness in wilderness areas; and the increasing operational use of snow-covered area (SCA) obtained from satellite imagery, which, combined with water equivalent from snow sensors, provides a high correlation to the volumes and rates of snowmelt runoff. Also examined are the advantages of remote sensing; the anticipated effects of a new input of basin-wide index of water equivalent, such as that obtained through microwave techniques, on future forecasting opportunities; and the future direction and goals of the California Snow Surveys Program.

INTRODUCTION

The snow survey and water supply forecasting program in California differs in several ways from similar programs elsewhere. First, it is coordinated by the State, with field data collection and associated research almost completely supported by cooperating federal, state, and local water management agencies. Second, the program's need for remote sensing capabilities to monitor the snow zone is more urgent than in other states because of the large and contiguous designated and proposed wilderness areas, which encompass almost three-quarters of the central and southern Sierra snow zone. See Figure 1, "Existing and Proposed Wilderness Areas in High Yield Snow Zone of Central and Southern Sierra." (Ground access for placement of data collection instrumentation is administratively restricted by the Wilderness Act, Public Law 88-577.)

Because state coordination of the snow survey program in California was set up over 50 years ago through the efforts of cooperating agencies, we have been particularly aware of our charge to coordinate, standardize, and advance the technologies necessary to continually improve water supply forecasting capabilities. And because the advent of automatic snow measurement instrumentation coincided with passage of the Wilderness Act, there is this additional reason to seek new ways to collect field data without the need for a tight network of field data sites, which so far has been unacceptable to the U.S. Forest Service within wilderness areas.

Consequently, support of ongoing research has become an integral part of our program, as illustrated by such recent endeavors as: attempting to incorporate long-range weather forecasts into runoff forecasting procedures (ref. 1 and 2); the use of satellite derived snowcovered area measurements in forecasting snowmelt (ref. 3 and 4); and the installation of an Earth Receive Station to directly monitor a growing network of snow Data Collection Platforms (DCP's) through an assigned channel on the Geostationary Operational Environmental Satellite (GOES).

An understanding of the operational needs of the snow survey and water supply forecasting programs, particularly our needs to perfect remote data collection techniques, may be helpful in influencing research directions toward the practical applications required by water management agencies.

THE CALIFORNIA COOPERATIVE SNOW SURVEYS PROGRAM

Traditionally, snow surveying, as devised by Dr. James E. Church in the early 1900's, has been the major data collection input into snowmelt runoff forecasting efforts. In California, more than 40 cooperating agencies collect over 1,100 monthly snow water equivalent samples on 317 snow courses each winter season. More than 150 aerial snow depth marker photographs are secured each winter by overflights to provide supplemental data from remote areas.

Other input collected by Department of Water Resources (DWR) includes a variety of data on water import, export, diversion, evaporation, consumptive use, etc., needed to compute the natural (unimpaired) runoff to date of the 25 major snowmelt rivers and 20 tributaries for which unimpaired runoff is forecast. Storage data from 143 reservoirs is collected at the end of each month to complete the necessary input to the runoff calculations and to monitor the status of California's water supply reserves.

In essence, the runoff forecasting procedure consists of computing the existing hydrologic balance in each river basin and then extending it into future months on the basis of snowmelt runoff potentials and median weather conditions. The primary input is a basinwide index of snow water equivalent as derived from the snow course measurements. The multi-regression equation method of forecasting April through July snowmelt volumes has been in use for almost half a century. The accuracy of these forecasts has been increased through the addition of factors that better define basin priming, i.e., antecedent precipitation, previous year's runoff, soil moisture, etc.

More recently, increased attention has been given to construction and manipulation of basin hydrologic models that can be updated rapidly as conditions on the watershed change. At present, forecast services to program cooperators consist of the basic April through July volumetric forecasts for the 25 snowmelt basins, computed on the first of February, March, April, and May -- plus weekly updates of these forecasts for eight of California's major rivers (based on daily snow water changes obtained through telemetered automatic snow sensors). An additional service involves use of hydrologic models on the Kings and San Joaquin Rivers, which simulate future flow based on various temperature regime inputs.

STATUS OF DATA COLLECTION ACTIVITIES

RADIOACTIVE SNOW GAGES

Snow data collection methods began to change in the 1950's with the first attempt to develop automatic snow sensors. Single-point radioactive gages using Cobalt 60 attenuating to an overhead rate meter were tried. Neutron soil moisture gages were adapted to snow density measurements in a 1962 U. S. Forest Service study supported by DWR. These studies led to the ultimate in radioactive gages, the Gamma Transmission Profiling Snow Density Gage, developed by the U. S. Forest Service, Southwest Forest and Range Experiment Station, and co-sponsored by DWR and the former U.S. Atomic Energy Commission. Volumetric comparison tests conducted by our staff, both at the Central Sierra Snow Lab and at our Alpha Instrument Evaluation Site, show this profiling gage measures to within 2 to 3 percent of true snow water equivalent.

A general statewide distribution of these gages for operational snow data collection is not being considered, however, because of high initial costs, maintenance costs, level of technical expertise required, and land use permit restrictions on use of unattended radioactive sources (in this case Cesium 137).

Another type of radioactive gage takes incremental measurements of total snowpack water equivalent by use of collimated Cobalt 60 sources arranged to "zig-zag" the gamma emissions from sources in one mast to radiation detectors in a parallel mast. The original prototype of this gage, using Cesium 137 sources, was tested at our Alpha site. Although its accuracy is acceptable, the same operational restrictions exist as with the U. S. Forest Service profiling type gage.

PRESSURE TYPE SNOW SENSORS

The Alpha Instrument Evaluation Site was established in 1965 following successful development of the pressure sensitive snow pillow by the U. S. Soil Conservation Service (SCS) at Mt. Hood. The Alpha site was established to provide DWR and cooperating agencies with the means to test and evaluate automatic snow sensing and related instrumentation under the same weather and snow conditions that would be encountered during actual operation. Parallel objectives were to encourage uniformity of operations, reduce false starts or duplication of effort among cooperating agencies engaged in snow data measurements, and assist cooperators in establishing reliable snow sensor networks. Results of 10 years of research at Alpha were published in 1976 (ref. 5).

Recent studies have further advanced the snow survey program to its present level of data collection and use and have begun to set the stage for eventual transition to completely automated data acquisition. As of May, 1980, 61 automatic snow sensors are operating in California.

The present snow sensor consists of either a 3.66 metre (12 foot) diameter rubber pressure pillow, or four 1.22-by-1.52 metre (4-by-5 foot)

stainless steel pressure tanks. These sensors are installed according to
standards developed at the Alpha Site. Either float operated recorders or
pressure transducers are used to interface with the telemetry.

To convince the U. S. Forest Service and the National Park Service that
we could install automatic snow sensors unobtrusively in wilderness areas, we
have designed and tested Data Collection Platform (DCP) configurations
specifically for that purpose. See Figure 2, "Standard Automatic Snow Sensor
Installations." The National Park Service has approved the installation of a
skeletal network of wilderness type snow sensors in Yosemite, Sequoia, and
Kings Canyon National Parks. We are still working with the U. S. Forest
Service to gain approval for installations in wilderness areas.

SATELLITE COMMUNICATION

The use of satellite communication of data is considered a major advance
for our program, particularly for collecting data from remote areas, where
ground radio paths are difficult to establish. In 1978-79 we successfully
tested a DCP in cooperation with NOAA/NESS, the operators of the GOES
satellite. Subsequently, DWR started a capital outlay program to place 30
DCP's into our snow sensor network in cooperation with the agencies who fund
the field installation and maintenance. In addition, the State was assigned a
data channel on the GOES satellite and, with the Department of Forestry, we
obtained direct access capability by installing a disk antenna and receive
terminal on our headquarters building in Sacramento. This system is now in
operation with the first three DCP's installed and transmitting. It is
expected that program cooperators and other agencies will install DCP's and
use the State's satellite link.

Progress is also being made in exchanging snow data and other information
by computer to computer links or through auxiliary terminals. As an example,
we now access the SCS Sno-Tel system computer in Portland for all California
data transmitted by meteor burst relay to the SCS data bank. This presently
involves 14 snow sensors located in east side Sierra watersheds that are
partially tributary to Nevada.

SNOWCOVER AREA BY SATELLITE

An operational study was conducted for four years (1975-1979) under a
NASA contract to evaluate the application of snowcovered area (SCA) obtained
from satellite imagery as an additional parameter in California's water supply
forecasting procedures. Photographs of the snowcovered areas were supplied by
NASA (Landsat) and by NOAA (GOES). Translation of SCA from photographs was
accomplished by direct overlay, or by use of a zoom transfer scope that
optically matched the satellite image to the base map on which the SCA was
traced. The percent of SCA in a basin was computed and an "effective
snowline" determined.

Because the use of SCA was found to have a significant relationship to
snowmelt runoff from April through July, as the snowpack melts and changes its

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boundaries, California is continuing to adapt its forecast procedures to include the use of SCA. There are benefits to be gained by the combination of SCA from satellite imagery and water content of the snowpack from snow sensors, coupled with sophisticated telemetry and computer capability. These improvements, and others yet to come, will lead to entirely new and faster means of surveying the snowpack, and to more frequent and timely forecasts of water supply.

ADVANTAGES OF REMOTE SENSING TO OPERATIONAL USERS

SOLAR ALBEDO

Measurement of solar reflectivity of snow may prove correlative with snow density. Measurement of albedo for purposes of accurately quantifying snow density would be a valuable addition to snow measurement technology, especially if such measurements could be accomplished by remote sensing. In California, with its marine climate and periodic warming of the snowpack, density estimates would be particularly useful throughout the snow season, provided it could be coupled with other data to derive total water equivalent of the snowpack.

SATELLITE DERIVED QUANTITATIVE PRECIPITATION ESTIMATES

Quantitive precipitation estimates from satellites using infra-red imagery, together with high resolution visible imagery, have been applied to predict flood intensities. Such information is also considered potentially useful in predicting increments of snow water accumulation or melt when surface temperatures below freezing can be delineated. We recommend this capability be enhanced with both more responsive satellite sensors and increased data analysis, and, if needed, ground truth correlative studies.

BACKGROUND RADIATION FOR SNOW SURVEYS

The measurement of changes in natural gamma radiation due to accumulation of snow over the land appears to offer much promise as a rapid method for snow surveying. Although the procedure is primarily being used on large level areas, as on Russian and Canadian plains, continued experience with the method may lead to its refinement and subsequent application to smaller areas, such as nearly flat mountain meadows, as well as to deeper snowpacks than those presently being measured with the technique.

MICROWAVE MONITORING OF SNOW

A considerable amount of work has been put into investigations of electromagnetic measurement of snow. The results have been encouraging, and subsequent research efforts will expand our understanding of these techniques. Present limitations associated with defining snow properties by microwave sig-

natures may eventually be overcome by higher resolution instruments. However, basic techniques based on the use of current data sources have already provided a broad understanding of how microwave emissions can be used to measure snow wetness, density, and depth.

California's snow survey and water supply forecasting program would be particularly benefited by the successful development of microwave sensing of snow water equivalent. For mountainous watersheds, the perfection of either active or passive microwave systems to monitor snowwater at selected locations would resolve the difficulties now encountered with field installations of instrumentation, including the problem of restrictions in wilderness areas.

FUTURE OF THE CALIFORNIA SNOW SURVEYS PROGRAM

Research in snow data collection and analysis by various methods will lead to new and more useful procedures in water supply forecasting. Such advancement will result in increased operational efficiency at all levels with a resulting positive economic impact on every aspect of water conservation and use.

From our observations of some of the current research projects measurement of snow characteristics, the most elusive element is the ability to accurately sample deep pack (or total pack) parameters, especially in mountains. An associated element is higher resolution of data from small areas in the mountainous terrain rather than the present practice of averaging data from scans of large areas. These elements are essential to our particular activities in water supply forecasting.

There is an opportunity through snow research to "mix" data types in new combinations in order to characterize the snowpack, its melt regimens, and its water producing potential. One combination could consist of radar observed depths combined with: (1) satellite derived snowcovered area, or (2) snow density or water equivalent as determined by albedo or microwave radiometer determinations, or (3) radiation measurements.

The role of satellites in measuring snow properties may be expanded to yield the basic parameters necessary for our work: snow depth, density, and water equivalent. The direction of our California Cooperative Snow Surveys Program is seen as one that will be kept sufficiently flexible to quickly absorb new concepts as they prove to be beneficial. This results in our operating more than one program at a time -- as we are doing now. Manual snow surveys, taken once per month, are still the backbone of our data input to volumetric snowmelt runoff forecasts. But we also make use of daily automatic snow sensor data and satellite imagery in some of our forecasting update procedures. New levels of forecasting would be opened up if such things as microwave techniques could be more completely understood and eventually applied to snow surveying. The result could be the elimination of many of the expensive on-ground devices, increased access to remote areas, increased timeliness of data (regardless of weather), and improved water supply forecasting capabilities and services. To the extent possible, we will be supportive of future research projects and the analysis and application of new forms of data.

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Figure I

CALIFORNIA COOPERATIVE SNOW SURVEYS

EXISTING AND PROPOSED
WILDERNESS AREAS
IN
HIGH YIELD SNOW ZONE OF
CENTRAL AND SOUTHERN SIERRA

GRANITE CHIEF

DESOLATION

LADEUX

SACRAMENTO ◉

MOKELUMNE

CARSON ICEBERG

EMIGRANT

HOOVER & EXTENSION

NEVADA

YOSEMITE

MINARETS

MINARETS SUMMIT

MERCED ◉

SIERRA

NOTE: 5,000 FOOT ELEVATION LINE
INDICATES LOWER LIMIT OF SNOW ZONE

SEQUOIA-KINGS

MONARCH

FRESNO ◉

JOHN MUIR

CREST

GOLDEN TROUT
GRAY MOSES

DOMELAND

BAKERSFIELD ◉

Figure 1

Figure 2

CALIFORNIA COOPERATIVE SNOW SURVEYS PROGRAM

STANDARD AUTOMATIC SNOW SENSOR INSTALLATIONS

STANDARD: NON-WILDERNESS

1. BURIED SNOW SENSOR
2. INSTRUMENT SHELTER
3. MISSILE PRECIP. GAGE
4. ANTENNA
5. SOLAR PANEL
6. TEMPERATURE GAGE

PROPOSED FOR WILDERNESS

1. BURIED SNOW SENSOR
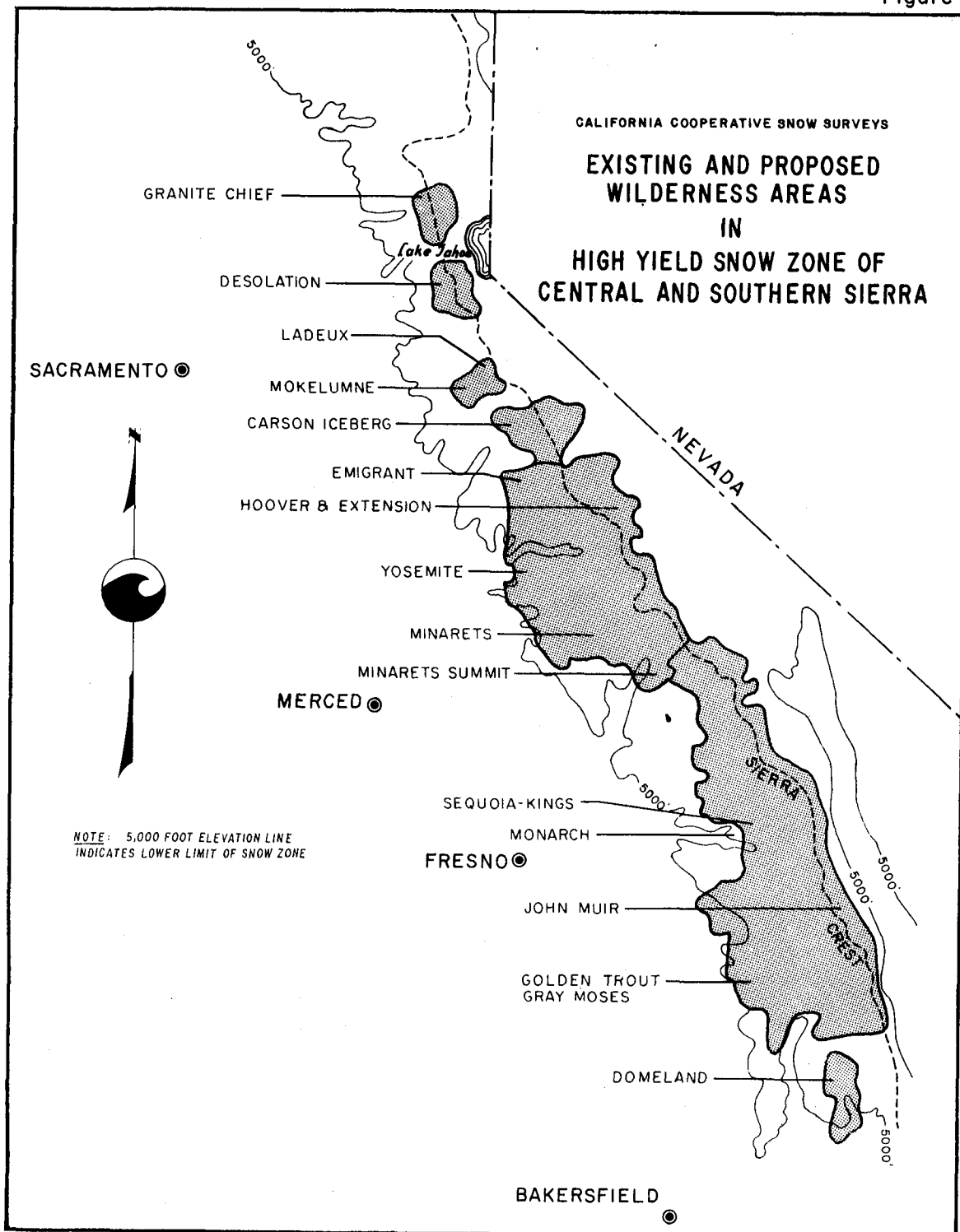2. BURIED DATA COLLECTION PLATFORM
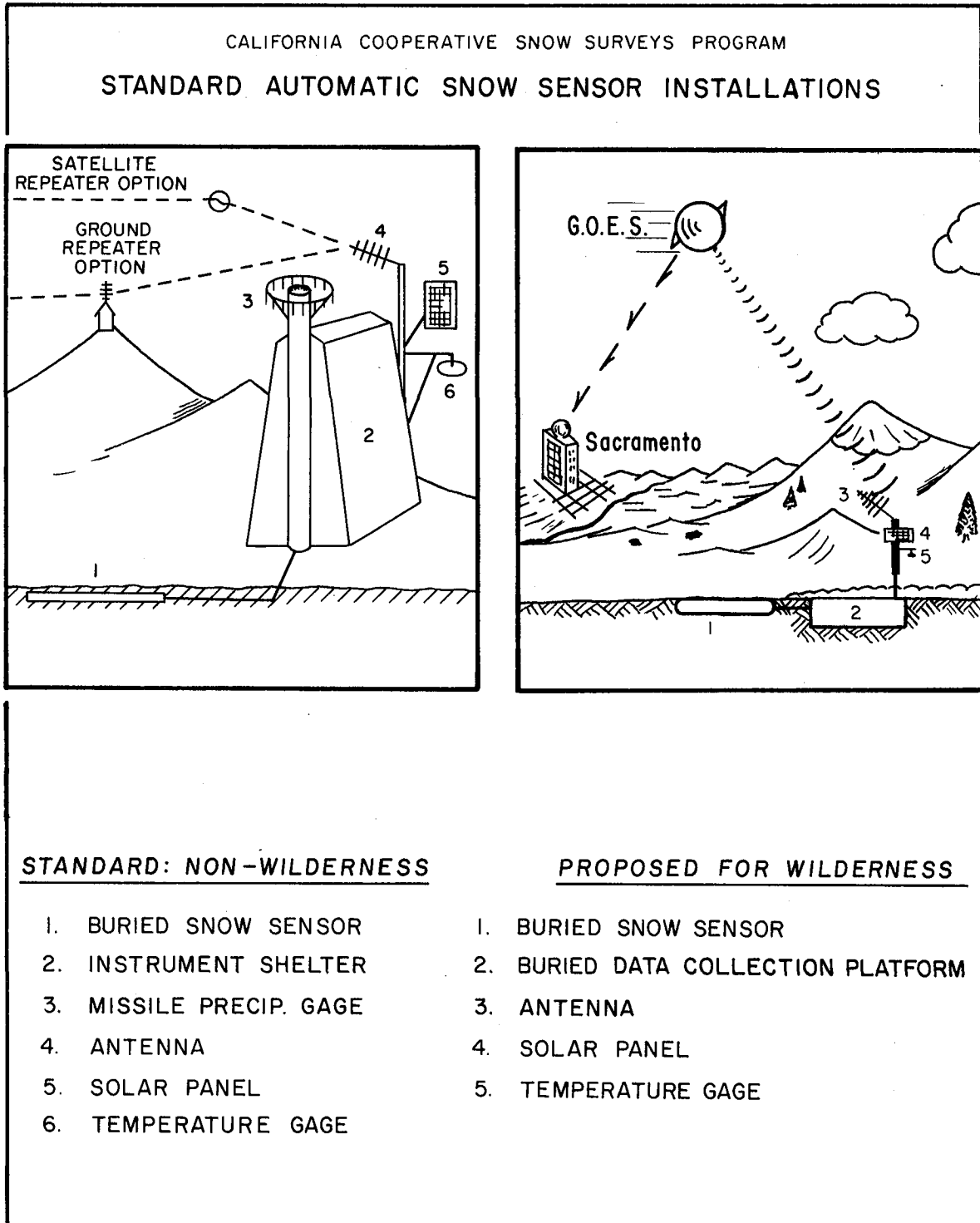3. ANTENNA
4. SOLAR PANEL
5. TEMPERATURE GAGE

Figure 2

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LIQUID DISTRIBUTION AND THE DIELECTRIC

CONSTANT OF WET SNOW

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ABSTRACT

The mixing theory of Polder and Van Santen is revised for application to three cases of wet snow. The dielectric constant is calculated for a range of liquid contents and porosities. These calculated values compare favorably with experimental data for the two cases in which data are available. The application to a snow cover with a heterogeneous distribution of liquid is discussed. The possibility of applying this theory to calculate the imaginary part of the dielectric constant must be explored further.

INTRODUCTION

The dielectric properties of wet snow are important because of the information which can be obtained from measurements in both the megahertz and gigahertz frequency range. Ambach and Denoth (ref. 1) developed a small instrument for measuring the liquid water content of snow at a frequency of 20 MHz. Measuring with a precision of about ± 0.5%, their device allows in situ measurements of the amount of liquid present, a parameter which affects the snow's strength, albedo, rate of metamorphism, and rate of liquid movement.

Snow is generally heterogeneous (i.e., more than one statistical distribution is required to describe properties such as grain size and density) hence large scale measurements are needed to average over the local conditions. This averaging is necessary on the macro-scale in the same sense that averaging is made over many individual snow grains when local measurements are made with a hand held instrument. Information on the larger scale can be obtained with active microwave sensing systems which in principle can provide much valuable information about the snow cover. However, proper interpretation of microwave signals depends upon our understanding of the dielectric properties of wet snow and much remains to be learned about the response of the different types of snow to excitation at microwave frequencies.

In this paper the theory of Polder and Van Santen (ref. 2) is simplified for ellipsoidal particles and is then used to calculate the dielectric constant for wet snow. This approach is based on the observed structure of wet snow. Three distinct types of wet snow are identified and the theory

is applied to each type. This allows the calculation of the dielectric constant for a wide range of conditions, for example, freely draining low density snow and high density slush layers.

THEORY FOR ELLIPSOIDAL PARTICLES

Polder and Van Santen (ref. 2) developed a general theory for calculating the effective dielectric constant of mixtures in which the individual particles or holes are assumed to be ellipsoidal. Denoth and Schittelkopf (ref. 3) showed the applicability of this theory to one type of wet snow without explaining the effect of the shape of the particles or fluid inclusions. This earlier work is continued here with the restriction that two of the principal axes of the ellipsoid are equal. Where the principal axes are a, b and c, a equals b but c may take various values. Thus we are assuming that all inclusions take the shape of spheroids in a continuum and that the continuum may be air, liquid or ice (these cases I, II and III are developed later).

Polder and Van Santen derive depolarization factors (A_i), which for our case (a=b) are

$$A_i = \frac{a^2 c}{2} \int_0^\infty \frac{du}{(a^2+u)\,(c^2+u)^{1/2}\,(i^2+u)} \tag{1}$$

where i represents a, b or c. It is useful to note that for a equal to be,

$$2A_a + A_c = 1 \tag{2}$$

The effective dielectric constant of the mixture (ε') is given by the implicit equation

$$\varepsilon' = \varepsilon_e \left[1 - \frac{1}{3} \sum_j \left\{ V_j (\varepsilon_j - \varepsilon_e) \sum_i \frac{1}{\varepsilon' + (\varepsilon_j - \varepsilon')A_i} \right\} \right]^{-1} \tag{3}$$

where ε_e is the dielectric constant of the continuum, V_j is the volume filling factor of the jth component (air, water and ice), and the second summation is over the three principal axes (a, b and c). Effectively the dielectric constant of the continuum is modified by j inclusions whose volume and dielectric constant is described in the first summation and whose shape is described in the second summation. For snow, the volume filling factors are described by porosity (ϕ), liquid water content (θ), and air content (θ_a), where

$$\theta + \theta_a = \phi \tag{4}$$

Snow density (ρ_s) and water density (ρ_w) are related to the density of ice (ρ_i) and porosity by

$$\rho_s = (1 - \phi)\rho_i + \theta\rho_w \tag{5}$$

The application of this theory is simplified by use of figures 1 and 2 which show the depolarization factors (A_i) and their ratio (m) as functions of the ratio of the principal axes of the spheroidal inclusions (n), or

$$n = c/a = c/b \tag{6}$$

and

$$m = A_c/A_a = A_c/A_b \tag{7}$$

These formulae are now applied to three cases of wet snow.

CASE I: HIGH ϕ AND LOW θ

For the seasonal snow cover the most important case to consider is that of low density, freely draining snow where the liquid content is typically 3 to 7% by volume and the dry density is below that of randomly packed spheres ($\sim$ 550 kg/m^3). This is the "pendular regime" of liquid saturation (ref. 4) where air is the continuum since it occurs throughout the medium in a continuous path. With the ice and liquid inclusions, the effective dielectric constant is given by the implicit equation

$$\varepsilon_a = \varepsilon' \left[1 - \frac{1}{3}(1-\phi)(\varepsilon_i - \varepsilon_a) \sum_j \frac{1}{\varepsilon' + (\varepsilon_i - \varepsilon')A_{j,s}} \right.$$
$$\left. - \frac{1}{3}\theta(\varepsilon_1 - \varepsilon_a) \sum_j \frac{1}{\varepsilon' + (\varepsilon_1 - \varepsilon')A_{j,1}} \right] \tag{8}$$

where ε_a is the dielectric constant of air, ε_i is the dielectric constant of ice, ε_1 is the dielectric constant of liquid water, $A_{j,s}$ is the depolarization factor for the solid inclusions, and $A_{j,1}$ is the depolarization factor for the liquid inclusions, and the summations are over the principal axes (a,b and c). Together with equation (7) this implicit equation in ε' can be solved for any particular porosity, liquid water content, and inclusion geometry.

The ice grains in wet snow at a low liquid content typically form clusters (ref. 5) whose unit cell is shown on figure 3. One grain from the cluster can be approximated by a spheroid as shown on figure 4 with axes (0.706, 0.706, 0.5). We adopt this solid figure of revolution as the solid inclusion in wet snow for Case I. Using the appropriate depolarization factors (0.289, 0.289, 0.422), the dielectric constant for the special case of dry snow ($\theta=0$) is shown on figure 5 versus snow density and porosity. The curve calculated from equation (8) and the data points from Cumming (ref. 6) and Sweeny and Colbeck (ref. 7) are in very good agreement. It must be noted here that the agreement would be nearly identical if we had assumed spherical particles since the effect of the geometry of the solid inclusions is slight as long as the principal axes of the solid inclusions are randomly oriented. Thus a tetrahedral packing of four grains or even an ellipsoidal cluster of many grains could be taken as the solid inclusion for Case I. Although the dielectric constant is insensitive to the shape of the solid inclusions, as shown next, the dielectric constant is sensitive to the geometry of the liquid inclusions.

The geometry of the liquid veins and fillets in three clusters are shown in figure 6. All assume a triangular type of cross-section, the vein with a small aspect ratio and the fillet with a large aspect ratio. Nevertheless, we represent these with the geometry described earlier, comparing our calculated values of dielectric constant with measured values to determine the most suitable aspect ratio. It is assumed here that the aspect ratio is more important than the cross-sectional shape of the inclusion. To test the effect of the shape of the inclusions on the dielectric constant we plot the dielectric constant versus liquid water content for various values of the aspect ratio (n) on figure 7. The largest value of n represents a needle shaped spheroid, the smallest value represents a disc shaped spheroid, and the value of one represents a spherical water inclusion. As shown in figure 6, the liquid water inclusions in low density grain clusters are long, thin liquid fillets and shorter liquid veins. Therefore we expect the aspect ratio to fall somewhere between one and ten, the values for spherical and needle shaped inclusions.

Ambach and Denoth (ref. 8) measured the dielectric constant of wet snow as a function of water content. Their data points are shown on figure 8 along with our calculated curve for the case where n equals 3.5. (Note that ε_D is the dielectric constant at zero liquid content). The result suggests that the aspect ratio of 3.5 describes the average liquid inclusion for this case. The liquid fillets could be described better with a larger aspect ratio while the liquid veins could be described better by a smaller ratio. Since about two-thirds of the liquid is contained in the veins, the average aspect ratio is weighted in favor of the short veins.

Given the excellent correlation between theory and experiment shown on figures 5 and 8, it seems reasonable to use this theory with an aspect ratio of 3.5 for any case of high porosity ($\phi \geq 0.4$) and low liquid water content ($\theta \leq 0.10$). There is some uncertainty about the upper limit of the liquid content at which Case I transforms into Case II. Case I represents the pendular regime of liquid content where the air phase is continuous

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throughout the pore space. Case II represents the funicular regime where the liquid is continuous throughout the pore space. As the liquid content changes, granular porous media typically flip-flop from one state to the other as the transition saturation is reached. That is, there is a sudden and abrupt transition from the pendular to the funicular regime as the liquid content increases. For granular materials this transition typically occurs at about 7% liquid content but, since low density snow forms into grain clusters with large air-filled pores, we expect the transition to occur at a slightly higher liquid content. Thus we suggest that the liquid path becomes continuous and the air exists in isolated bubbles at a liquid content of about 10%.

CASE II: HIGH ϕ AND HIGH θ

Above the transition liquid content the grain clusters shown in figure 3 break down and the grains round off. The air exists in isolated bubbles trapped in the pores and the liquid water exists in continuous paths throughout the pore space. Thus we describe this case with the liquid being the continuum and the ice particles and air bubbles being spherical inclusions. Equation (3) then assumes a relatively simple form,

$$\varepsilon_1 = \varepsilon' + 3\varepsilon' (1-\phi)(\varepsilon_1-\varepsilon_i)/(2\varepsilon'+\varepsilon_i)$$
$$+ 3\theta_a\varepsilon'(\varepsilon_1-\varepsilon_a)/(2\varepsilon'+\varepsilon_a) \tag{9}$$

The calculated values of dielectric constant are shown on figure 9 for various porosities. While there is a strong effect of liquid content, there is relatively little effect due to porosity. Unfortunately there is no data against which to test this case.

CASE III: LOW ϕ

At low porosities (high ice content) we take ice as the continuum. At liquid contents in the funicular regime (above a transition liquid content of about 7 to 10%) the air occurs in isolated bubbles and the liquid occurs in the large pores at four grain contacts and in liquid-filled veins at three grain contacts. At low liquid contents and low porosities, the air can still be described as spherical inclusions while the liquid occurs in veins and fillets as described in Case I. Thus this case covers the entire range of liquid contents for porosities less than 0.4.

Equation (3) can now be expressed as

$$\varepsilon' - \varepsilon_i = \frac{1}{3}\,\theta\varepsilon'(\varepsilon_1-\varepsilon_i)\ \frac{\varepsilon'(m^2+7m+10) + (2m^2+5m+2)\varepsilon_1}{2\varepsilon'^2(m+1) + \varepsilon'\varepsilon_1(m^2+m+2)+\varepsilon_1^2 m} \tag{10}$$
$$+ 3\theta_a\varepsilon'(\varepsilon_a-\varepsilon_i)/(2\varepsilon'+\varepsilon_a)$$

where m is determined from the aspect ratio of the liquid inclusions as shown on figure 2. As in Case I, we take n as 3.5 to represent an average aspect ratio for the liquid veins and fillets. The calculated dielectric constant (minus the value for dry snow) is shown against liquid water content on figure 8. The data points of Sweeney and Colbeck (ref. 7) for low porosity ($0.27 < \phi < 0.39$) snow were corrected for a porosity of 0.32 and are shown along with Ambach and Denoth's data. The theories for high and low porosity (Cases I and III) describe the experimental data rather well (although there is considerable scatter in the data of Sweeny and Colbeck because no calorimeter measurements were made to check the liquid content).

APPLICATION TO A SNOW COVER

As liquid water passes the surface of a snow cover, liquid water moves downward as a wave which has been described (ref. 9) and observed (ref. 10) in many natural snow covers. During the early stages of melt, snow covers are usually composed of a sequence of layers with varying characteristics (ref. 11). One of the most important of these characteristics is the low porosity "ice layer" which tends to impound the infiltrating water and divert some flow down distinct flow paths (refs. 12 and 13). Both the ice layer and overlying soaked layer have distinctly different dielectric constants than the surrounding snow. Likewise, a layer of liquid soaked snow covers the ground surface when the moving water exceeds the infiltration capacity of the underlying soil. These soaked layers are an important feature of the snow cover since the dielectric constant of both low and high porosity snow are highly dependent on the liquid content (see figs. 8 and 9). For low porosity seasonal snow covers, the effect is particularly important since the dielectric constant of high porosity "slush layers" may be twenty times larger than that for dry snow of the same porosity.

The nature of an infiltrating wave of meltwater suggests two layers (ref. 14) which can be explained with well known principles of unsaturated infiltration (ref. 9). The simplest characterization of these two layers would attach a dielectric constant to each depending on its liquid water content. As the "wetting front" moves downward, the upper layer expands and the lower layer shrinks. If the size of the layers could be determined remotely then the rate of movement of the wetting front could be determined from successive measurements. This would allow a prediction of the time when the water would reach the ground and be available for runoff and/or infiltrating into the soil.

DISCUSSION AND CONCLUSION

The real part of the dielectric constant at high frequencies can be calculated for a variety of types of wet snow using the Polder and Van Santen model as modified here. The calculated and measured dielectric constants varied with porosity and liquid content in consistent manners for the dry snow and two types of wet snow for which experimental results were available. Since twice as much information could be obtained if the

imaginary part of the dielectric constant could be determined as well, it is important to examine this theory to see if it can be extended. Glen and Paren (ref. 15) suggest that formulae originally derived for the real part may also be valid for the imaginary part of the complex dielectric constant. If both could be determined, both ice and liquid volume fractions could be determined directly without resorting to another measurement (such as a separate density measurement). This would allow the maximum amount of information to be determined by measurements of the dielectric properties of wet snow.

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Figure 1.   Depolorization factors (A_i) are shown against the ratio of the c to a axes (n).

Figure 2.   Ratio of A_c to A_a (m) is shown against the ratio of the c to a axes (n).

Figure 3.   The unit cell of a grain cluster showing a liquid vein at the three grain junction and three liquid fillets at grain boundaries (from ref. 5).

Figure 4.   One grain from the cluster shown on figure 3. The grain is approximated by a spheroid with axes (0.706, 0.706, 0.5).

Figure 5.   Dielectric constant for dry snow shown against snow density and porosity. The line was calculated using equation (8) and the data points are from refs. 6 and 7. ε_a is 1, ε_i is 3.15, and ρ_i is 0.917 g/cc.

Figure 6a. The plan view of the liquid vein and fillets of the three grain cluster shown in figure 3.

Figure 6b. An artist view of the liquid vein and fillets.

Figure 7.   Calculated values of the dielectric constant shown against liquid water content for various values of the aspect ratio n. (For Case I with a porosity of 0.651).

Figure 8.   Measured values of dielectric constant (minus the dielectric constant for dry snow) are shown against liquid water content. The lines represent the values calculated from Case I (for the air continuum) and Case III (for the ice continuum). The values of Sweeny and Colbeck are corrected for a porosity of 0.32.

Figure 9.   Calculated values of dielectric constant shown against liquid water content for various values of porosity (from equation 9).
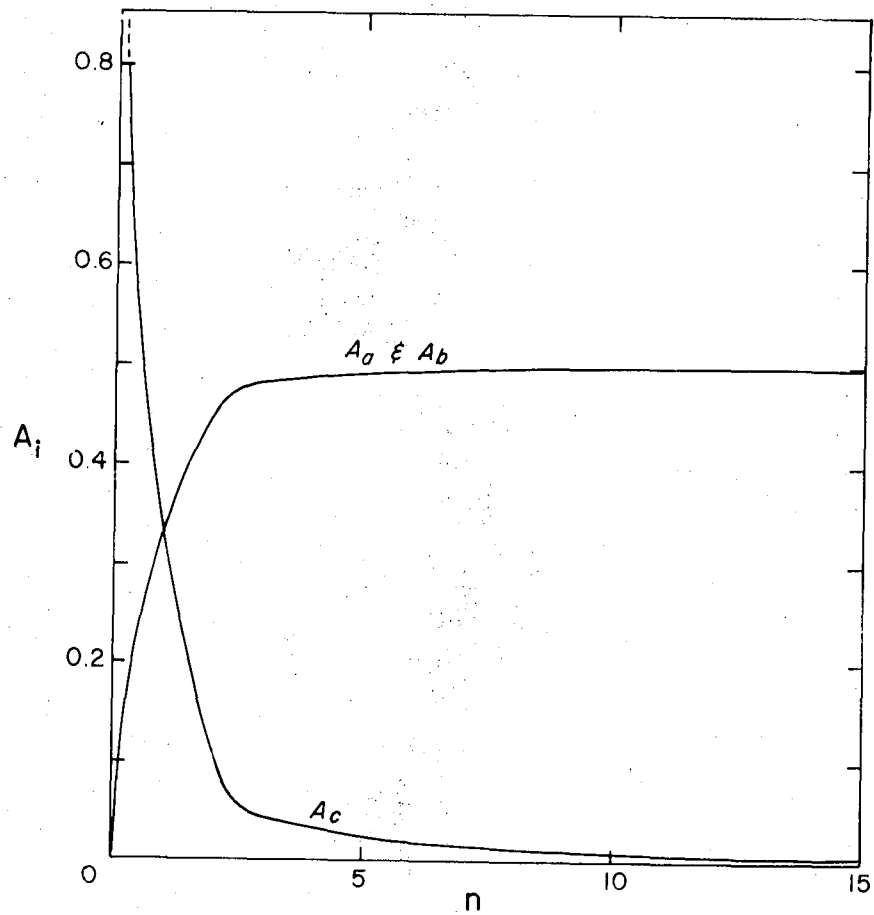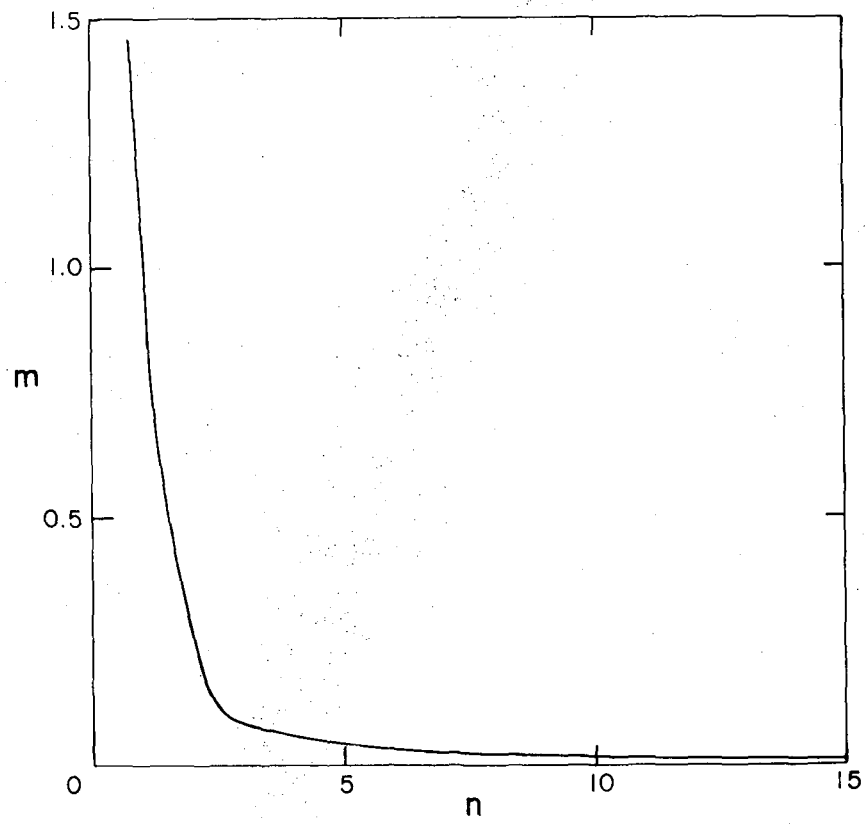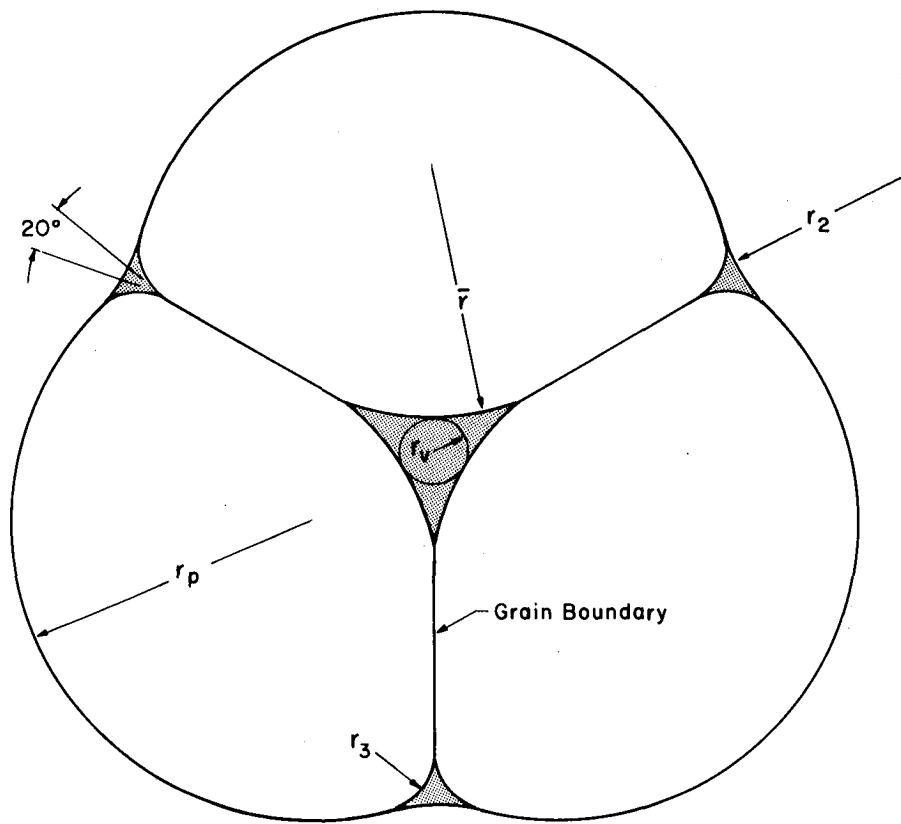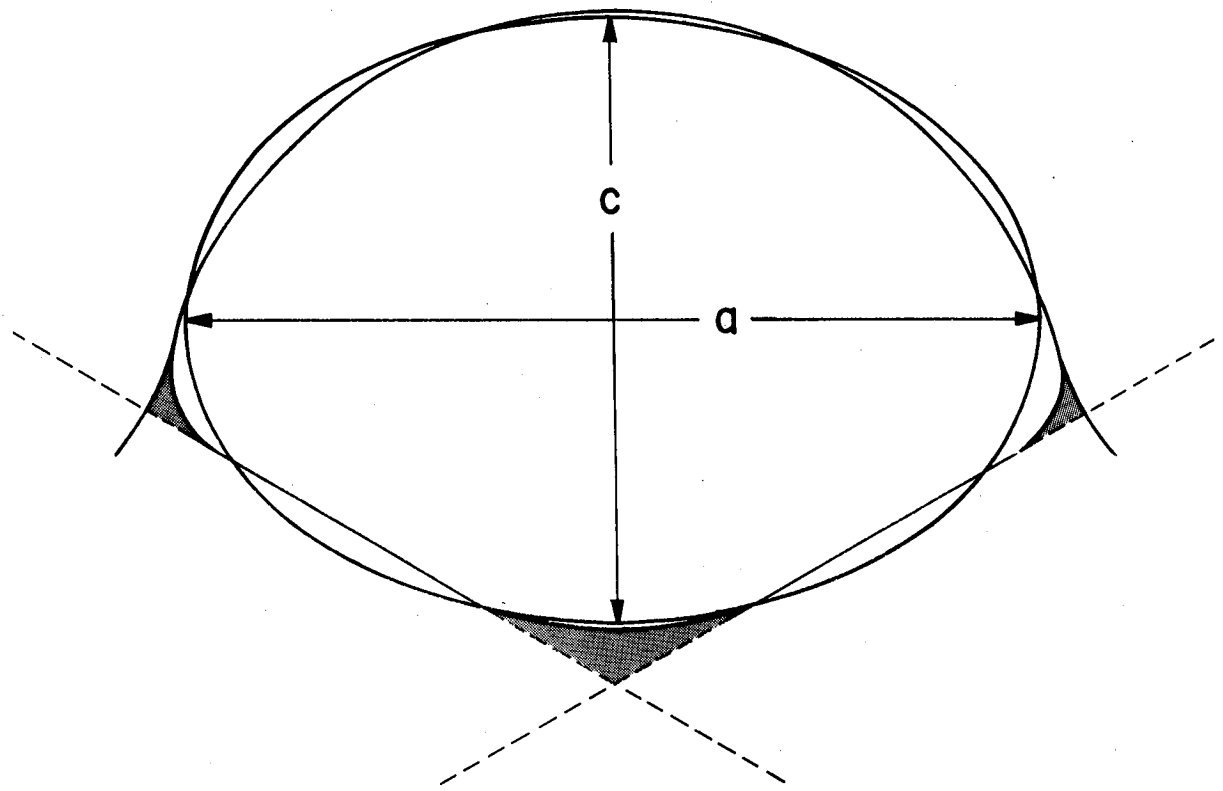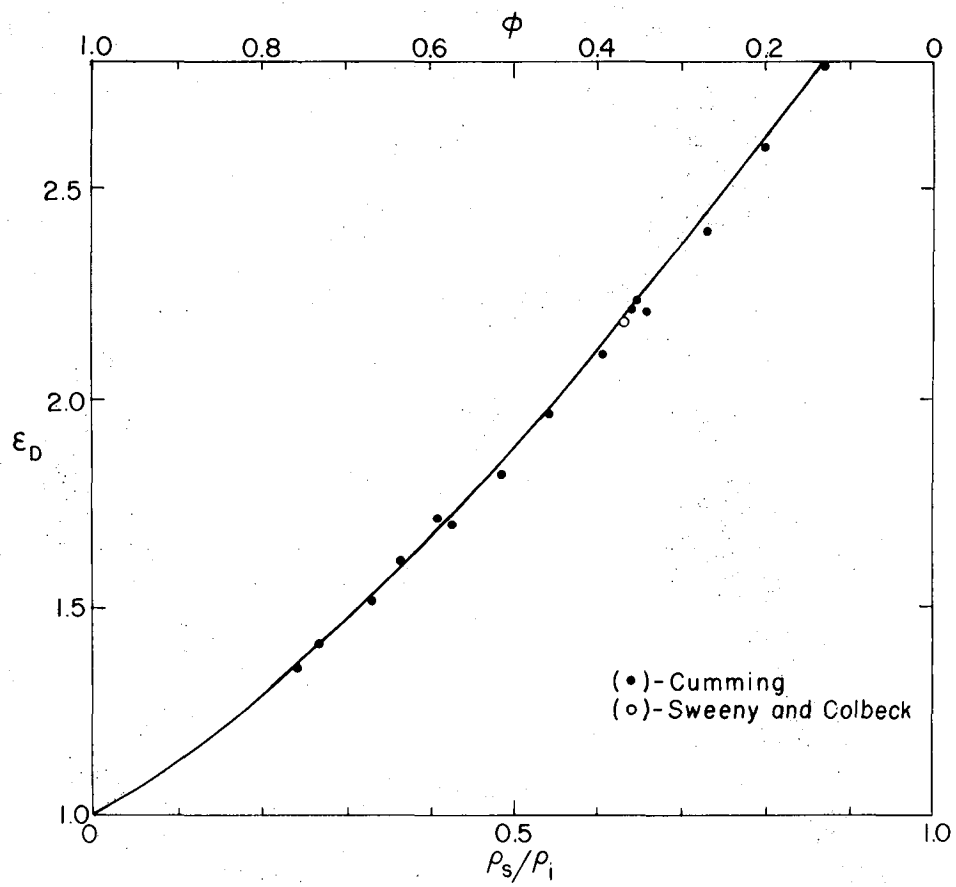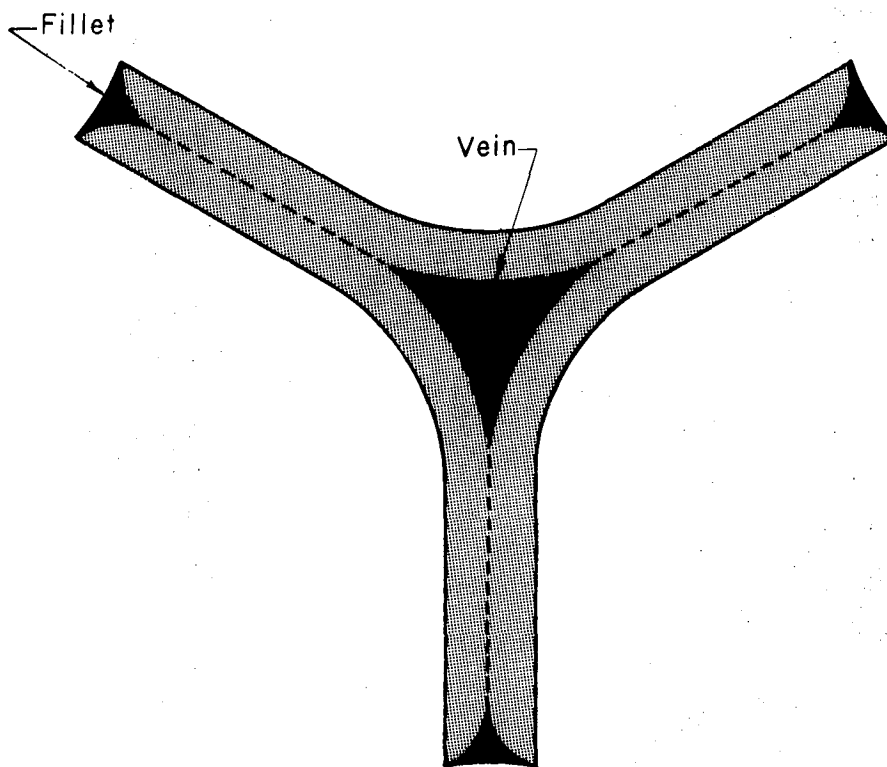
Figure 1

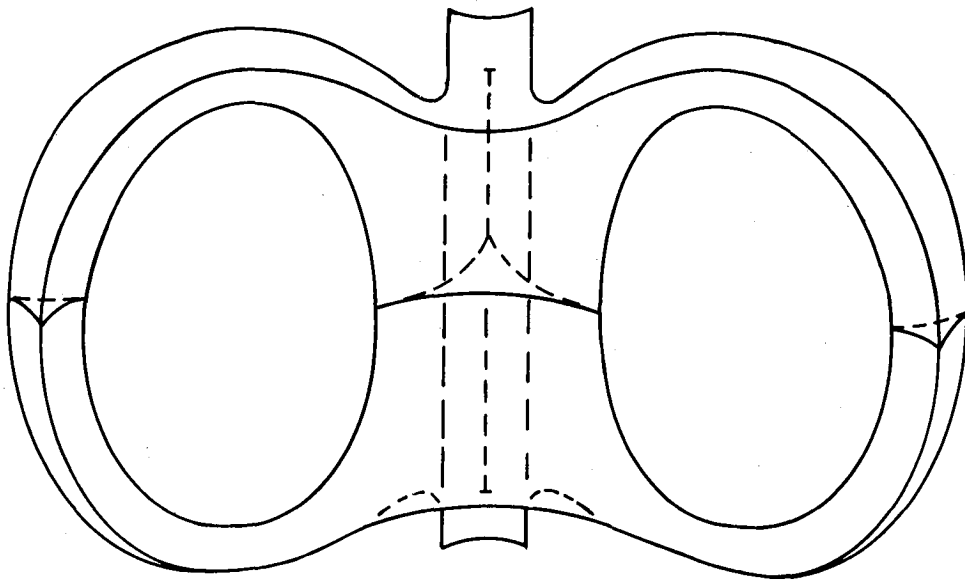Figure 2

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Figure 3

Figure 4
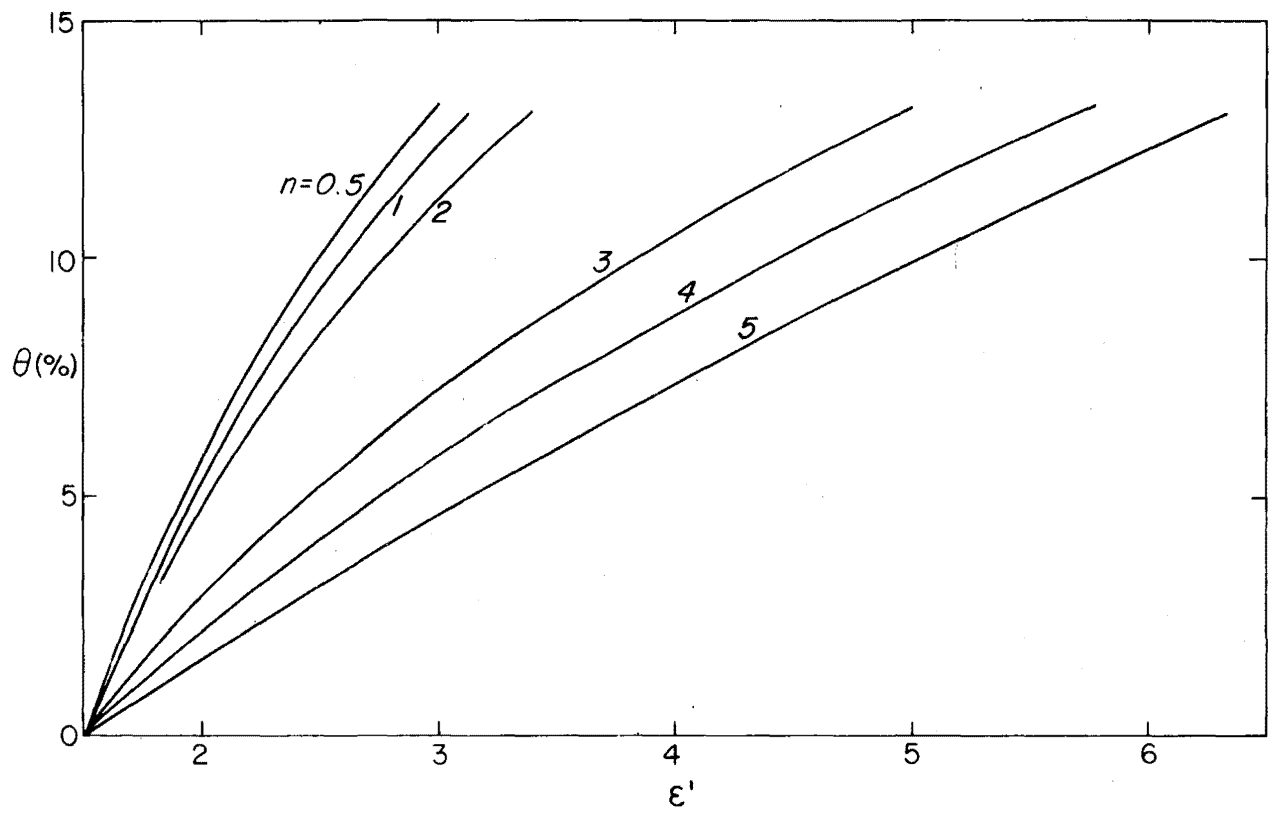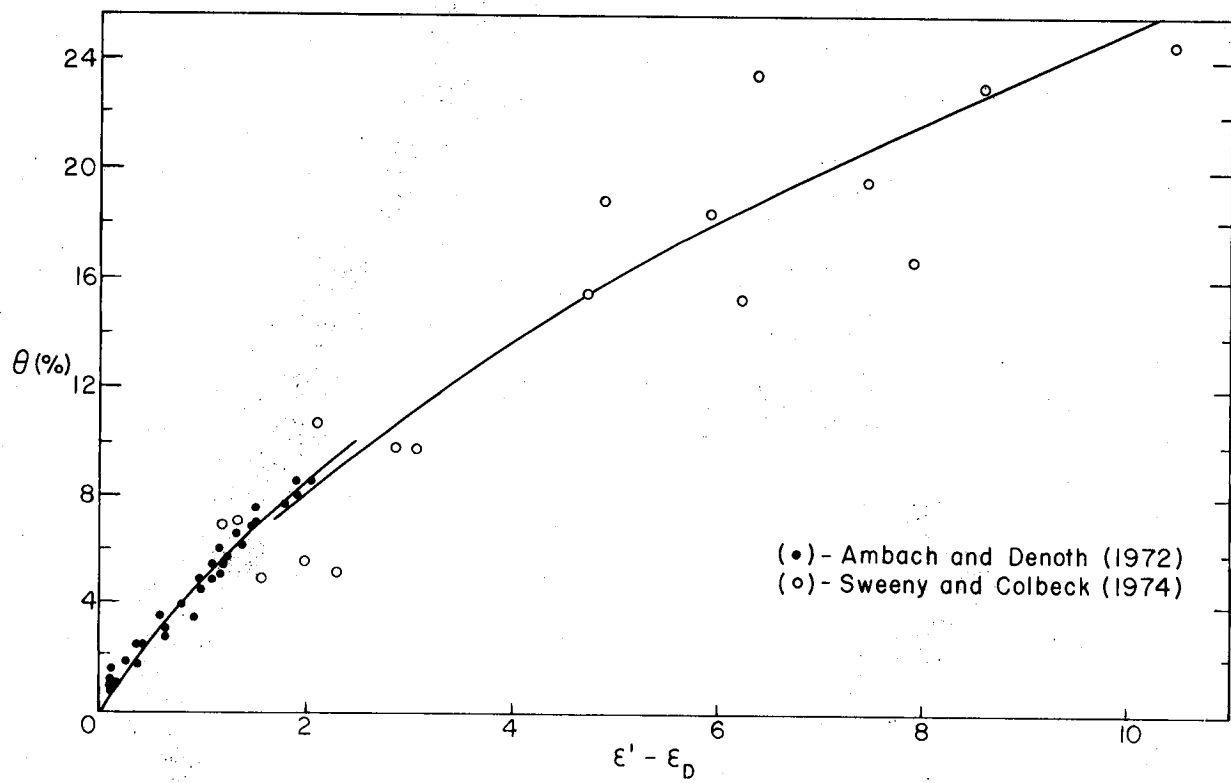
33

Figure 5

Figure 6a

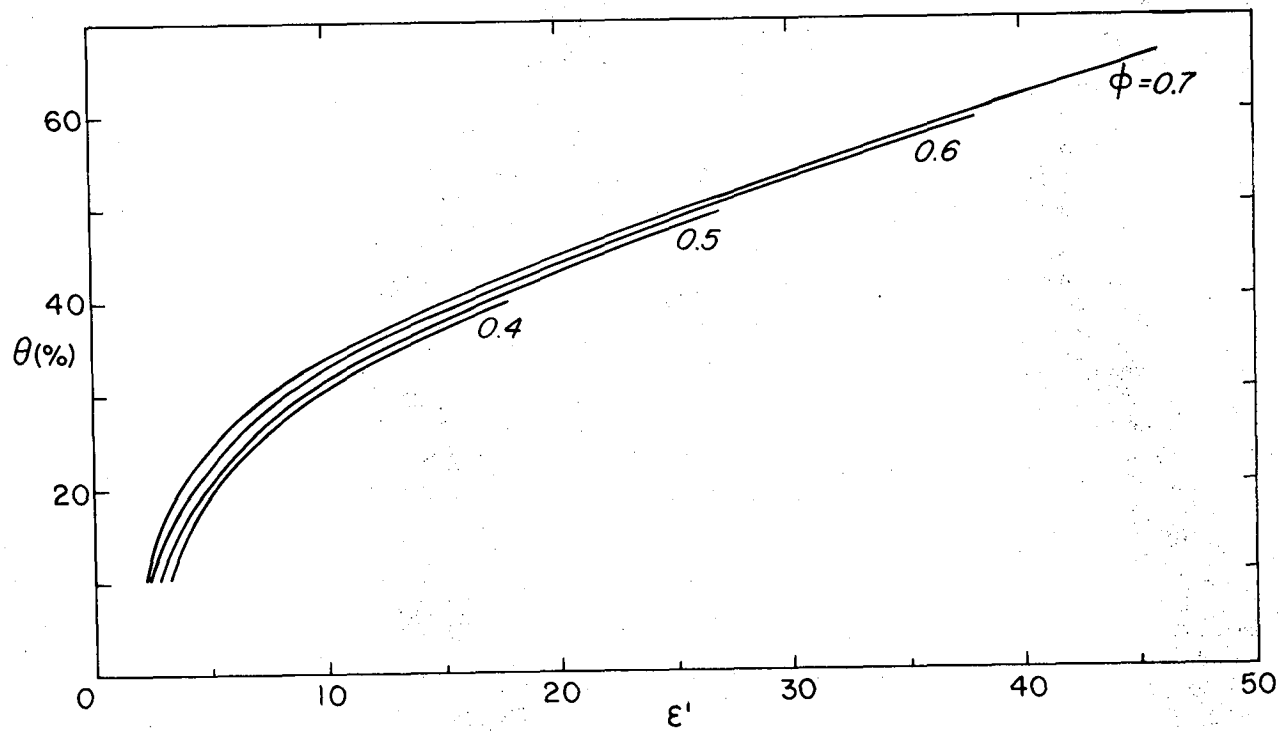Figure 6b

Figure 7

Figure 8

Figure 9

MEASUREMENT OF LIQUID WATER CONTENT IN A MELTING SNOWPACK USING COLD CALORIMETER TECHNIQUES*

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ABSTRACT

Liquid water in a snowpack has been recognized for some time as a quantifiable parameter of hydrological significance. It is also important in the interpretation of snowpack remote sensing data using microwave techniques. One acceptable approach to measuring liquid water content of a snowpack (by weight) is the cold calorimeter. This technique will be presented from theory through application. Silicon oil has been used successfully as the freezing agent. Consistent results can be obtained even when using new operators with a minimum of training. Data can be obtained approximately every 15 minutes by using two calorimeters and three operators. Accuracy within one to two percent can be achieved under reasonable field conditions.

INTRODUCTION

Liquid water in a snowpack has been recognized for some time as a quantifiable parameter of hydrological significance. However, this is not a regularly reported snowpack characteristic due to the difficulties of measuring liquid water in the field. The amount of liquid water is a parameter that is important in forecasting runoff, predicting the timing of wet avalanche release, and in the interpretation of snowpack remote sensing data using microwave techniques.

The liquid water content of a snowpack, sometimes called free water, snow water content, or liquid phase water, includes three categories of water: gravitational water moving downward through the snowpack, capillary water held by surface tension between the individual snow crystals, and hygroscopic or adsorbed water held in thin films on the individual snow crystals.

Various techniques have been used or proposed to measure the liquid water content of a snowpack. Most of these techniques can be broadly categorized as centrifugal, dielectric, and calorimetric and range from laboratory techniques to remote-sensing concepts. Some investigators have tested additonal methods including Shoda (ref. 1) who used a measurement of volume expansion upon freezing, and Bader (ref. 2) who used the concept of dilution of a solution by the liquid water in the snowpack.

Centrifugal separation of the liquid water from a snow sample was described by Kuroda and Furukawa (ref. 3) and Carroll (ref. 4). Langham (ref. 5, 6) subsequently improved the method by isolating the amount of melt that occurs during the centrifuging process. The centrifuge technique has been compared

with the calorimetric technique by the National Bureau of Standards (ref. 7). It was concluded that the centrifuge and freezing calorimeter method results do not measure the same phenomena and that water may be present in the snowpack in a form which is detectable by the freezing calorimeter method but not by the centrifuge method.

The measurement and comparison of the dielectric constant of wet and dry snow has provided a method of determining the liquid water content of the snow-pack (ref. 8). Linlor et al. (ref. 9), Linlor & Smith (ref. 10), and Linlor et al. (ref. 11) have used several methods for measuring the liquid water based on the dielectric constant, as well as experimenting with the attenuation of microwave beam transmission through a snowpack.

Several calorimetric methods have been proposed. The melting of a given quantity of snow with a measured amount of hot water and recording of the resulting temperature has been used (ref. 12, 13, 14). This method is diffi-cult to use in the field, however. A related procedure employing the melting of a snow sample by a measured amount of electrical energy was described by de Quervain (ref. 15) and Hansen and Jellinek (ref. 16). A third class of calorimetric techniques wherein the "negative heat" required to freeze a snow sample is measured was described by Radok et al. (ref. 17) and has been effec-tively used in the field by Leaf (ref. 18).

Howell et al. (ref. 19) and Bergman (ref. 20) have shown utilization of the freezing calorimetric technique with toluene as the freezing agent. Although toluene is a very satisfactory freezing agent, its toxic properties and rela-tively high flash point make its use less than desirable. This paper reports on several improvements to the freezing calorimeter technique in connection with investigations of snowpack properties using remote-sensing techniques. Alternative methods for calculating either snowpack-thermal quality or snow quality, both indicators of liquid water content, are presented for consider-ation of the user employing the freezing calorimeter approach.

CALORIMETRIC ANALYSIS

The freezing or cold calorimeter technique was used in the course of these liquid water determinations. This method was selected primarily because it is relatively inexpensive, it is based on known and documented physical phenomena, and it has been used as a method for calibrating other techniques[1]. In addi-tion, it appears that because a small amount of liquid water is frozen in the cold calorimetric approach as opposed to a large amount of snow melted in the hot water calorimetric method, the freezing calorimetric method is more sensi-tive to variations in liquid water content. Leaf (ref. 18) and Langham (ref. 5) estimated that the freezing calorimetric method has errors of about ±1% of the liquid water content by weight.

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[1]Leaf, Charles F., 1976. Personal Communication.

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A calorimetric analysis is based on the concept that the heat gained by the system must equal the heat lost. The cold calorimeter is no exception. The calorimeter and the freezing agent gain heat, the snow sample and the liquid water lose heat. These terms can be expressed as follows:

snow sample $\quad\quad\quad (W_3 - W_2)\ (t_2 - t_3)C_i$

liquid water $\quad\quad\quad (W_{fw})\,(L)$

calorimeter $\quad\quad\quad\quad E(t_2 - t_1)C_f$

freezing agent $\quad\quad\quad (W_2 - W_1)(t_2 - t_1)C_f$

Where W_{fw} = weight of the liquid water

$\quad\quad L$ = latent heat of fusion in cal/gm

$\quad\quad C_f$ = specific heat of freezing agent at $\dfrac{t_1 + t_2}{2}$ in cal/gm/$^\circ$C

$\quad\quad C_i$ = specific heat of ice at $\dfrac{t_2 + t_3}{2}$ in cal/gm/$^\circ$C

$\quad\quad E$ = calorimeter constant in gms

$\quad\quad W_1$ = tare weight of calorimeter in gms

$\quad\quad W_2$ = weight of calorimeter and freezing agent in gms

$\quad\quad W_3$ = weight of calorimeter, freezing agent and snow in gms

$\quad\quad t_1$ = initial temperature of the freezing agent in $^\circ$C

$\quad\quad t_2$ = final temperature of freezing agent and snow mix in $^\circ$C

$\quad\quad t_3$ = initial temperature of the snow at sample point

If one then sets the heat gained terms equal to the heat lost terms the following equation is obtained:

$$(W_3 - W_2)(t_2 - t_3)C_i + W_{fw}L = [(W_2 - W_1) + E](t_2 - t_1)C_f \quad\quad (1)$$

Let P be the percentage of liquid water in the sample so that:

$$W_{fw} = P(W_3 - W_2)$$

Then substituting in equation (1) the following results are obtained:

$$P = \left[\frac{[(W_2 - W_1) + E](t_2 - t_1)C_f}{L(W_3 - W_2)} - \frac{C_i(t_2 - t_3)}{L}\right] \quad\quad (2)$$

Snow quality, Q_f, is equal to $1 - P$ and is defined by Bruce and Clark (ref. 21) as the percent (by weight) of water in the solid state. Thus if the snowpack was totally dry, i.e., no free water, it would have a snow quality of 100%. Snow quality can be computed by using the following equation:

$$Q_f = 1-P$$

$$Q_f = 1 - \left[\frac{[(W_2 - W_1) + E](t_2 - t_1)C_f}{L(W_3 - W_2)} - \frac{C_i(t_2 - t_3)}{L} \right] \tag{3}$$

During development of the method to calculate snow quality, the t_3 parameter was added to the original formulations presented by Radok et al. (ref. 17) and Leaf (ref. 18) to improve physical understanding and to empirically limit Q_f from exceeding 100%.

Another concept closely related to snow quality is thermal quality. The thermal quality of a snowpack (Q_t) is the ratio of heat necessary to produce a quantity of melt from an existing snowpack to the heat necessary to produce the same quantity of melt from pure ice at 0°C and is usually expressed as a percentage (ref. 22). The snowpacks colder than 0°C would have thermal quality values greater than 100%, and 0°C snowpacks holding liquid water would have thermal qualities less than 100%. The equation used to compute thermal quality of a snowpack is as follows:

$$Q_t = 1 - \left[\frac{[(W_2 - W_1) + E](t_2 - t_1)C_f}{L(W_3 - W_2)} - \frac{C_i t_2}{L} \right] \tag{4}$$

The equation is identical to equation (3) except the t_3 term is deleted, thus making values greater than 100% possible. It should be noted that this also occurs when $t_3 = 0$ C in equation (3). Thus, it is possible to have snow quality and thermal quality be numerically equal when the temperature of the ice crystals in the snow equals 0 C.

FREEZING CALORIMETER EQUIPMENT AND PROCEDURES

FIELD EQUIPMENT

Listed below is the basic equipment required for utilizing the cold calorimetric technique in the field

- Calorimeter. For this experiment, the calorimeter was constructed from a wide mouth vacuum bottle made of stainless steel.

- Scales. The scales should be capable of weighing up to 2,000 grams and should be readable to the nearest tenth of a gram.

- Temperature Probes. Various types are available and can be used, but they should be easily read and of a design that allows for field use. In this experiment, 12-inch stainless steel thermocouple probes were used. A multi-channel digital readout was employed so that air and snowpack temperatures could be simultaneously read.

In addition to these basic requirements, several other items of equipment are necessary:

- cold chest for storing the freezing agent;

- thermal containers for obtaining the snow samples;

- timing devices;

- miscellaneous tools such as, shovels, trowels, spoons, etc.

Figure 1 shows the calorimeter with the temperature probe. Figure 2 shows the typical equipment used for field operations in the back of a four-wheel-drive support vehicle. Figure 3 shows the field facility in operation at the 1979 test site near Fraser, Colorado.

FREEZING AGENTS

Cold calorimetry requires the use of a freezing agent to freeze the liquid water in the snow sample. It is recognized that some of the hygroscopic water will not be frozen, but the resulting error is considered to be very small.

Ideally, a freezing agent should have low viscosity (down to about -60°C), high flash point, be easily obtainable, and be non-toxic. Also, it is desirable that the freezing agent leave no residue or film in the calorimeters. The initial freezing agent used was a chemical grade toluene and subsequently a good industrial grade toluene was found to work equally well. Toluene, although working well as a freezing agent, has two major drawbacks. First, it is somewhat toxic, and second, it has a flash point low enough to create a potential fire and explosion hazard. This means that considerable care has to be exercised in its storage and use.

Because of these safety problems, a very light silicon oil was tested for use as a freezing agent. The specific substance used was General Electric SF-96-5.[2] This product is more expensive and requires more careful cleaning of equipment; but, because of the safety gained, it is considered superior to the toluene.

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[2]The use of a trade name in no way implies endorsement. The specific name is given so that the reader may investigate its properties.

Part of the additional expense of the silicon can be offset by the fact that much of it is reclaimable by first letting the ice silicon mixture warm so that all the ice is melted. The water can then be removed and the silicon filtered. Only a small portion of the silicon is lost.

CALORIMETER CONSTANT DETERMINATION

The determination of snow liquid water content through the use of the freezing calorimetry involves a heat-balance relationship occurring between a freezing agent and the liquid water contained in a snow sample as they are mixed together in a closed container. The type of container which is primarily used for the mixing process is a vacuum insulated bottle with a temperature probe and a tightly fitted rubber stopper. Experience indicates that a commercial, 1-quart (0.946 liter), stainless steel, wide-mouth vacuum bottle works quite well. The temperature probe is used to monitor the changes in temperature that occur in the vacuum bottle during the mixing process. The heat-balance equation is dependent upon accurate measurements of temperature changes occurring inside the bottle during a typical mix. Knowing that the vacuum insulated bottle is not a perfect system and some heat will be gained by the bottle itself, the heat-balance equations (Equations 1 and 2) contain a calorimeter constant, E. Each individual calorimeter bottle will have its own constant and must be determined independently. For convenience of use in the heat-balance equations, the calorimeter constant is expressed in terms of equivalent weight of freezing agent. Various methods may be used to determine this constant. This discussion considers only the one developed in the course of this study.

References are made throughout this discussion to one typical calorimeter constant determination shown in figure 4. This determination was actually only one in a series of twelve used to obtain an adequate sample from which to compute the mean value for a specific calorimeter bottle. The freezing agent used was a silicone fluid. The calibration of the calorimeter bottle was done in the laboratory using the same equipment that is used in actual field operations. Quantities and time frames were approximated to match those used in field procedures. Temperatures were varied from run to run to simulate changing field conditions.

Theory

The calorimeter constant is determined by a basic heat-balance equation. When a warm fluid is mixed with a cold fluid in a calorimeter bottle, the heat which is lost by the warm fluid must be equal to the heat gained by the cold fluid and the bottle itself. The heat-balance equation is:

$$\underbrace{\left[Wt_{warm}\right]\left[\frac{Cs_w + Cs_2}{2}\right]\left[T_w - T_2\right]}_{\text{Heat lost by warm fluid}} = \underbrace{\left[Wt_{cold} + E\right]\left[\frac{Cs_1 + Cs_2}{2}\right]\left[T_2 - T_1\right]}_{\text{Heat gained by cold fluid + bottle}}$$

where:

Wt_{warm} = Weight of warm fluid.

Wt_{cold} = Weight of cold fluid.

E = Calorimeter constant expressed in equivalent grams of fluid.

T_w = Initial temperature of warm fluid before mixing.

T_1 = Initial temperature of cold fluid before mixing.

T_2 = Final temperature of warm-cold fluid mix.

Cs_w = Specific heat of fluid at temperature of warm fluid.

Cs_1 = Specific heat of fluid at temperature of cold fluid.

Cs_2 = Specific heat of fluid at final temperature of mix.

The fluid weights are directly obtainable and can be determined with reasonably high accuracy. Also, the specific heats of the fluid can be obtained directly from charts after the corresponding temperatures have been determined. Table 1 shows the specific heat values for the silicon oil used. The determination of the initial and final temperatures of the fluid is the most critical and time-consuming part of the process. These determinations are also the greatest potential source of errors in the system. The temperature values are determined by standard calorimeter techniques which involve the extrapolation of tempera-ture curves.[3] The cold fluid (approximately 1 pint) is poured into the calorim-eter bottle (a 1-quart (0.946 liter) size vacuum insulated bottle) at time zero. The initial warm up of the fluid and cooling of the bottle occurs in the first 3 minutes, and then the slope of the temperature curve becomes fairly uniform indicating the bottle itself has cooled and is now gaining heat through the walls from the outside at an almost constant rate. The temperature of the fluid was monitored for 8 minutes in order to get a good definition of the slope of the curve. In the period between 8 and 9 minutes the warm fluid (approximately 1/2 pint), with a measured temperature T_w, is added and mixed thoroughly. Temperature recordings are started at the 9-minute mark and are recorded until the slope of the second curve stabilizes and becomes relatively constant (about 5 minutes). Note that the bottle should be shaken lightly throughout this whole period to insure uniform temperatures throughout the interior of the bottle and the fluid. The curves are plotted as shown in figure 4 and both are extrapolated to the mid-point in time (8.5 min.) when the transfer of heat during the mix is assumed to occur. The initial (T_1) and final (T_2) temperatures can be picked off the curves with reasonable accuracy. The calorimeter constant can then be calculated as shown at the top of figure 4.

Results

When the calibration of a calorimeter bottle is performed under laboratory conditions, the accuracy of the measuring equipment and the technique and experience of the calibrator will determine the precision and repeatability of the results. Accurate scales and temperature measurements can eliminate errors

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[3] Corrin, Myron L., 1978, Personnel Communication.

in the actual data taking. However, plotting and interpretation of the temperature data is a source of error which is difficult to eliminate. Thus, in practice the authors have used the average of ten to twelve calibrations to determine the calorimeter constant. The heat-balance equation is quite sensitive to the two temperature factors. Therefore, care should be given to the plotting of the temperature curves and the extrapolation of the slopes to the required points. Actual laboratory procedures using a beam scale accurate to 0.1 gram and an electronic digital thermometer accurate to 0.1 degree Centigrade have produced sets of calorimeter constants of 12 values that are within 10 percent of the mean value. This range of values is within the limits of the system as a 10 percent change in the value of the calorimeter constant will result in less than 0.5 percent change in the snow-quality factor equation developed by Radok (ref. 17) and Leaf (ref. 18).

FIELD PROCEDURE

The freezing agent was stored in 1-pint glass bottles in an insulated ice box containing dry ice. The temperature of the freezing agent was maintained in the range of $-40^{\circ}C$ to $-50^{\circ}C$. It is important to maintain this range of temperature particularly during periods of high air temperatures and high liquid water contents in the snowpack so that the resultant temperature, t_2, will be less than $0^{\circ}C$. The calorimeter was pre-cooled to the average internal operating temperature prior to taking the first set of data. During the actual measurements, the calorimeter will remain cooled due to the use of the freezing agent. After the calorimeter bottle has been cooled for approximately 20 minutes (the average time per run) using 1-2 pints of freezing agent, the bottle is emptied, wiped dry, and a tare weight (W_1) is taken. One pint of freezing agent is then poured into the bottle and the cap is sealed on. At this point the total weight of the bottle and freezing agent (W_2) is obtained and recorded. From this point on, the procedure used in the current set of measurements (1979) varies from that used in earlier experiments (1976-1978). Initially, the bottle was shaken and the temperature of the freezing agent was checked until a stable reading was obtained (t_1). The use of a more precise digital thermometer in the later measurements showed that a "stable" temperature is actually never achieved until the temperature of the freezing agent equals that of the air around the bottle. Investigation of standard calorimetric techniques[4] led to obtaining an initial temperature (t_1) for the freezing agent through extrapolation of a temperature curve in the same manner as was done in the calorimeter constant determination (see figure 4). The temperature of the freezing agent is monitored and recorded every 30 seconds while the bottle is shaken. After the first 3-5 minutes of shaking, the slope of the temperature curve becomes fairly uniform. The temperature monitoring is continued for 8 minutes in order to get a good definition of the slope of the curve. A pre-cooled thermos bottle is used to collect and store the snow sample and return it to the field laboratory site. The snow (approximately 200-225 grams) is then added to the cold freezing agent in the calorimeter and the cap is again sealed

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[4]Corrin, Myron L., 1978, Personnel Communication.

on. The bottle is then shaken to thoroughly mix the contents. The 1976 procedure called for shaking the bottle until a "stable" temperature (t_2) is reached (in 3-6 minutes). The 1979 procedure calls for monitoring the temperatures again in 30 second intervals until the temperature curve becomes well defined (about 4-5 minutes). The total weight of the bottle, freezing agent, and snow (W_3) is measured and recorded to complete one run of the routine. The 1979 procedure is dependent on a good record of time being kept throughout the entire process. The initial temperature curve is obtained from observations taken every 30 seconds during the 0-8 minute interval. The snow is added between minutes 8-10. Mixing takes place during the interval from minutes 10-14, and a second temperature curve is obtained from every 30 second observation made during this interval. Both temperature curves are extrapolated to the 9 minute point when the transfer of heat during the mix is assumed to occur. At this point the temperatures t_1 and t_2 are determined from the two curves as shown in figure 4. The field procedures used in the 1976 and 1979 measurements are similar in time frames and general methods. The field procedure has been refined in the 1979 measurements through the use of more precise measuring equipment and basic refinement of the calorimeter technique itself.

Operating the cold calorimeter apparatus under winter field conditions leads to difficulties in weighing samples and measuring temperatures. Scales and temperature systems are checked for accuracy prior to field operations to minimize basic calibration errors. Field operations are normally conducted with some type of simple shelter to minimize wind and precipitation problems. Despite shelters and precautions, minimal errors are still present.

Figure 5 shows the form on which data are recorded for the liquid water analysis using the cold calorimetric technique. The example form illustrates typical values one might expect in working with a moderately wet snow. Figure 6 shows the graphical determination of t_2 and t_1.

FIELD APPLICATIONS OF COLD CALORIMETRIC TECHNIQUE

Two sites in Colorado were used in 1979 for free water determinations (ref. 23). They were located in the southern portion of the Fraser Valley and just south of Steamboat Springs in the Yampa Valley. These locations were used as part of a microwave remote sensing field experiment program. The purposes of these calorimeter tests were to determine:

● Typical diurnal patterns of free water;

● Repeatability of results using various calorimeter operators;

● Possibilities for improvements in field operation techniques.

FRASER VALLEY SITE

The Fraser Valley site is located in an open meadow just west of Berthoud
Pass off U.S. Highway 40 between Hideaway Park and Fraser. This is the same
site that has been used by NASA and the National Bureau of Standards (NBS) for
their microwave snowpack remote-sensing experiments (ref. 24). The dates
selected for this work (March 13, 14, and 15, 1979) at the Fraser Valley site
were chosen so that they would coincide with the ongoing microwave experiment
conducted by NASA and NBS.

The weather at the site was excellent for the work. The morning of
March 13 was cloudy and some light snow was still falling. However, by mid-
morning it began to clear and was almost totally clear by noon. During the
morning of the first day, a snow pit was prepared and the snow properties
characterized as shown in table 2. Data collection on the snow liquid water
using the cold calorimeter began shortly after noon. The weather held for the
remainder of the 3 days, and with the exception of some overcast skies on the
second day, conditions were mostly clear. Data for the computation of the
percentage of liquid water in the snow and the results are shown in table 3,
and plotted in figure 7. Air temperatures shown in figure 7 are those taken
in a shaded condition.

STEAMBOAT SPRINGS SITE

Tests were conducted at the Steamboat Springs site on March 28, 29, and
30, 1979. This timing was selected so that an isothermal snowpack would be
available, but it would still have adequate depth so as to minimize the ground
effects. The location of the Steamboat Springs site was on the previously
established 5-mile Steamboat Springs flight line used by NASA in connection
with their airborne microwave remote-sensing research.

Considerable overcast conditions, as well as numerous brief snow showers,
were encountered. Due to the overcast skies, low nighttime temperatures
approached the freezing level, but it was not cold enough to completely freeze
the snowpack. Thus, the measurements were taken on a snowpack that presented
different conditions from those observed at Fraser. The first task at this
site, as at Fraser Valley, was to dig a snow pit and characterize the profile,
as shown in table 4.

The results of the first 2 days of field operations are presented in
table 5. A review of these data shows considerable variation in the liquid
water on the first day, but much less variation on the second day. This was
attributed to the overcast conditions that reduced the nighttime back-
radiation and the incoming daytime radiation. Since the weather for the third
day (March 30, 1979) was forecast to be a repeat of that of the previous day,
it was decided to take advantage of this relatively stable weather and run
liquid water content determinations of the various layers identified in the
snow pit. These data are shown in table 6. For each liquid water determina-
tion the snow pit was extended horizontally so as to expose a new face for
sampling.

DISCUSSION OF DATA

Perhaps the most striking difference between the data taken at the two sites is the relative flatness of the diurnal liquid water curves at Steamboat Springs as compared to those at Fraser Valley. This is attributed to the relatively clear skies at Fraser Valley.

The data from both sites tend to present a sawtooth pattern rather than a smooth curve. Some of this may be attributed to changes in incoming radiation and random errors inherent in the system, but it is hypothesized that most of it is representative of the manner in which the upper layers of the snowpack melt. Water may build up on a given ice crystal or series of ice crystals due to the melting of the crystals or other crystals. At some point in this melt process, the mass of water held by capillary forces becomes large enough for the gravitational forces to overcome the tension or capillarity. When this occurs, the water is released downward into the snowpack. The best examples of the sawtooth curve tend to appear near the time of the greatest amount of incoming solar radiation (also a time of high melt potential). This phenomena needs further documentation, for if this hypothesis is correct, the snowpack viewed on a micro-scale will have considerable differences in liquid water content over short periods of time and short distances. For remote sensing purposes, however, on a meso- or macro-scale these variations, because of their statistical nature of occurrence, may present a relatively smooth curve.

This study was also designed to test operators in their ability to learn the technique and produce meaningful results. During this study, it was found that different operators could be field-trained within a few hours. Figure 7 shows the results by operator, and no consistent differences were found between operators after training.

Finally, the study was to be a test of field techniques. Wind continued to be a problem. The scales had to be sheltered. In deeper snow, a snow pit can serve as shelter, or in this case, where a road was nearby one can work in the protection of a vehicle. In running the tests it was found that three people are definitely needed to run two calorimeters. Since each calorimetric determination takes approximately 20 to 25 minutes, a dual calorimeter system is fairly dependable for readings every 15 minutes. Operators must be dedicated to obtaining good data, since the continuous running of liquid water determinations can become a rather tedious task.

All the field equipment used appeared to be well suited to field applications. When working at one location, recording equipment would be desirable for documentation of air and snowpack temperatures and solar radiation at 1- or 2-minute intervals.

It is difficult to precisely estimate the errors in this technique. However, it currently appears to be in the range of 1 to 2 percent for determinations made in the field. This figure was arrived at by running liquid water contents on extremely cold snow in which one would expect the liquid water content to approach zero.

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CONCLUSIONS

Formulations were presented for the calculation of the liquid water content of snow as a percent by weight as well for the determination of the associated parameters, snow quality and thermal quality. These formulations were based on the use of methodology associated with freezing or cold calorimetry.

The equipment and procedures currently being utilized in freezing calorimeter determinations of snow liquid water content are relatively simple and easy to use. The substitution of a light silicon oil for toluene as the freezing agent has markedly improved the safety factor as well as reduced the total amount of freezing agent required. The use of more precise digital thermometers in the field operation has allowed more rapid and accurate determinations of solution temperatures leading to a basic refinement of the calorimeter technique.

Differences in amplitude of the diurnal liquid water curves at the two test sites were attributed to the presence or absence of clouds. Clear skies allowed significant incoming solar radiation and a sharp rise up to as much as 29.5% liquid water content (by weight) from an initial value of 0.0%. Radiation cooling under cloudless skies at night allowed the liquid water to go to zero as a result of freezing. Generally, much less diurnal variation is experienced under overcast conditions. Under both situations, a sawtooth pattern rather than a smooth curve was apparent for the diurnal liquid water variation. This was attributed to both variations in incoming radiation and to an apparent periodic release of meltwater accumulated in the upper layers of the snowpack.

Training of new operators is accomplished easily, and no consistent differences in liquid water determinations were found between operators. Each calorimetric determination takes about 25 minutes so that with continuous operation of two calorimeters, liquid water readings can be obtained about every 15 minutes. To accomplish this, three operators should be employed. This frequency of measurement is appropriate for coordinated remote sensing observations during the snowmelt season.

The errors associated with this cold calorimeter technique when used in the field appear to be on the order of plus or minus one to two percent (employing the equipment described).

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TABLE 1.- SPECIFIC HEAT OF SILICONE FLUID
(General Electric SF - 96 -5)

t °C	Cs Cal/g/°C	t °C	Cs Cal/g/°C	t °C	Cs Cal/g/°C
25	.4440	-1	.4392	-27	.4343
24	.4439	-2	.4390	-28	.4341
23	.4437	-3	.4388	-29	.4339
22	.4435	-4	.4386	-30	.4337
21	.4433	-5	.4384	-31	.4335
20	.4431	-6	.4382	-32	.4333
19	.4429	-7	.4380	-33	.4331
18	.4427	-8	.4378	-34	.4329
17	.4425	-9	.4377	-35	.4328
16	.4424	-10	.4375	-36	.4326
15	.4422	-11	.4373	-37	.4324
14	.4420	-12	.4371	-38	.4322
13	.4418	-13	.4369	-39	.4320
12	.4416	-14	.4367	-40	.4318
11	.4414	-15	.4365	-41	.4316
10	.4412	-16	.4363	-42	.4314
9	.4410	-17	.4361	-43	.4313
8	.4409	-18	.4360	-44	.4311
7	.4407	-19	.4358	-45	.4309
6	.4405	-20	.4356	-46	.4307
5	.4403	-21	.4354	-47	.4305
4	.4401	-22	.4352	-48	.4303
3	.4399	-23	.4350	-49	.4301
2	.4397	-24	.4348	-50	.4299
1	.4395	-25	.4346		
0	.4393	-26	.4345		

TABLE 2.—SNOWPACK PROFILE CHARACTERIZATION AT THE FRASER VALLEY SITE, MARCH 13-15, 1979

PIT Nº __1__   LOCATION: __Fraser Valley Site_____   3/13 +0.7°C
                                                                  3/14 +6.4°C
DATE: _3/13/79_   TIME: _____   TOTAL DEPTH: _92 cm_   AIR TEMPERATURE: 3/15 -3.2°C TABLE Nº ____

DEPTH TO BOTTOM OF LAYER	THICKNESS OF LAYER	SNOW CLASSIFICATION	GRAIN SIZE	LAYER TEMP. (°C)			DENSITY (kg/m³)		REMARKS
				3/13	3/14	3/15	3/14		
1½ cm	.1½ cm	melt-freeze crust		+1.2	0.0	-0.6			
3½ cm	2 cm	soft-new snow	<1 mm ave.	+0.1			180		irregular, sharp crystals
4½ cm	1 cm	frozen layer (not ice lens)							
14½ cm	10 cm	melt-freeze metamorphosed	1 mm ave.	-1.1	-0.5	-4.8	200		loose, granular crystals rounded
15 cm	½ cm	ice lens							
19 cm	4 cm	melt-freeze metamorphosed	1 mm ave.	-3.8	0.0	-4.4	280		
19½ cm	½ cm	ice lens							
23½ cm	4 cm	melt-freeze metamorphosed	1 mm ave.	-3.7	-4.0	-3.5			
66½ cm	43 cm	melt-freeze metamorphosed firnification	1 mm+ ave.	-3.0	-3.1	-2.4	320		flat platelets
92 cm	25½ cm	depth hoar	2 mm ave.	-1.3	-2.5	-0.6	250		bottom 4 cm well developed
Grd				+0.4	0.0	+0.4			

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TABLE 3.—COLD CALORIMETER DATA FOR LIQUID WATER CONTENT DETERMINATIONS ON MARCH 13-15, 1979 AT THE FRASER VALLEY SITE

Date	Time	Air temp. (°C)	E (g)	W_1 (g)	W_2 (g)	W_3 (g)	T_1 (°C)	T_2 (°C)	T_3 (°C)	Q_f	Liquid water (%/100)	Oper.	Remarks
3/13/79	1337	2.2	92.2	1260.0	1615.3	1775.4	-21.8	-14.6	0	0.9798	.0202	SH	
	1348		78.7	1254.6	1625.1	1809.5	-25.8	-16.3	0	0.9736	.0264	SO	
	1403		92.2	1261.2	1628.1	1769.3	-23.7	-16.5	0	0.9733	.0267	SH	
	1421		78.7	1249.0	1601.0	1786.0	-35.3	-22.6	0	0.9761	.0239	SO	
	1436		92.2	1260.8	1610.0	1749.2	-21.6	-15.9	0	0.9987	.0013	SH	
	1446		78.7	1258.4	1630.7	1779.2	-42.2	-29.2	0	0.9607	.0393	SO	
	1458	2.7	92.2	1260.9	1644.2	1755.7	-34.6	-26.5	0	0.9717	.0283	SH	
	1512	2.7	78.7	1258.1	1619.6	1759.6	-46.1	-32.3	0	0.9573	.0427	SO	
	1524	2.4	92.2	1260.7	1629.0	1761.2	-31.2	-22.0	0	0.9590	.0410	SH	
3/14/79	0821	-10.0	92.2	1260.0	1631.4	1753.3	-31.2	-27.4	-12.5	1.0088		SH	
	0845	-8.8	92.2	1260.3	1644.4	1768.4	-36.1	-30.6	-8.8	1.0133		SH	
	0915	-6.9	92.2	1260.6	1657.0	1795.5	-36.7	-29.9	-5.3	1.0134		SH	
	0939		92.2	1260.3	1667.8	1793.2	-37.8	-30.4	-2.8	1.0039		SH	
	1007		78.7	1254.4	1600.4	1751.8	-39.0	-29.2	-3.0	1.0067		SO	
	1016		92.2	1260.2	1602.9	1764.1	-34.7	-25.3	-2.9	0.9964	.0036	SH	
	1030		78.7	1259.3	1670.8	1858.6	-51.0	-37.2	-2.8	1.0071		SO	
	1042		92.2	1260.1	1659.6	1820.0	-43.4	-31.6	0	0.9923	.0077	SH	
	1053		78.7	1257.1	1652.5	1844.0	-55.4	-38.0	0	0.9912	.0088	SO	
	1105		92.2	1260.9	1645.7	1813.8	-44.4	-31.7	0	0.9936	.0064	SH	
	1126		78.7	1257.1	1660.0	1776.1	-57.3	-43.9	0	0.9562	.0438	SO	
	1133		92.2	1261.0	1657.5	1804.2	-44.6	-32.0	0	0.9632	.0368	SH	
	1206		78.7	1256.1	1659.9	1818.5	-56.5	-38.5	0	0.9312	.0688	SO	
	1215		92.2	1260.8	M	1823.0	-42.8	-28.4		M		SH	Data rejected
	1240		78.7	1255.0	1608.2	1760.7	-49.3	-31.3	0	0.9109	.0891	SO	
	1245		92.2	1260.4	1624.2	1773.2	-34.8	-22.0	0	0.9205	.0795	SH	
	1305		78.7	1256.7	1623.1	1832.7	-53.0	-23.7	0	0.8068	.1932	SO	
	1313		92.2	1260.7	1637.7	1799.8	-35.4	-20.3	0	0.8859	.1141	SH	
	1330		78.7	1255.3	1639.9	1807.0	-46.4	-27.6	0	0.8835	.1165	SO	
	1339		92.2	1260.4	1638.8	1798.9	-35.2	-17.5	0	0.8242	.1758	SH	
	1354		78.7	1256.1	1641.2	1816.3	-49.8	-30.9	0	0.9136	.0864	SO	
	1405		92.2	1260.5	1621.3	1791.0	-36.0	-21.1	0	0.9138	.0862	SH	
	1420		78.7	1255.3	1646.7	1820.7	-52.4	-31.0	0	0.8725	.1275	SO	
	1429		92.2	1260.2	1673.4	1866.4	-42.5	-21.1	0	0.8242	.1758	SH	
	1452		78.7	1254.7	1623.6	1856.9	-49.6	-20.9	0	0.8288	.1712	SO	
	1523		92.2	1260.0	1639.7	1793.9	-37.7	-21.6	0	0.8635	.1365	SH	
	1545		92.2	1260.2	1643.2	1803.8	-41.5	-25.9	0	0.9056	.0944	SH	
	1615		92.2	1260.1	1635.5	1774.9	-39.9	-27.9	0	0.9490	.0510	SH	
	1640		92.2	1260.5	1658.6	1832.3	-43.1	-29.4	0	0.9663	.0337	SH	
	1706		92.2	1260.3	1652.9	1822.3	-44.0	-31.1	0	0.9854	.0146	SH	

Date	Time	Air temp	E (g)	W_1 (g)	W_2 (g)	W_3 (g)	T_1 (°C)	T_2 (°C)	T_3 (°C)	Q_f	Liquid water (%/100)	Oper. [1]	Remarks
3/15/79	0905		92.2	1261.1	1655.1	1776.8	-38.4	-31.3	-5.6	0.9977	.0023	SH	
	0926		92.2	1261.1	1646.7	1818.4	-43.5	-32.1	-2.8	1.0016		SH	
	1005		92.2	1261.3	1651.3	1780.5	-38.7	-28.3	-2.4	1.0021		SH	
	1017		78.7	1255.7	1658.8	1804.0	-47.1	-33.7	0	0.9596	.0404	HU	
	1027		92.2	1260.9	1632.7	1789.2	-33.9	-23.6	0	0.9766	.0234	SH	
	1044		78.7	1256.2	1634.4		-52.4	-34.2	0	M		HU	
	1053		92.2	1261.1	1677.0	1873.5	-43.6	-28.4	0	0.9572	.0428	SH	
	1107		78.7	1255.7	1642.1	1768.7	-56.3	-41.2	0	0.9421	.0579	HU	
	1116		92.2	1261.9	1642.1	1794.6	-42.2	-28.1	0	0.9318	.0682	SH	
	1135		78.7	1256.3	1646.1	1794.0	-53.4	-33.7	0	0.8631	.1369	HU	
	1158		92.2	1261.4	1651.0	1821.3	-42.8	-22.8	0	0.8313	.1687	SH	
	1208		78.7	1255.6	1629.7	1775.2	-53.1	-32.3	0	0.8427	.1573	HU	
	1221		92.2	1261.5	1651.5	1826.3	-42.9	-24.8	0	0.8789	.1211	SH	
	1246		78.7	1255.2	1654.8	1859.7	-44.9	-18.2	0	0.7732	.2268	HU	
	1254		92.2	1261.6	1660.4	1849.9	-43.8	-20.6	0	0.7993	.2007	SH	
	1317		78.7	1255.1	1655.1	1838.8	-53.1	-26.0	0	0.7748	.2252	HU	
	1337		92.2	1261.9	1676.2	1887.8	-42.2	-15.0	0	0.7387	.2613	SH	
	1349		92.2	1261.7	1688.2	1867.3	-40.0	-18.4	0	0.7727	.2273	SH	
	1401		78.7	1254.7	1657.5	1821.0	-44.7	-18.9	0	0.7033	.2967	HU	
	1420		92.2	1263.0	1644.0	1856.1	-40.0	-12.6	0	0.7456	.2544	SH	
	1447		78.7	1255.5	1668.1	1832.2	-50.0	-23.2	0	0.7062	.2938	HU	
	1508		92.2	1262.0	1637.6	1792.6	-35.0	-14.6	0	0.7549	.2451	SH	
	1608		78.7	1255.0	1670.0	1867.7	-45.9	-20.3	0	0.7770	.2230	HU	
	1640		92.2	1261.1	1651.3	1837.4	-37.8	-17.4	0	0.8191	.1809	SH	
	1650		78.7	1255.2	1620.4	1780.7	-50.9	-29.1	0	0.8478	.1522	HU	

[1] Operators -

SH = Steven Howell
SO = Stephen Olt
HU = Hub Ulrich

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TABLE 4.—SNOWPACK PROFILE CHARACTERIZATIONS AT THE STEAMBOAT SPRINGS SITE, MARCH 28, 1979

PIT Nº __3___   LOCATION: Steamboat Valley - 50'N. of 3-Mile Road on Flight Line

DATE: _3/28/79_   TIME: _1010___   TOTAL DEPTH: _71 cm___   AIR TEMPERATURE: _+4.5_   TABLE Nº _____

DEPTH TO BOTTOM OF LAYER	THICKNESS OF LAYER	SNOW CLASSIFICATION	GRAIN SIZE	LAYER TEMP. (°C)	DENSITY (kg/m³)		REMARKS
8 cm	8 cm	new snow	¼ to ½ mm	0	150		
24 cm	16 cm	melt-freeze metamorphosed	½ to 1 mm	0	460		
25 cm	1 cm	ice lens		--			
55 cm	30 cm	melt freeze metamorphosed firnification	½ to 1 mm	0	420		
71 cm	16 cm	depth hoar	½ to 1 mm	0	380		

TABLE 5.—COLD CALORIMETER DATA FOR LIQUID WATER CONTENT DETERMINATIONS ON MARCH 28-29, 1979 AT THE STEAMBOAT SPRINGS SITE.

Date	Time	Air temp °C Shd Sun	E (g)	W_1 (g)	W_2 (g)	W_3 (g)	T_1 (°C)	T_2 (°C)	T_3 (°C)	Q_f	Liquid water (%/100)	Oper.	Remarks
3/28/79	1030	5.2 5.0	78.7	1251.1	1659.9	1801.2	-47.2	-28.5	0	.8217	.1783	SH	Overcast sky
	1038	5.3 4.0	92.2	1257.8	1661.1	1803.6	-34.5	-18.5	0	.8105	.1895	SO	Overcast, light snow shower
	1055	7.5 5.8	78.7	1253.0	1638.2	1767.5	-53.6	-31.2	0	.7522	.2478	SH	Overcast, snow stopping
	1117	5.1 9.1	78.7	1252.8	1635.8	1805.9	-52.0	-24.2	0	.7391	.2609	SH	Sun shining through
	1139	3.1 8.6	92.2	1259.0	1645.3	1850.4	-37.2	-13.7	0	.7863	.2137	SO	Thin, overcast (T_4 could be -13.0)
	1145	3.1 8.6	78.7	1252.6	1634.8	1842.7	-50.6	-22.5	0	.7992	.2008	SH	Partial overcast (T_2 could be -21.7) (.2151)
	1205	4.4 8.8	92.2	1258.6	1630.6	1825.1	-36.3	-08.8	0	.6975	.3025	SO	Partial overcast
	1215	2.9 5.2	78.7	1252.3	1659.1	1850.0	-50.1	-19.3	0	.6935	.3065	SH	Solid overcast
	1236	2.6 4.7	92.2	1258.2	1643.0	1806.0	-37.1	-12.5	0	.6858	.3142	SO	Solid overcast, light snow starting (T_2 could)
	1255	5.3	78.7	1252.3	1648.1	1837.6	-50.2	-22.7	0	.7648	.2352	SH	Snowing heavily (.2945)(be -13.4)
	1316	5.8 4.6	92.2	1258.8	1649.0	1847.5	-39.5	-12.5	0	.7206	.2794	SO	Snowing moderately
	1357	3.5 7.2	78.7	1252.0	1666.7	1863.6	-44.5	-14.5	0	.6813	.3187	SH	Snow stopped, breaks in overcast
	1425	6.2 5.4	92.2	1258.3	1670.2	1859.3	-41.8	-15.6	0	.7166	.2834	SO	Partial overcast
	1448	5.3 9.1	78.7	1252.1	1648.7	1855.2	-44.7	-13.4	0	.6916	.3084	SH	Partial overcast, bright sun at times
3/29/79	1125	5.0 8.2	78.7	1320.7	1726.7	1901.8	-44.1	-25.8	0	.8809	.1191	SO	Partial overcast, N-NW winds 5-7 mph
	1137	4.2 6.2	92.2	1265.1	1662.0	1830.0	-35.8	-18.5	0	.8392	.1608	HU	Solid overcast, W-NW winds 7-10 mph
	1151	3.2 3.9	78.7	1321.1	1706.9	1904.7	-52.9	-28.3	0	.8574	.1426	SO	Solid overcast, W-NW winds 10-15 mph
	1208	2.8 3.6	92.2	1264.3	1640.2	1794.6	-41.4	-23.0	0	.8367	.1633	HU	Solid overcast, W-NW winds 10-15 mph, drifting snw
	1219	2.8 3.6	78.7	1320.9	1710.7	1858.5	-53.7	-34.2	0	.8691	.1309	SO	Solid overcast, W-NW winds 15 mph, blovng snow
	1242	3.0 4.7	92.2	1264.4	1650.6	1783.0	-41.3	-27.4	0	.8923	.1077	HU	Partial overcast, N-NW winds 15 mph, blowng snw
	1254	2.9 5.5	78.7	1320.6	1689.0	1848.8	-55.1	-32.3	0	.8479	.1521	SO	Partial overcast, N-NW winds 15 mph, blowng snw
	1315	3.4 5.8	92.2	1263.8	1673.7	1775.3	-40.1	-30.5	0	.9249	.0751	HU	Partial overcast, N winds 15-20 mph
	1328	2.7 6.1	78.7	1320.7	1689.9	1795.7	-52.1	-38.6	0	.9186	.0818	SO	Partial overcast, N winds 5-10 mph, Snw surfce very irregular due to new snow blown around
	1355	2.6 6.0	78.7	1320.5	1692.3	1883.2	-56.4	-29.6	0	.8356	.1644	SO	Partial overcast, N winds 5-10 mph
	1414	1.7 4.3	92.2	1264.8	1632.4	1746.1	-36.7	-24.6	0	.8828	.1172	HU	Partial overcast, N winds 5-10 mph, crust frmng on surface
	1425	1.7 3.3	78.7	1320.8	1729.1	1847.3	-48.4	-34.0	0	.8810	.1190	SH	Solid overcast, N-NE winds 5-10 mph
	1443	1.3 2.5	92.2	1264.3	1683.7	1790.0	-40.4	-29.9	0	.9048	.0952	HU	Solid overcast, N-NE winds 5-10 mph, light snow starting
	1446	1.3 2.5	78.7	1321.2	1693.0	1808.4	-55.2	-40.2	0	.9200	.0800	SH	Solid overcast, N-NE winds 5-10 mph, light snow

TABLE 6.—COLD CALORIMETER DATA FOR LIQUID WATER CONTENT DETERMINATIONS AT VARIOUS LAYERS IN THE SNOWPACK ON MARCH 30, 1979 AT THE STEAMBOAT SPRINGS SITE.

Depth	Time	Air temp °C shd	Air temp °C sun	E (g)	W_1 (g)	W_2 (g)	W_3 (g)	T_1 (°C)	T_2 (°C)	T_3 (°C)	Q_f	Liquid water (%/100)	Oper.	Remarks
0-6cm	0846	1.8	0.4	92.2	1351.1	1754.2	1812.6	-43.6	-38.1	0	.9719	.0281	HU	Partial overcast, winds calm
	0848	2.0	2.8	78.7	1320.0	1722.3	1787.1	-54.5	-46.7	0	.9586	.0414	SH	Sunny, calm
	0913	0.8	6.1	92.2	1351.9	1736.2	1812.0	-44.7	-36.5	0	.9367	.0633	HU	Sunny, calm
	0917	0.8	6.1	78.7	1320.4	1709.6	1800.3	-60.8	-48.6	0	.9421	.0579	SH	Sunny, calm
6-12cm	0943	0.7	9.1	92.2	1352.2	1762.1	1924.2	-45.4	-31.4	0	.9524	.0476	HU	Partly cloudy, calm
	0946	0.8	5.3	78.7	1321.0	1697.8	1895.5	-60.1	-37.4	0	.9386	.0614	SH	Partly cloudy, calm
12-23cm	1010	1.6	5.3	92.2	1351.7	1760.5	2017.2	-40.2	-18.4	0	.8812	.1188	HU	Overcast, calm
	1015	2.9	4.4	78.7	1321.0	1712.4	2000.0	-59.2	-29.5	0	.9144	.0856	SH	Overcast, calm
24-60cm	1037	5.8	5.1	92.2	1351.8	1735.4	1876.2	-40.2	-24.4	0	.8577	.1423	HU	Overcast, calm
	1044	2.9	7.2	78.7	1320.0	1682.0	1851.8	-52.7	-29.8	0	.8570	.1430	SH	Partial overcast, calm
60-70cm	1108	5.6		92.2	1351.9	1760.0	1920.8	-37.7	-23.1	0	.8932	.1068	HU	Partly cloudy, sunny, calm
	1113	5.8		78.7	1321.6	1717.8	1914.7	-52.6	-30.9	0	.9015	.0985	SH	Partly cloudy, sunny, calm
0-6cm	1134	4.6	4.4	92.2	1351.5	1722.4	1838.0	-36.7	-19.9	0	.7557	.2443	HU	Partly cloudy, sunny, calm
	1139	4.3	6.4	78.7	1320.7	1698.1	1821.6	-53.3	-31.1	0	.7428	.2572	SH	Partly cloudy, sunny, calm
0-6cm	1158	3.0	8.5	92.2	1352.5	1711.6	1877.6	-36.7	-14.7	0	.7652	.2348	HU	Sunny, calm
	1200	3.0	8.5	78.7	1321.4	1700.1	1844.7	-53.5	-26.7	0	.7025	.2975	SH	Sunny, calm
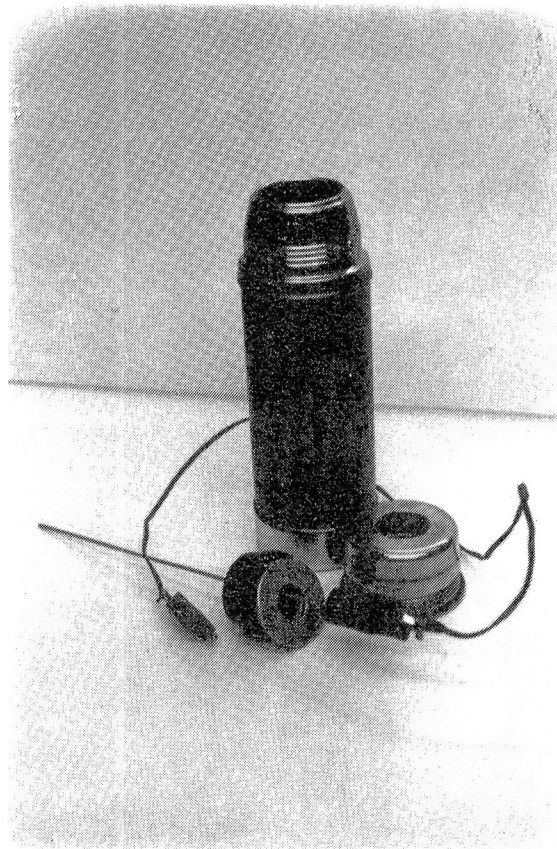
Figure 1. Calorimeter Bottle with Temperature Probe
and Rubber Stopper.
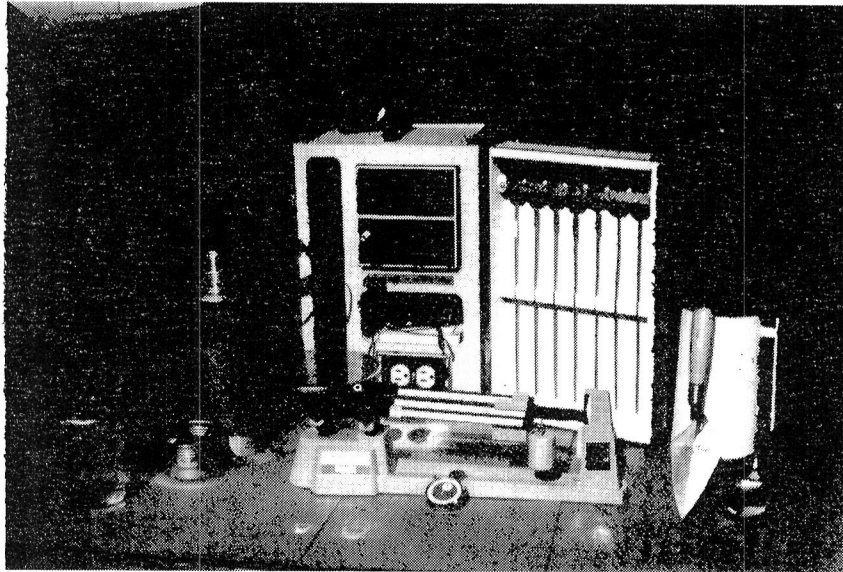
Figure 2.   Cold calorimeter equipment set including calorimeter
bottle, digital readout for temperature probes, thermos
bottle for collection of samples, scales, trowel and spoon,
temperature probes, clock, stopwatch, and container of
freezing agent.

Figure 3.   Calorimeter field facility in operation near
Frazer, Colorado.

TARE WT. OF BOTTLE = 1040.0 g
TARE WT. + COLD SILICONE = 1371.7 g   Wt.cold = 331.7 g
TARE WT. + COLD + WARM SILICONE = 1478.4 g   Wt.warm = 106.7 g

$T_W = +7.6°C$

$$(Wt_{warm})(\frac{Cs_W + Cs_2}{2})(T_W - T_2) = (Wt_{cold} + E)(\frac{Cs_1 + Cs_2}{2})(T_2 - T_1)$$

$$(106.7g)(.4373)(37.6°) = (331.7g + E)(.4328)(9.7°)$$

$$E = 86.2 g$$

$T_2 = -30.0°$

$\Delta T = 9.7°$

$T_1 = -39.7°$

TIME	TEMP
0	-45.5
1	-44.2
1.5	-43.6
2	-43.1
2.5	-42.8
3	-42.5
3.5	-42.3
4	-42.0
4.5	-41.8
5	-41.5
5.5	-41.3
6	-41.0
6.5	-40.8
7	-40.4
7.5	-40.2
8	-39.9
8.5	
9	-29.7
9.5	-29.8
10	-29.6
10.5	-29.5
11	-29.4
11.5	-29.2
12	-29.1
12.5	-29.0
13	-28.8
13.5	-28.7
14	-28.6

TEMPERATURE (°C)

TIME (MINUTES)
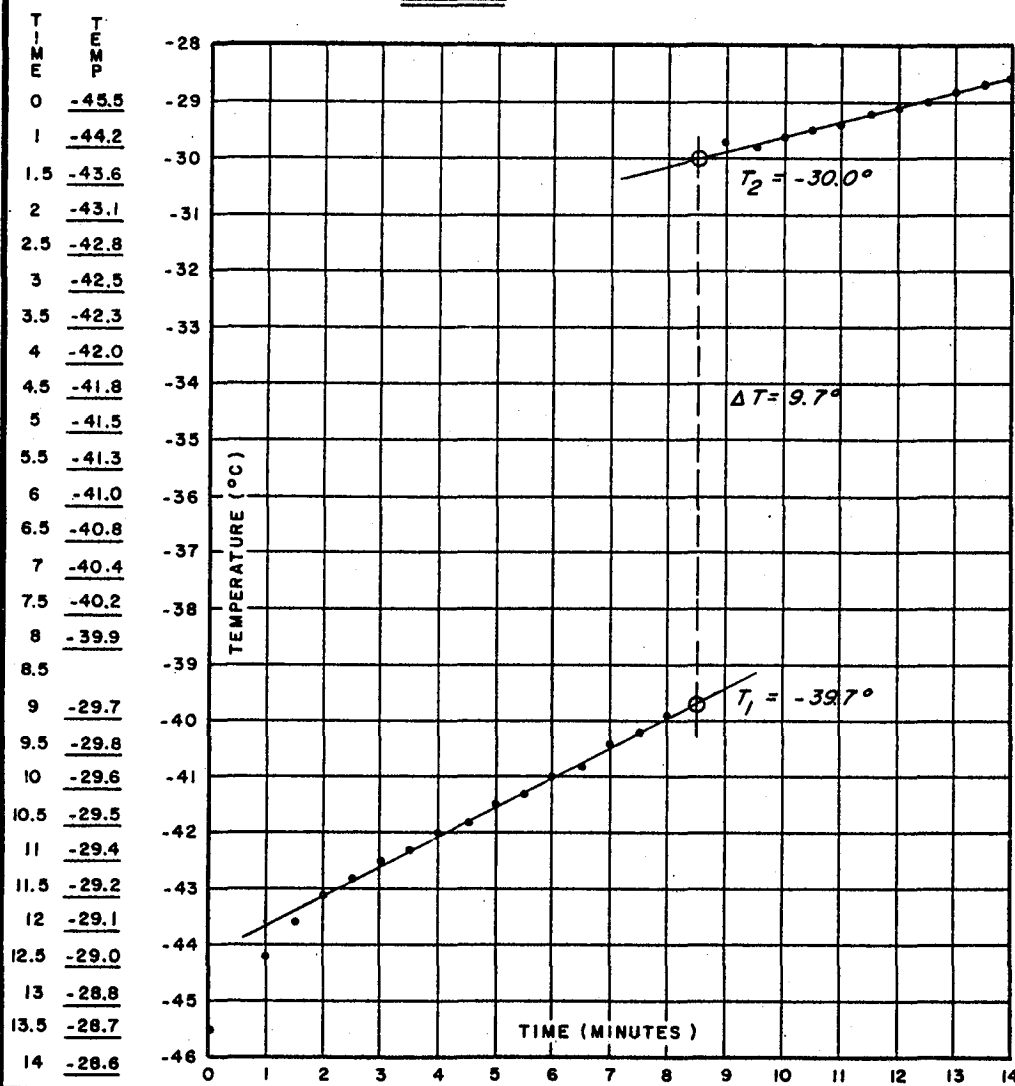
Figure 4. Typical calorimeter constant calculation.

MEASUREMENT OF SNOW QUALITY
(Freezing Calorimetric Technique)

Time	Temp°C
0	-58.5
1	-55.3
1.5	-54.2
2	-53.3
2.5	-52.3
3	-51.4
3.5	-50.6
4	-49.8
4.5	-49.1
5	-48.3
5.5	-47.6
6	-46.8
6.5	-46.1
7	-45.5
7.5	-44.8
8	-44.2
8.5	—
9	—
9.5	—
10	-26.4
10.5	-24.1
11	-23.6
11.5	-23.3
12	-23.0
12.5	-22.7
13	-22.4
13.5	-22.1
14	-21.8
14.5	—
15	—
15.5	—
16	—
16.5	—
17	—
17.5	—

Station __Fraser Valley__ Observer __SH__

Date __3/15/79__   Hour __1221__

Location and description of sampling point _____

__Top 4 cm of snow at Fraser Valley Snow Pit__

_____

Data

Sample thermos No. __105__ Air Temperature __+4.0__ °C

Height of sample from ground surface __34.5__ inches.

(1) Tare weight of calorimeter __1261.5__ gr. (W_1)

(2) Weight of calorimeter and silicone fluid __1651.5__ gr. (W_2)

(3) Weight of calorimeter + silicone fluid + snow __1826.3__ gr. (W_3)

(4) Calorimeter constant (E) __92.2__ gr.

(5) $t_1 = -42.9°$ $t_2 = -24.8°$ $t_3 = 0°$

(6) (2) − (1) = __390.0 gr.__

(7) (3) − (2) = __174.8 gr.__

Snow Quality

$$Q_f' = 1 - \frac{[WiC_f (t_2 - t_1) - C_i (t_2 - t_3)]}{LS \quad \quad L} \quad = \boxed{0.879}$$

% free water = $100 (1 - Q_f')$ $= \boxed{12.1\%}$

where

$Wi = W_2 - W_1 + E$, gr.

C_f = specific heat of freezing agent at average temp.
$$\frac{(t_1 + t_2)}{2}$$

t_1 = initial temp. °C
t_2 = final temp. °C
t_3 = snow temp. °C

C_i = specific heat of ice at average temp.
$$\frac{(t_2 + t_3)}{2}$$

$S = W_3 - W_2$, gr.
L = latent heat of melting

RCI form # 751-A
1-79

Figure 5. Data form for calculation of snow quality and liquid water content with data taken at the Frazer Valley site at 1221 hours on March 15, 1979.
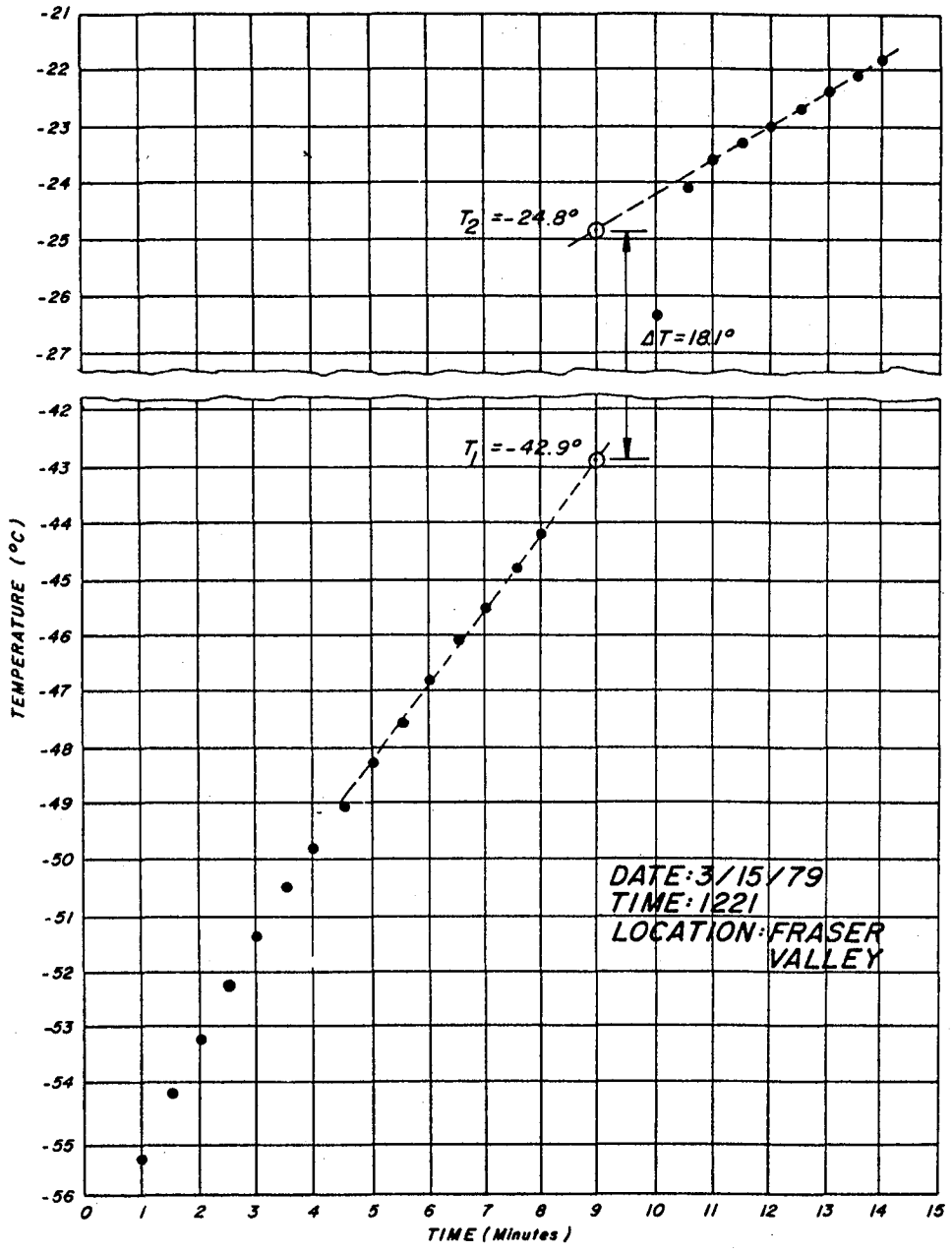
Figure 6. Graphical determination of t_1 and t_2 at the Frazer Valley site at 1221 hours on March 15, 1979.
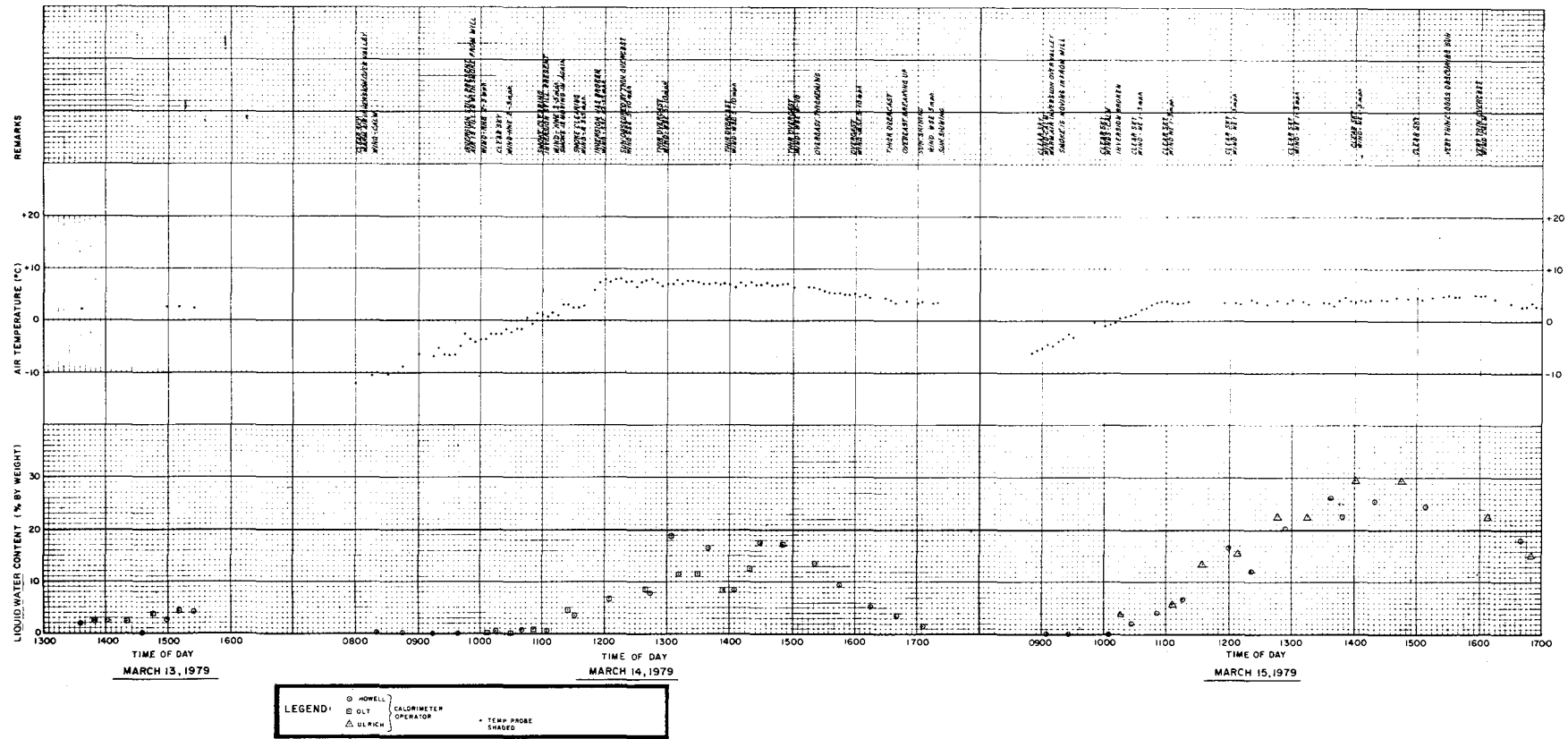
Figure 7. Graphical presentation of air temperature and liquid water at
the Frazer Valley site, March 13-15, 1979

THE DIELECTRIC BEHAVIOUR OF SNOW: A STUDY

VERSUS LIQUID WATER CONTENT

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ABSTRACT

Snow is treated as a heterogeneous dielectric material consisting of ice, air and water. The great difference in the high frequency relative permittivity of dry snow and water allows to determine the liquid water content by measurements of the relative permittivity of snow. A plate condenser with a volume of about 1000 cm^3 was used to measure the average liquid water content in a snow volume. Calibration was carried out using a freezing calorimeter.

An important aspect in the investigation of the snow surface by remote sensing techniques using VHF or UHF is the augmented liquid water content near the snow surface. In order to measure the liquid water content in thin snow layers, a comb-shaped condenser was developed, which is the two-dimensional analogon of the plate condenser. With this moisture meter the liquid water content was measured in layers of a few millimeters in thickness, whereby the effective depth of measurement is given by the penetration depth of electric field lines which is controlled by the spacing of the strip lines. Results of field measurements with both moisture meters, the plate condenser and the comb-shaped condenser, are given.

In order to find the proper measuring frequency for the moisture meters, the frequency dependence of the relative permittivity of wet snow was measured in the range of up to 100 MHz. The influence of both the stage of metamorphosis and the distribution of liquid water in snow is discussed in detail. The formula of Polder and van Santen is applied for calculating the relative permittivity of snow from the components ice, air and water. This model is a basis for the physical conception of snow moisture meters. From this analysis it results that the calibration of the dielectric moisture meters may be replaced by model calculations according to the formula of Polder and van Santen.

INTRODUCTION

The liquid water content of a snow cover determines many of its physical characteristics, so that attention must be attributed to its precise and efficient measuring. In connection with

remote sensing techniques it is of interest how the liquid water content of the snow pack influences the velocity and attenuation of electromagnetic waves (refs. 1, 2, 3), the albedo (ref. 4) and the brightness temperature in the microwave range.

METHODS OF MEASURING THE LIQUID WATER CONTENT

A comparison of various methods of determining the liquid water content of snow shows that dielectric methods are outstandingly suited for field measurements (refs. 6, 7).

The dielectric methods come under the indirect measuring methods of liquid water content, as they register characteristics of the material to be measured that depends on the liquid water content. Therefore a relationship between the dielectric constant and the liquid water content must be found by calibration. However, the dielectric method offers the advantage of quick readings (ref. 8).

In contrast to this advantage, the direct measuring methods of the liquid water content are in general time-consuming and only partly suited for field measurements (ref. 6). Among these methods the following have been described in detail in relevant literature: calorimetric ones with a freezing calorimeter (ref. 9) or with a hot water calorimeter (ref. 10) and centrifugal ones (ref. 11).

DIELECTRIC METHODS OF MEASURING THE LIQUID WATER CONTENT

The complex dielectric constant of a material is given by the two components ε' and ε'', whereas ε' means the relative permittivity and ε'' the loss factor. Dielectric measuring methods of the liquid water content make use of the substantial difference between the relative permittivities of dry snow and of water. As previous investigations have shown, the complex dielectric constants of ice and snow have a distinct frequency dependence in the range of $\nu < 100$ kHz (refs. 12, 13, 14). In this connection it is essential to analyse over a wide frequency range how the liquid water content and the stage of metamorphosis influence the relative permittivity. These analyses supply the physical bases of choosing a suited measuring frequency for the measuring of the liquid water content of snow by means of the dielectric method.

SNOW AS A HETEROGENEOUS DIELECTRIC MATERIAL

Snow is considered as a heterogeneous dielectric material consisting of ice, air and water. The relative permittivity and the loss factor of this system are determined by the relative permittivities and the loss factors of the individual components. Furthermore, the volume filling factors, the distribution and the shape of the individual components control the relative permittivity of snow. However, this complex system can be seen in a simplified version as a two-component system consisting of dry snow and of water inclusions.

RELATIVE PERMITTIVITY OF SNOW VERSUS FREQUENCY

An essential requirement of a dielectric moisture meter is that its reading does not depend on the size and shape of the snow grains varying in the course of the metamorphosis. Measurements of the frequency dependence of the relative permittivity of wet snow in the range of between 80 Hz and 100 MHz have shown that in the case of measuring frequencies exceeding 10 MHz the relative permittivity only depends on the liquid water content and on the porosity, but not on the stage of metamorphosis. In the range of between 80 Hz and 20 MHz the measurements were carried out by means of a Wien-Robinson bridge and a twin-T bridge; the measurements in the range of between 500 kHz and 100 MHz were made by means of two-port measurements with the voltage method using a vector-analyzer. In fig. 1, for comparison purposes, the frequency dependence of both the permittivity and the loss tangent of two types of snow with approximately the same liquid water content but different stage of metamorphosis is illustrated. Fig. 1a shows the frequency dependence in a log-scale of a fine grained sample (D < 0,5 mm), fig. 1b of a coarse grained one (D > 0,5 mm). In both snow samples a wide relaxation regime in the range of $\nu <$ 100 kHz occurs, whereas the relaxation frequency of the snow samples is substantially higher than in the case of pure ice (ref. 15).

The fine grained sample (fig. 1a) shows in comparison with the coarse grained sample (fig. 1b) a wide relaxation regime and the loss tangent decreases monotonously, without passing through a marked maximum. With the coarse grained sample (fig. 1b), however, a marked maximum of the loss tangent in the range of about 30 kHz occurs. Both samples show a constant value of the relative permittivity in the frequency range exceeding 10 MHz, the fine grained sample reaching a constant value at approximately 6 MHz, the coarse grained sample at 1 MHz already.

In a previous paper the frequency dependence of the relative permittivity and the loss tangent of snow samples have been analyzed by means of the Debye and the Cole-Cole models (ref. 14).

The result of these analyses is that in general for coarse grained samples the simple Debye model with just one relaxation frequency can be applied. In the case of fine grained samples the Cole-Cole model can be applied, which is based on a spectrum of relaxation frequencies.

For frequencies exceeding 10 MHz, regardless of the grain size and hence of the stage of metamorphosis, a value of the permittivity is reached that does not depend on the frequency. This constant value means the high frequency relative permittivity ε'_∞. Fig. 2, for example, illustrates how the density-reduced relative permittivity $e = \varepsilon'-1-2,2\varrho$ depends on the liquid water content of coarse and fine grained snow samples, using a measuring frequency of 13,6 MHz. Results for measuring frequencies of 1,0 MHz, 5,0 MHz, 9,3 MHz and 18,6 MHz are given in ref. 16. Instead of the relative permittivity ε' here the density-reduced permittivity e was introduced in order to allow for the dependence of the permittivity on the density of snow. Within the standard deviation $\pm\sigma$ no significant difference between coarse grained samples (D > 0,5 mm) and fine grained samples (D < 0,5 mm) can be observed. It results, that the density-reduced permittivity e thus only depends on the liquid water content and not on the stage of metamorphosis.

HIGH FREQUENCY RELATIVE PERMITTIVITY, WET SNOW DENSITY AND LIQUID WATER CONTENT

Dry snow samples

Fig. 3 illustrates the influence of porosity on the high frequency relative permittivity $\varepsilon'_{\infty,d}$ of dry snow. As a good approximation a linear relationship between the relative permittivity $\varepsilon'_{\infty,d}$ and the porosity is obtained:

$$\varepsilon'_{\infty,d} = 1 + 2,02(1-\phi) \tag{1}$$

The dry snow samples, shown in fig. 3, were divided into three groups as to their grain size, i.e. D < 0,4 mm, 0,4 ≤ D < 1 mm and D > 1 mm. No systematic influence of the grain size and consequently of the stage of metamorphosis can be observed.

Wet snow samples

Fig. 4 illustrates the dependence of the density-reduced high frequency relative permittivity $e_\infty = \varepsilon'_\infty -1 - 2,2\varrho$ on the liquid water content in the case of wet snow samples. The wet

snow samples were divided into two groups according to their grain size, i.e. $D \leq 0,5$ mm and $D > 0,5$ mm, the liquid water content being determined by means of a freezing calorimeter. From fig. 4 a linear relationship results between the density-reduced high frequency relative permittivity e_∞ and the liquid water content W, whereas no systematic influence of the grain size having been observed:

$$e_\infty = 0,213 \ W \tag{2}$$

THE MODEL OF POLDER AND VAN SANTEN APPLIED TO WET SNOW

Of the many models quoted in relevant literature (refs. 17, 18) that lead to the calculation of the permittivity of a heterogeneous material, the model of Polder and van Santen is especially suited for wet snow (refs. 19, 20). As compared to others it has the outstanding advantage of offering a possibility of expressing the orientation and the shape of the components of a heterogeneous system by means of shape factors. The mixing formula of Polder and van Santen adapted to the heterogeneous system of wet snow (ref. 20) reads as follows:

$$\mathcal{E}_\infty' - \frac{W}{3}\left[\mathcal{E}_W - \mathcal{E}_{\infty,d}'\right]\mathcal{E}_\infty' \cdot \sum_i \frac{1}{\mathcal{E}_\infty' + (\mathcal{E}_W - \mathcal{E}_\infty')g_i} - \mathcal{E}_{\infty,d}' = 0 \tag{3}$$

The high frequency relative permittivity $\mathcal{E}_\infty'$ of the heterogeneous system can be calculated from the relative permittivity of water, the porosity of snow and the depolarizing factors g_i by means of equation (3).

Fig. 5 shows a comparison between the calculated relative permittivity $\mathcal{E}_\infty'$ (model) obtained from equation (3) and the measured value $\mathcal{E}_\infty'$ (measurement), the depolarizing factor g_1 having been optimated under the assumption of water inclusions in the form of oblate spheroids. In this particular case the components g_2, g_3 are given by $g_2 = g_1$ and $g_3 = 1 - 2g_1$. An influence of the grain size, and thus of the stage of metamorphosis, cannot be observed. The slope of the regression line in fig. 5 shows that the model of Polder and van Santen is suited to the application to wet snow. In order to calculate the relative permittivity of wet snow, snow density and liquid water content have to be known. The agreement between measured and calculated values of the high frequency relative permittivity of wet snow allows the conclusion that the liquid water content of wet snow samples can be calculated from equation (3). This analysis may be suitable to replace the calorimetric calibration.

Whereas in fig. 5 the regression line was obtained by applying an averaged depolarizing factor, in fig. 6 the depolarizing factor g_1 was calculated for each individual sample (refs. 21, 22). From this analysis a dependence of the depolarizing factor g_1 on the liquid water content is obtained (fig. 6). At low values of the liquid water content the depolarizing factor g_1 increases monotonously, whereas in the range of between 5% to 7% per volume a significant change of it is marked (ref. 23). This change may be attributed to the transition from the pendular to the funicular regime.

THE MODEL OF WIENER

The formzahl u in Wiener's mixing formula (ref. 24) is also suited to describe wet snow as a heterogeneous system, although with certain restrictions (ref. 20). Applying the model of Wiener, the formzahl u depends on both the liquid water content and the stage of metamorphosis (refs. 25, 26). In fig. 7 the course of the formzahl u versus liquid water content is shown for fine grained ($D < 0,5$ mm) and coarse grained ($D > 0,5$ mm) snow samples. For values of the liquid water content exceeding 4% per volume both groups of samples (figs. 7a, 7b) show the identical course of the formzahl u, whereas for values lower than 4% per volume an additional influence of the stage of metamorphosis, i.e. grain size, is clearly to be observed. However, the transition of the distribution of liquid water from the pendular to the funicular regime, which may occur in the range of 5% to 7% per volume is not reflected by a change of the formzahl u.

A comparison of the models of Wiener and the one of Polder and van Santen shows that the model of Polder and van Santen is preferably used to describe wet snow as heterogeneous dielectric material: The model of Polder and van Santen is a more comprehensive one, containing the model of Wiener as a special case (ref. 17).

MEASURING THE BULK WATER CONTENT

MEASURING METHOD

For measurements of the complex dielectric constant in the frequency range of up to approximately 100 MHz bridge-methods have proved their efficiency. In the frequency range exceeding 10 MHz, which is of particular interest in connection with measuring the liquid water content, twin-T bridges and a bridged-T bridge are preferable for measurements of the complex dielectric constant. On the basis of the experimental results

discussed here, two types of instruments with set frequencies in the range of approximately 20 MHz have been developed to determine the liquid water content of snow.

THE TWIN-T BRIDGE

The twin-T bridge (refs. 27, 28) was built to a set frequency of 18 MHz; a plate condenser with a measuring volume of 1000 cm^3 consisting of 5 plates with a plate spacing of 3 cm was connected thereto by means of a coaxial cable. After completed tuning both the relative permittivity $\mathcal{E}'$ and the loss factor $\mathcal{E}''$ can be determined separately.

The twin-T bridge has been mentioned on purpose, as in contrast to other bridges, it allows a common earthing of the sample to be measured, of the generator and the reading instrument. Absolute calibration of the instrument was carried out in the laboratory by means of a freezing calorimeter. During field measurements an absolute accuracy of ± 0,5% per volume is reached when determining the liquid water content. It is assured by the accuracy of the adopted calibration method.

BRIDGED-T BRIDGE

In order to simplify the rather time-consuming tuning operation for both, the component $\mathcal{E}'$ and the component $\mathcal{E}''$, a bridged-T bridge was developed. Comparative measurements, using a twin-T bridge and a bridged-T bridge show that the tuning operation for the loss factor can be avoided, if proper components of the network of the bridged-T bridge are chosen: in this way a one-knob-moisture meter is obtained (ref. 29).

By means of the moisture meters developed in this way, the twin-T bridge with two-knob-tuning and the bridged-T bridge with one-knob-tuning, the average liquid water content of a snow volume of approximately 1000 cm^3 can be measured. A substantial reduction of the measuring volume, which would be desirable for many purposes, is possible only to a certain extent. When the spacing of the plates is too small, the mechanical obstruction of the sample is very relevant; if the plates are too short the electric stray field of the plate condenser cannot be neglected any longer.

For measurements of the liquid water content in small volumes or in thin layers a special sensor has been developed, which is described in the following section.

MEASURING THE LIQUID WATER CONTENT IN THIN LAYERS

The increased liquid water content and the big gradient of the liquid water content near the surface are important quantities when analyzing the snow surface by means of VHF and UHF remote sensing techniques. For this special case of measuring the liquid water content in thin layers a comb-shaped condenser was developed (ref. 29), which is the 2-dimensional analogon of the plate condenser. Fig. 8 illustrates the geometrical shape of the condenser in use. As a good approximation the capacity per square unit of such a comb-shaped condenser can be expressed as follows

$$\frac{C}{A} = \frac{1}{1+s} \frac{\varepsilon_0}{\pi} \text{arcosh}[\frac{1}{s} + 1] \qquad (4)$$

and the effective depth of penetration of the electric field as a figure of the measured snow volume is approximately given by (refs. 29, 30)

$$b = \sqrt{\frac{1}{2}(\frac{1}{2} + s)} \qquad (5)$$

This effective measuring depth b is illustrated in fig. 9 for various strip-line spacings s depending on the ratio l/s, width to spacing of the strip lines. The effective measuring depth b can be varied over a wide range by choosing suitable spacings and widths of the strip lines.

Using a moisture meter with a comb-shaped sensor for measuring the liquid water content in thin layers, the change in the capacitance of the sensor is indicated by a change in the frequency of an oscillator. An analog reading of the liquid water content is obtained by a frequency to voltage converter. The comb-shaped sensor in this moisture meter was designed for an effective depth of measurement of 3,5 mm, the total surface of the sensor being 4x4 cm^2. Calibration was carried out with homogeneous samples by means of a freezing calorimeter.

CONCLUSIONS

The dielectric measuring method for the determination of the liquid water content of snow is particularly suited for field work. The reasons are that these dielectric moisture meters are simple to handle and allow fast measurements. Accuracy

is assured due to the calibration procedure reaching $\pm$ 0,5% per volume, a freezing calorimeter being the best calibration instrument. Two embodiments of a moisture meter have been tested: one instrument with a plate condenser for measuring the average liquid water content in a snow volume, and a second one with a comb-shaped condenser for measuring the liquid water content in thin snow layers near the surface of a snowpack.

Systematic studies of the frequency dependence of the complex dielectric constant of snow samples of various stages of metamorphosis and liquid water contents were made, which offered the physical basis for the development of the instruments. Both dielectric moisture meters, the one with the plate condenser and the other with the comb-shaped condenser, work with a set frequency. The frequency must be higher than 10 MHz, whereby a set frequency of 20 MHz was proved to be most suitable. Using this frequency the relative permittivity of snow only depends on both the liquid water content and the porosity, but is independent of the grain size and consequently of the stage of metamorphosis.

SYMBOLS

A	surface of the comb-shaped condenser
b	penetration depth of the electric field of the comb-shaped condenser
C	capacity of the comb-shaped condenser
e	density-reduced dielectric constant
e_∞	density-reduced high frequency dielectric constant
g	depolarizing factor in the mixing formula of Polder and van Santen
l	width of the strip lines of the comb-shaped condenser
n	number of strip lines of the comb-shaped condenser
s	spacing of the strip lines of the comb-shaped condenser
u	formzahl in the mixing formula of Wiener
W	water content; water volume/total volume
z	length of the strip lines of the comb-shaped condenser
ε'	relative permittivity
ε'_∞	high frequency relative permittivity of wet snow
$\varepsilon'_{\infty,d}$	high frequency relative permittivity of dry snow
ε'_W	relative permittivity of water at $0^\circ C$
ε_0	permittivity of free space
ε''	loss factor
ν	frequency
ϱ	density of dry snow
σ	standard deviation
ϕ	porosity; pore volume/total volume

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LEGEND TO FIGURES

Fig. 1: Frequency dependence of the relative permittivity and the loss tangent of a fine grained snow sample (a) and a coarse grained samble (b). Fine grained snow samples are samples with an average grain size $D < 0,5$ mm, coarse grained with $D > 0,5$ mm.

Fig. 2: Dependence of the density-reduced relative permittivity $e = \mathcal{E}' - 1 - 2,2\varrho$ on the liquid water content of coarse grained and fine grained samples at a measuring frequency of 13,6 MHz. The bottom diagram shows that the course of fine grained and of coarse grained samples coincide within the standard deviation $\pm \sigma$. Fine grained snow samples are samples with an average grain size $D < 0,5$ mm, coarse grained with $D > 0,5$ mm. Diagram of other set frequencies cf ref. 16.

Fig. 3: Dependence of high frequency relative permittivity $\mathcal{E}'_{\infty,d}$ on the porosity ϕ of dry snow samples. The samples were divided into 3 groups according to their grain size D.

Fig. 4: Empirical relationship between the density-reduced high frequency relative permittivity $e_\infty = \mathcal{E}'_\infty - 1 - 2,2\varrho$ and the liquid water content obtained by a calorimetric method (% per volume) for coarse grained and fine grained samples. No systematic difference between both kinds of samples can be observed. Fine grained snow samples are samples with an average grain size $D < 0,5$ mm, coarse grained with $D > 0,5$ mm.

Fig. 5: A comparison of measured values of the high frequency relative permittivity $\mathcal{E}'_\infty$ (measurement) with the calculated values $\mathcal{E}'_\infty$ (model) using the mixing formula of Polder and van Santen for coarse grained and fine grained samples. Dry snow samples are marked by empty symbols, wet snow samples by full ones. The inserted regression line shows that on an average the calculated values coincide satisfactorily with the measured ones. Fine grained snow samples are samples with an average grain size $D < 0,5$ mm, coarse grained with $D > 0,5$ mm.

Fig. 6: Dependence of the depolarizing factor g_1 on the liquid
water content of snow samples, the liquid water content
being measured by means of the calorimetric method.
Coarse grained and fine grained snow samples have dif-
ferent symbols. In the range of the transition from
the pendular regime into the funicular one the depola-
rizing factor g_1 undergoes a change. Fine grained snow
samples are samples with an average grain size $D < 0,5$ mm,
coarse grained with $D > 0,5$ mm.

Fig. 7: Plot of the formzahl u according to Wiener versus the
liquid water content obtained by means of a freezing
calorimeter for coarse grained and fine grained snow
samples. Fine grained snow samples are samples with an
average grain size $D < 0,5$ mm, coarse grained with
$D > 0,5$ mm.

Fig. 8: Geometrical shape of the comb-shaped condenser

Fig. 9: Effective measuring depth of a comb-shaped condenser
depending on the shape of the comb-shaped condenser,
l meaning the width, s the spacing of the strip lines
(cf fig. 8).
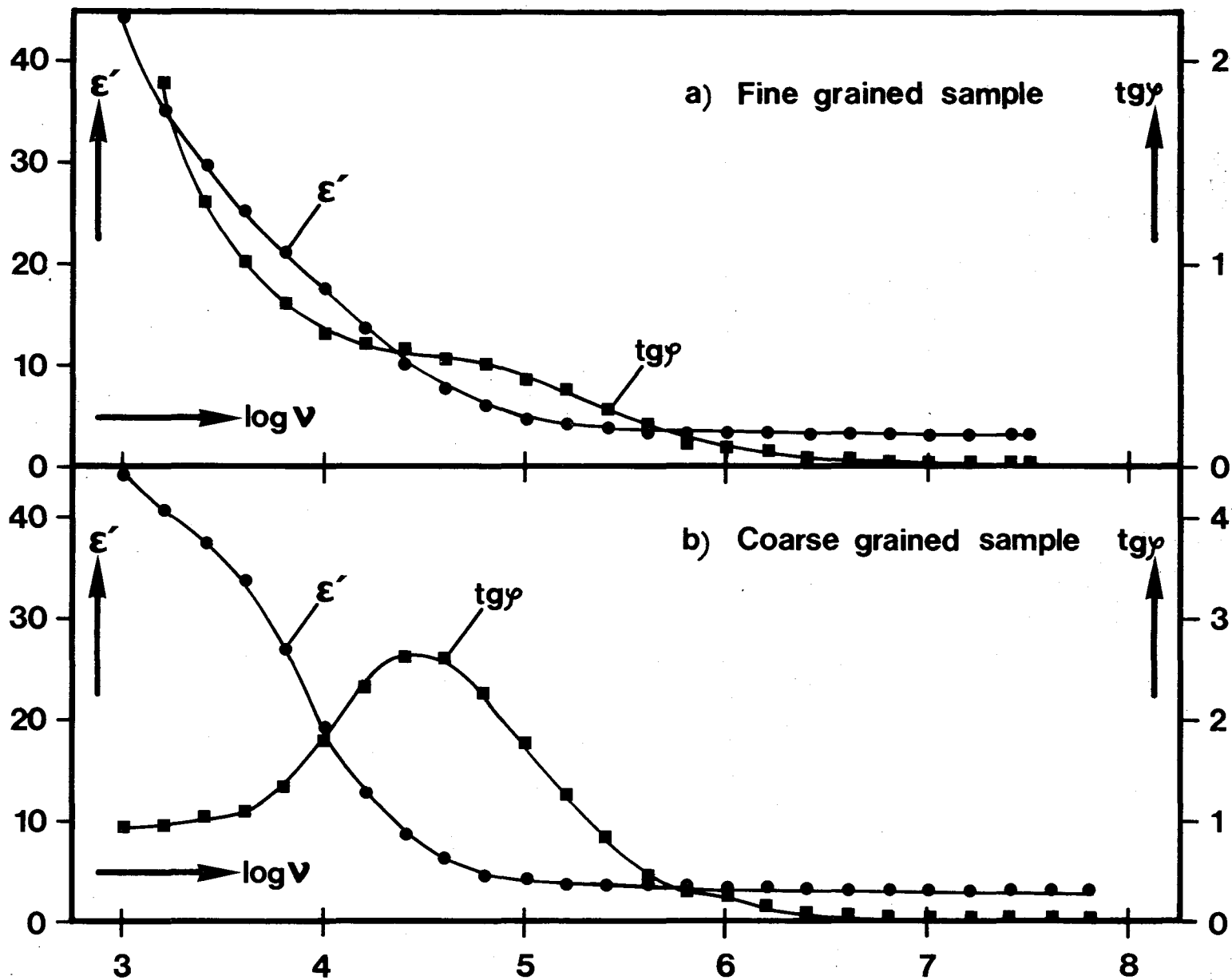
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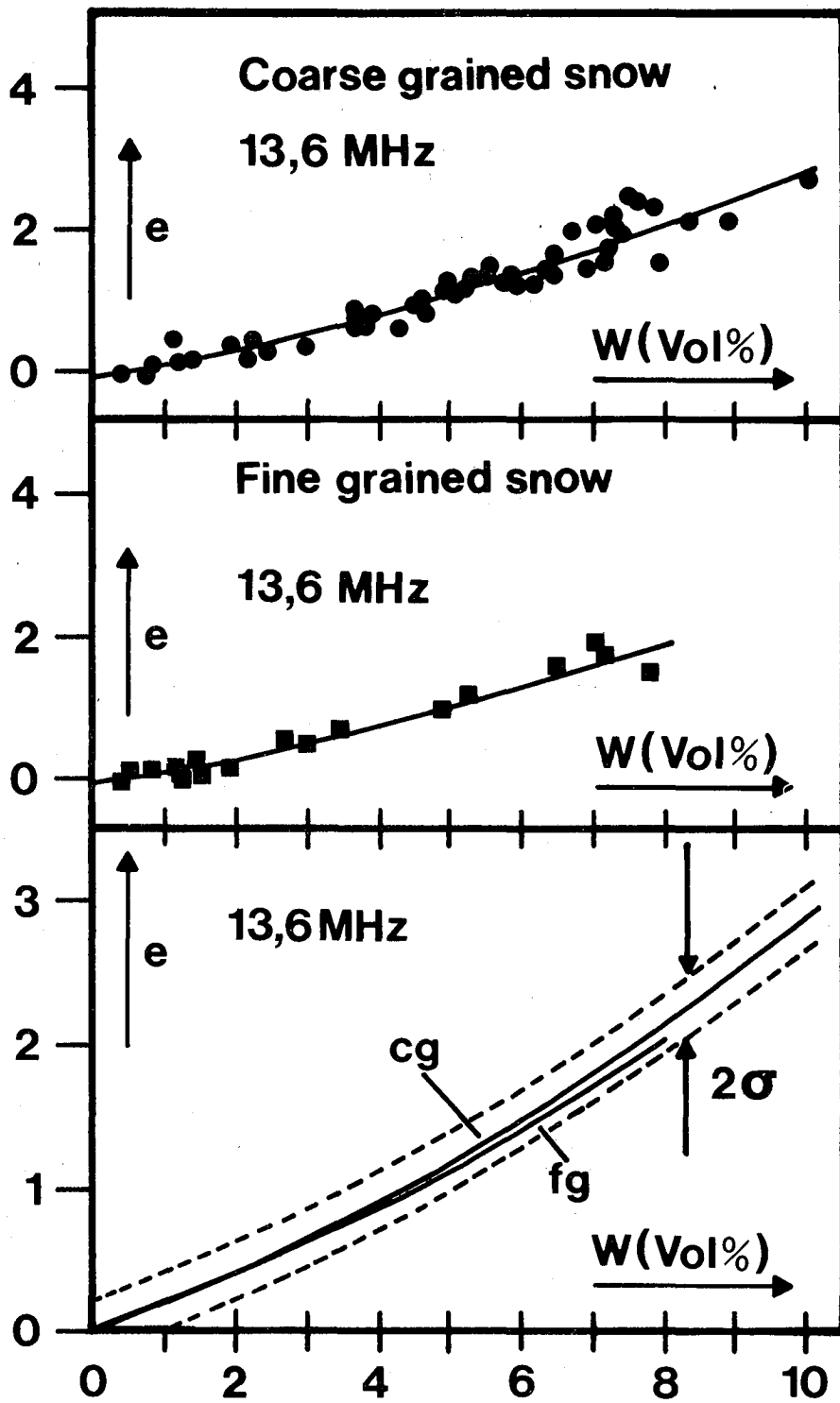
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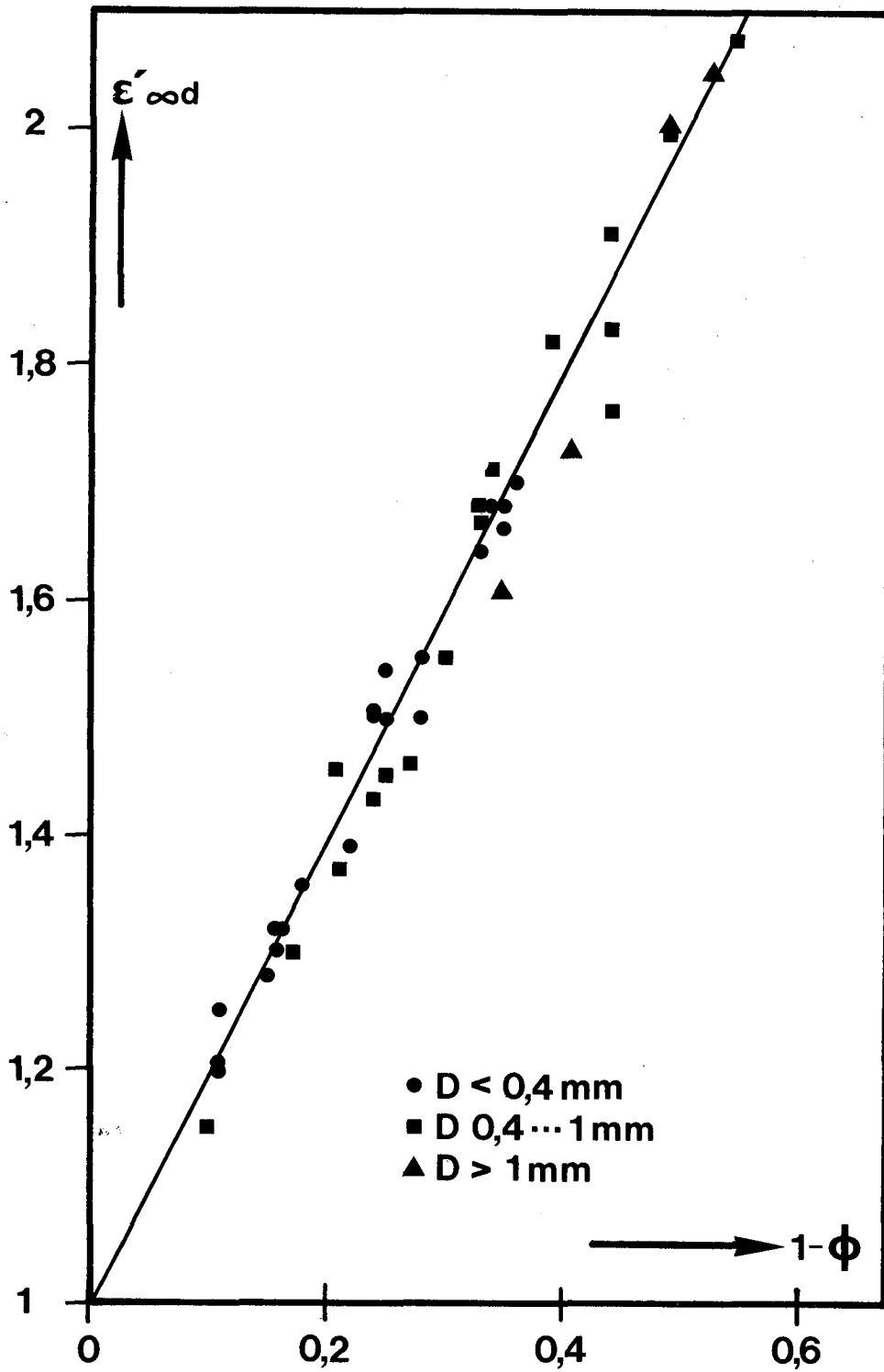
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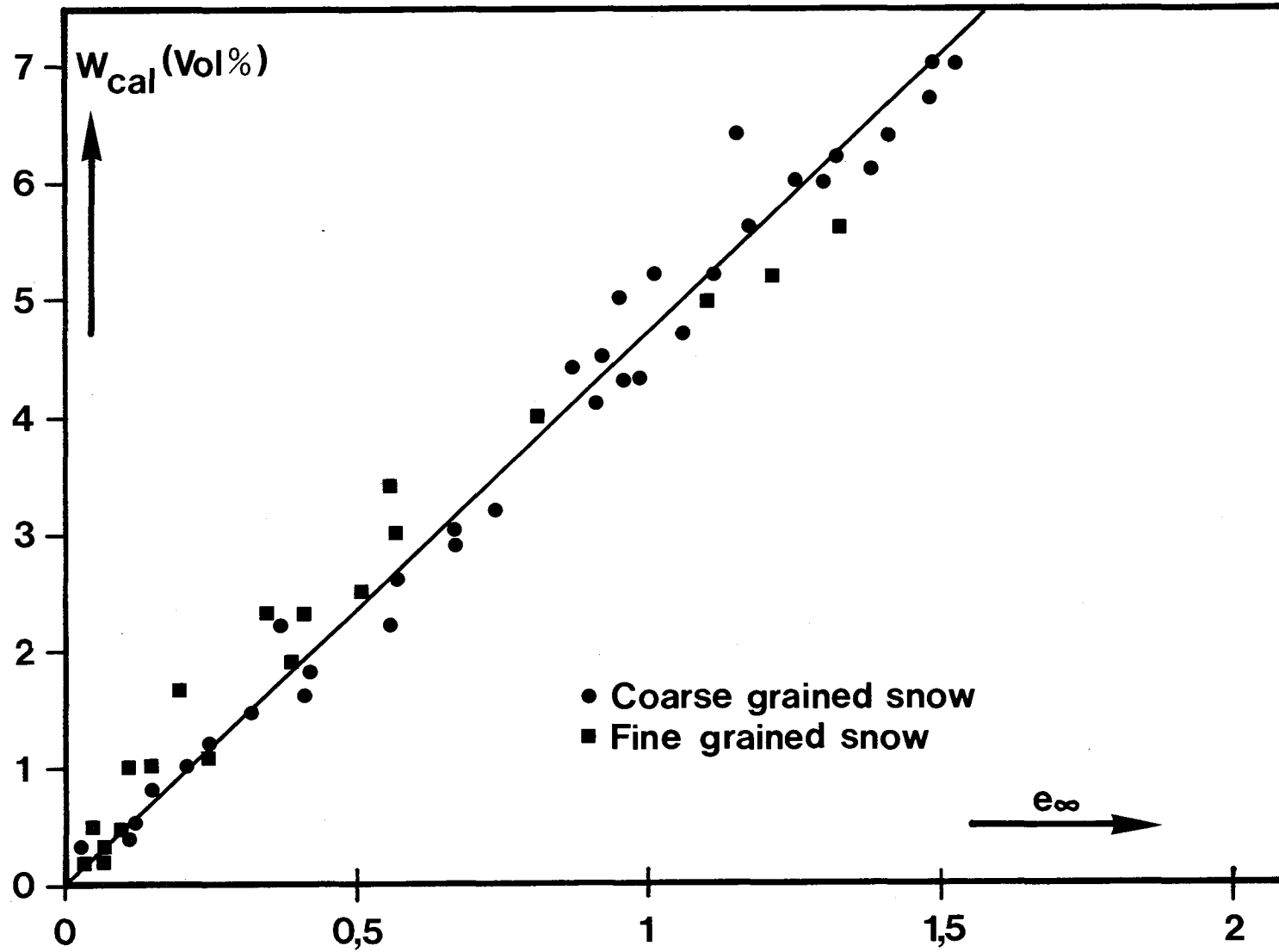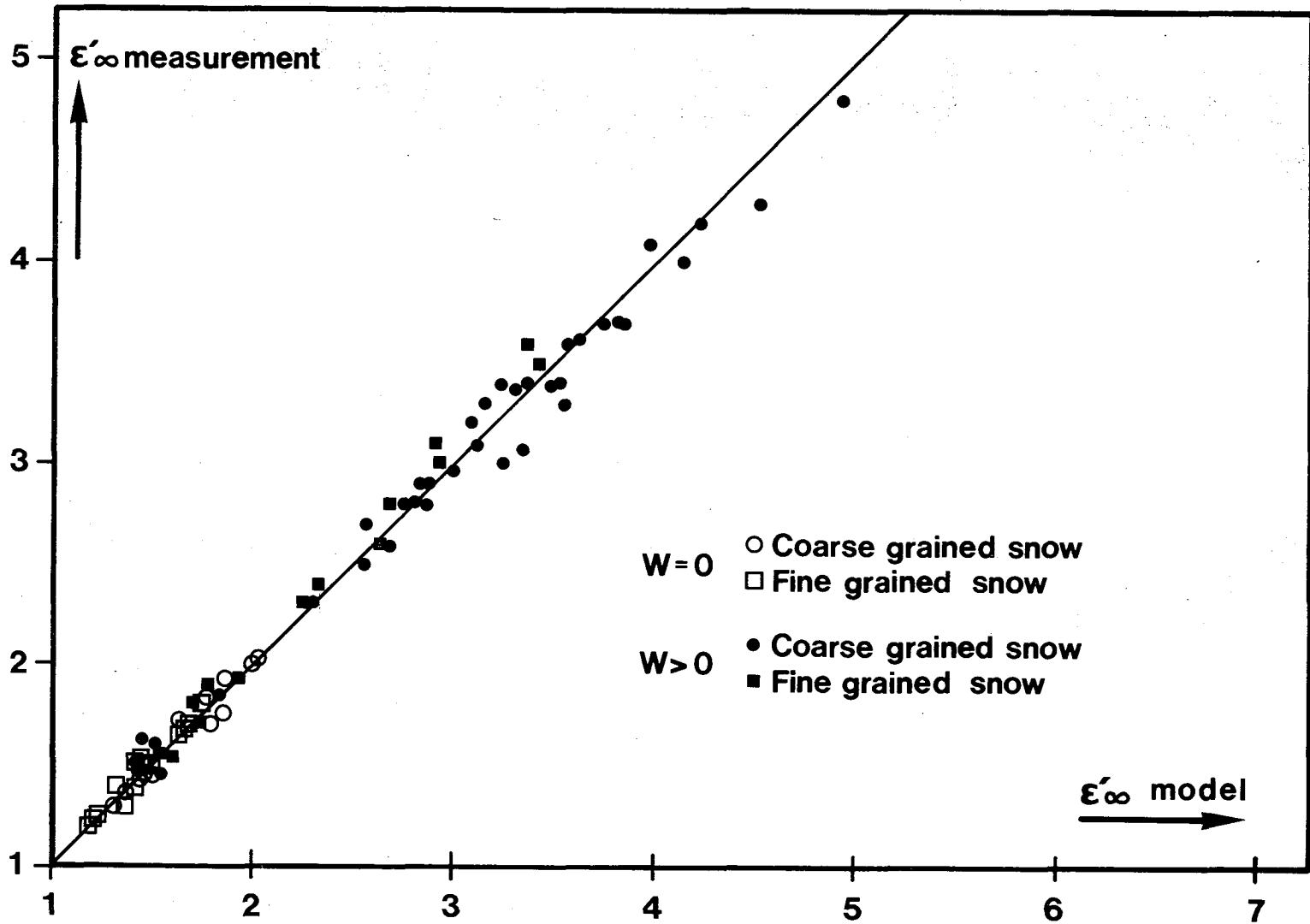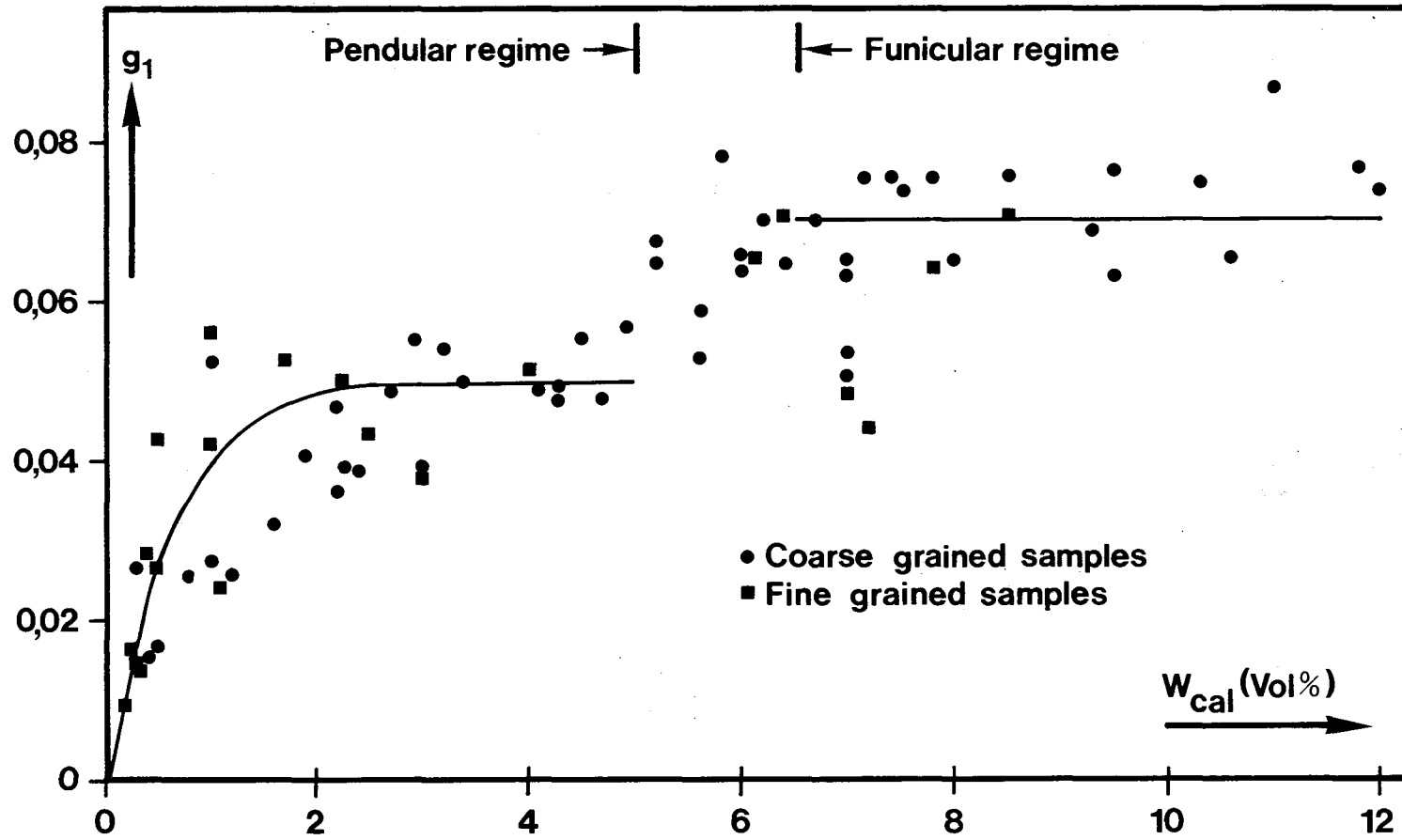
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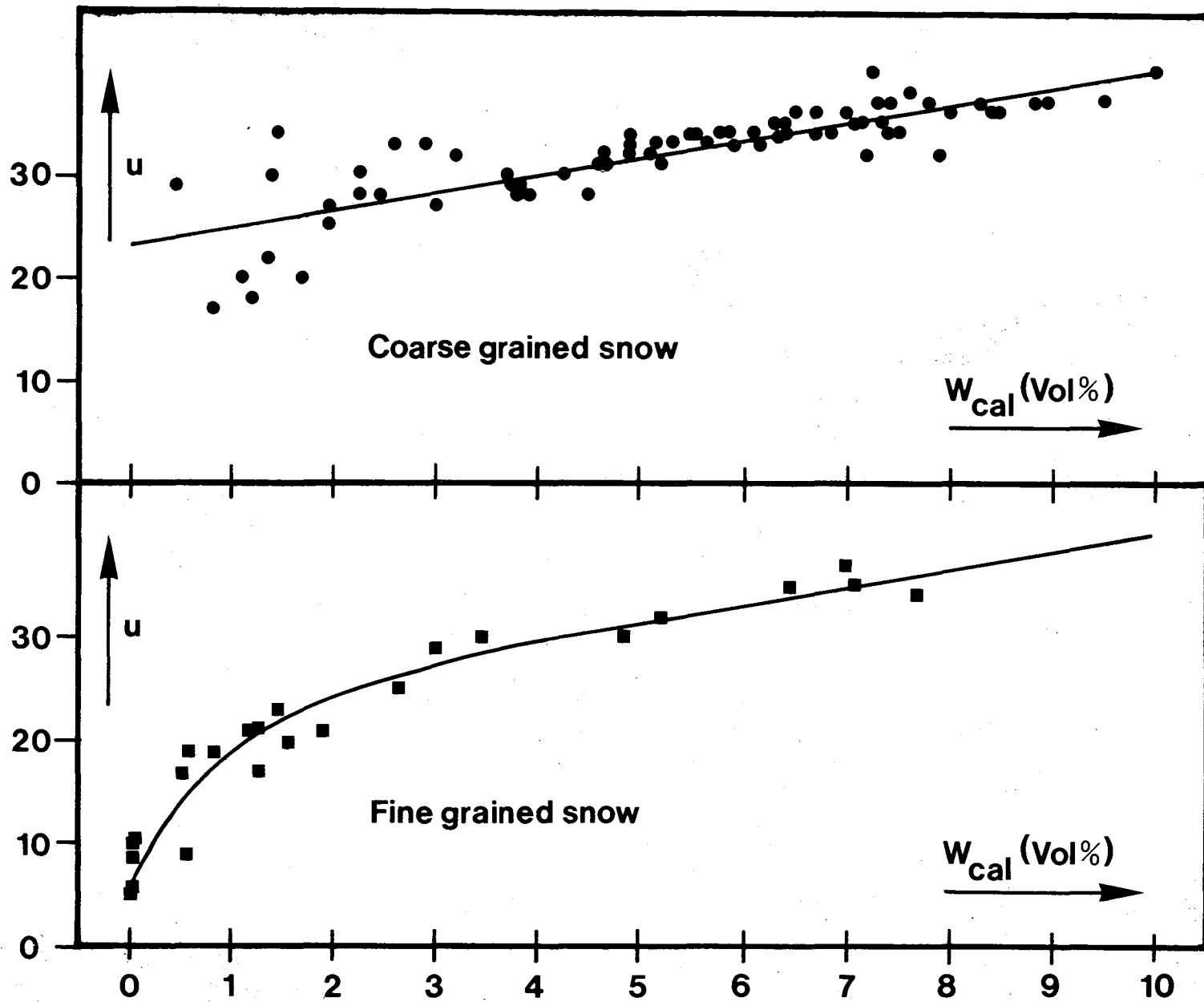
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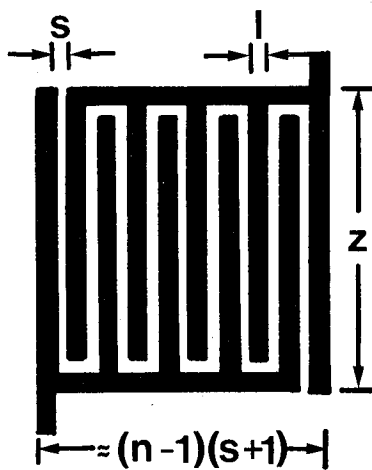
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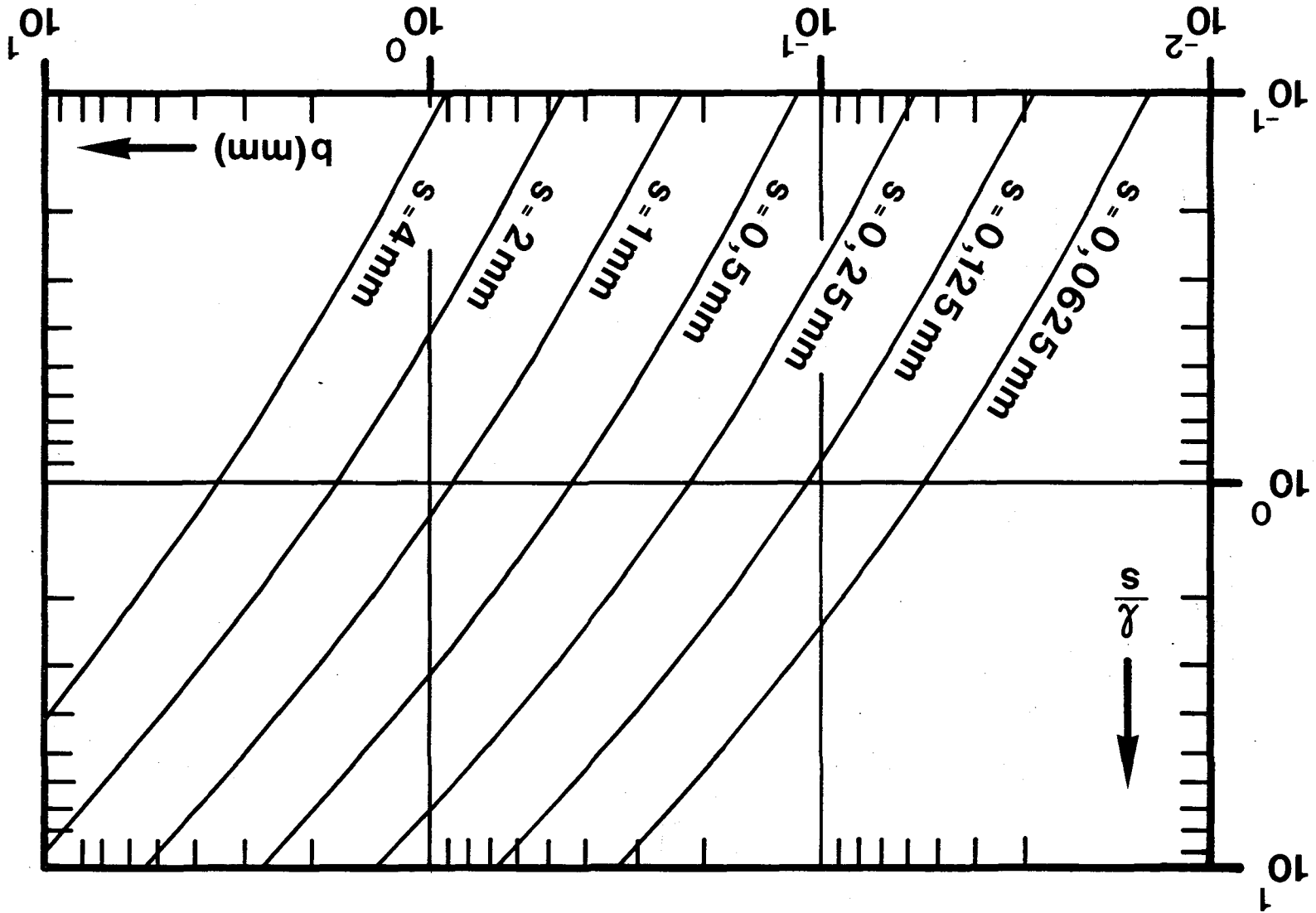
Figure 9

SNOW ELECTROMAGNETIC MEASUREMENTS*

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ABSTRACT

An electromagnetic system is described for measuring the dielectric constant and attenuation of snow samples in the frequency range of 4 to 12 GHz. System components consist of a swept-frequency source, microwave horns, network analyzer, and XY plotter. Procedure in calibrating the effect of wetness on the snow properties is described. Equations are given that express the experimentally determined relation between attenuation per unit length and volume percent wetness at any frequency between 4 and 12 GHz. Permittivity can be calculated from the snow density, attenuation per unit length, and frequency. Some applications of the techniques are described such as runoff forecasting from mountain snowpacks.

INTRODUCTION

The measurement of mountain snowpacks is one part of the NASA programs for remote sensing of earth resources, employing active and passive electromagnetic systems. To plan experiments and to interpret results, one must have values for the permittivity and attenuation of snow for the range of wetness encountered under natural conditions. The frequencies of interest start at a few gigahertz (GHz).

Published information regarding the attenuation of wet snow in the GHz region is extremely limited. Cumming (ref. 1) gives curves relating the loss tangent of snow to liquid-phase water for two densities at the frequency of 9.375 GHz. Sweeny and Colbeck (ref. 2) measured the dielectric constant and loss factor at 6.0 GHz for wet snow, their values of wetness ranging from 0 to 24% by volume. They also performed tests using glass beads as the host medium in place of snow. Linlor (ref. 3) measured the permittivity and attenuation of wet snow between 4 and 12 GHz.

The present paper addresses topics of interest to snow specialists, in the context of the Linlor paper (ref. 3), such as calibration procedures for

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obtaining snow samples having known amounts of liquid-phase water. Previous publications (refs. 4-14) are listed as references without discussion because this is not intended to be a "review paper." However, the data of references 1 and 2 are compared with our results.

SNOW SAMPLES HAVING KNOWN WETNESS

The preparation of samples of snow having known amounts of liquid-phase water is a difficult problem. Sweeny and Colbeck (ref. 2) employed a "capillary pressure system" in which the snow sample was first saturated with ice water, then drained, using a porous plate under a negative pressure gradient. The water content was calculated by taking the difference between the water added and that removed.

Jones, Rango, and Howell (ref. 15) measured the water content in naturally occurring snow rather than preparing samples. They employed a "cold calorimeter" to measure the heat absorbed in freezing the liquid-phase water. Such a measurement does not provide information as to the effect on the dielectric constant and attenuation produced by liquid-phase water when present in snow.

Our approach is based on adding known amounts of water to initially dry snow, mixing thoroughly, and measuring the sample to obtain the permittivity and attenuation in the frequency range of 4 to 12 GHz.

SNOW SAMPLE SIZE

Large samples were chosen so that crystal orientation effects would be averaged, and so that any inhomogeneities (if present after mixing) would also be averaged. A Plexiglas cell was used, having inside dimensions of 39.1 cm × 39.1 cm × 14.5 cm thick. The walls were 0.635 cm (0.25 in.) thick. Snow density was obtained from cell weights, full minus empty.

WATER ADDITION

Tests were made to determine how ice water should be added to dry snow to obtain a reasonably uniform final state. If feasible, the water should be added as "in nature," with downward diffusion producing the uniformly wet sample. For our purposes the rate of adding the ice water had to be sufficiently rapid to achieve at least several samples per day.

An exploratory test was made on a large cube of new, dry, homogeneous snow, approximately 1 m on each side, all of which had been deposited in a single storm. This block was isolated from a snowbank by digging vertical clearance channels, and a metal plate was inserted at the bottom. Ink-colored water was sprayed gently and uniformly on the top, at the rate of approximately 600 cm^3/min on the area of 1 m^2. The spray-water application was stopped when water was observed at the metal plate, and the snow block was examined. Upon removal of the sides of the block, it was found that the water

had formed "veins" vertically down through the block, although it had been sprayed quite uniformly. The final distribution of water was distinctly non-uniform throughout the volume.

To check whether this type of vertical channeling occurs naturally or was produced because of the rate of addition of water, snowpacks were examined by carefully removing snow. Vertical "fingers" were found, originating at layers within the snowpack, having dimensions approximately 0.3 to 0.6 cm in diameter and vertical lengths of 5 to 10 cm. It was not known whether these fingers were produced solely by heat melting (sun and air) or if some unobserved (i.e., unreported) rain had occurred. This phenomenon is identified as a problem for further investigation because the presence of such fingers would obviously affect measurements via either active or passive microwave systems.

On the basis of the above information it was concluded that mechanical mixing of wetted snow samples is necessary to achieve uniformity in the final sample.

EFFECTS OF MIXING

Tests were made to determine whether the mechanical mixing of wetted snow would produce undesirable effects such as melting and major grain structure modification.

Dry snow samples were placed in a mixing chamber and vigorously stirred using manually operated wooden paddles. The force exerted on the snow averaged about 10 lb, acting through a distance of about 3 ft in 1 sec, as the maximum work rate. During 15 min of mixing, the average power was about half the maximum. The effect of mixing was measured by noting the increase in snow temperature, starting at, for example, $-15°$ C, with a room temperature of $-5°$ C. We found that the snow reached room temperature within 10 min or less, and remained at the room temperature despite mixing for an additional 15 min. The conclusion was that the mechanical mixing produced a rise in snow temperature of not more than $1°$ C.

The preceding conclusion is confirmed by calculation of the heat produced by mechanical work. The dry snow sample had a mass of 20,000 gm. The manual work rate (average) of 15 ft-lb/sec is equivalent to 20 W. For the time of 15 min, the energy is 18,310 J, or 4,380 calories. Since the specific heat of snow is about 0.5 calorie/gm/°C, the 20,000 gm of snow would be raised in temperature by $0.44°$ C.

If the dry snow were at $0°$ C, the energy of 4,380 calories would produce melting of 55 gm of snow (heat of fusion being 80 calories/gm), or about 0.3% wetness by weight. For snow of density 0.5 gm/cm^3, this represents a volume wetness of 0.14%.

In some cases the snow was mixed using motor-driven paddles. It is difficult to estimate how much of the motor power was transferred to the snow, but experimentally it was found that temperature of dry snow behaved similarly to the case of the manually stirred snow.

Examination of snow crystals at 8× magnification before and after the mixing did not show any significant change in structure.

Tests were also made to determine the effect of room temperature on the snow during the mixing time. The conclusion was reached that for 1 cm^3 of snow in the laboratory (no sunlight and no appreciable air motion) having temperature of +1° C, approximately 1% volume wetness is produced in 15 min. Obviously, to minimize this type of melting, samples having a large volume to surface ratio should be employed. Our samples (see above) had volumes of the order of 40,000 cm^3.

We estimate that the combined effect of mechanical work and ambient room temperature melting during snow mixing produced an uncertainty of less than 0.5% in the volume wetness.

CALIBRATION PROCEDURE

For calibration tests, snow samples were employed which had a known history and which had never experienced melting. For example, newly fallen snow at the temperature of −10° C was collected and placed in thermally insulated containers. Other collections of snow were placed in freezer chests for about 6 months at −15° C.

In preparation for calibration tests, the snow was brought to essentially 0° C by placing it in a waterproof container and surrounding it with moist snow (i.e., 0° C environment). When the snow was less than 1° away from zero, it was briefly stirred, placed in the test cell, and measured to confirm that it was dry. This test (without measurable attenuation) was performed on all calibration samples before water was added.

Next, a known amount of snow was selected, typically 20,000 gm. Water was chilled to 0° C, and 200 gm added and mixed; this produced 1% wetness by weight. After mixing for about 15 min, some of the snow was placed in the Plexiglas test cell and compacted to ensure uniform density. The snow-filled test cell was examined under a bright light to determine if it looked uniformly gray (transmitted light). The sample was then placed in a thermally insulated foam box and the EM measurements made (see below).

After being tested, the snow in the cell was returned to the table and the full (20,200 gm) supply received additional ice water, so as to encompass the desired wetness range. In some cases, 20% of the dry snow weight was added in the form of ice water to determine whether a single large step was equivalent to the sum of perhaps 10 small steps. Although the total mixing and exposure to ambient temperatures were different, we found that both procedures gave essentially the same results, providing that the ambient temperature was less than 1° away from 0° C.

Each sample was weighed after the EM measurements to provide the average density of the wet snow. The snow crystals were examined under low-power (8×) magnification before and after the EM measurements to determine average grain

diameters. If the mechanical mixing produced any effects, they were not discernible at the low-power magnification, in comparison to natural snow.

INSTRUMENTATION

Phase shift and transmission loss were measured, using two identical Plexiglas containers, one as the reference unit (empty) and the other with the snow sample. The containers had inside dimensions of 39.1 cm by 39.1 cm in the plane perpendicular to the microwave beam, and a depth of 14.5 cm in the direction of the beam. The walls were 0.635 cm (0.25 in.) thick.

Three pairs of conventional microwave horns were used, each horn having nominal gain of 20 dB at its midfrequency point, operating, respectively, in the ranges 4.0 to 6.0, 6.0 to 8.2, 8.2 to 12.0 GHz. Low-loss cables were used. A network analyzer measured the phase shift and transmission for the reference unit and the snow sample, the output being plotted versus frequency on an X-Y plotter that had separate pens for the phase and transmission channels. Thus, a simple subtraction of one curve from its counterpart (i.e., test minus blank) yielded the data for the snow. A conventional sweep generator was used for the microwave source. To verify that the initial supply of snow was adequately dry, a portion was placed in the test cell and measured. No significant attenuation was obtained. This procedure was followed for each snow supply to verify the initial dry state. Also, the absence of appreciable attenuation at any frequency up to 12 GHz shows that "beam scattering" is negligible in the test cell distance of 14.5 cm.

Impedance mismatch effects at the cell interfaces are negligible. For normal incidence on a medium having dielectric constant k_2 from a medium having dielectric constant k_1, the well-known equations are

$$\left(\frac{\sqrt{k_2} - \sqrt{k_1}}{\sqrt{k_2} + \sqrt{k_1}} \right)^2 = \text{Reflection power}$$

$$\frac{4\sqrt{k_1 k_2}}{\left(\sqrt{k_2} + \sqrt{k_1} \right)^2} = \text{Transmission power}$$

The Plexiglas walls have a dielectric constant of about 2.2. Dry snow of density 0.4 gm/cm^3 has a dielectric constant of 1.8 in the GHz region up to 12 GHz. Wet snow of 6% volume wetness and density of 0.5 gm/cm^3 has a dielectric constant of about 2.8 at 8 GHz. For these values, the greatest loss in transmission would occur if a single step in dielectric constant is assumed, namely, from 1 to 2.8. For such a step, the power transmission is 0.937, representing a loss of 0.28 dB. For interfaces having dielectric constants of 2.2 and 2.8, respectively, the power transmission is 0.996, representing a loss of 0.02 dB. The aggregate effect of the interfaces for the various

combinations is less than 1 dB loss; in comparison, the attenuation for wet snow of 6% volume wetness at 8 GHz is 21 dB in 14.5 cm.

Multiple reflections within the test cell under low loss conditions can be identified by the response to swept frequency. When the transmission loss through the 14.5 cm distance exceeds about 5 dB, the effect of multiple reflections becomes small. For most of our measurements, the attenuation in one passage of the test cell exceeded 10 dB at the 8 to 12 GHz range, so multiple reflections represented unimportant perturbations.

MEASUREMENTS

Many exploratory runs were necessary to determine appropriate operating procedure. Our final calibration curves are shown in figures 1-3. The original information was obtained as continuous X-Y curves. A dual-trace (X-YY) plotter gave the frequency as the X axis, the transmitted signal and the phase as two independent ordinates. Successive tracings of the curves were coincident to within the width of the tracing pen. Smooth curves were fitted to the tracings to eliminate obvious (and small) irregularities caused by interfaces. Data "points" were selected at each integral frequency value and are depicted in the figures as X marks.

Figure 1 shows the permittivity for dry snow in the frequency range of 8.2 to 12.0 GHz. The attenuation for the sample, which was 14.5 cm thick in the direction of the beam, was less than 1 dB; it is not plotted.

In figure 2, the permittivity and decibels per centimeter are plotted versus frequency in the range of 8.2 to 12.0 GHz for a snow sample having 2.51 volume percent wetness and density of 0.442 gm/cm^3. Also plotted are the loss factor ξ'' and tan δ, which are obtainable from the permittivity and attenuation per centimeter.

In figure 3, for the frequency range of 4.0 to 12.0 GHz, the permittivity, decibels per centimeter, loss factor, and tan δ are shown for a snow sample having 6.24 volume percent and density of 0.558 gm/cm^3.

SNOW WETNESS EQUATIONS

From the measured phase shift (in deg), the permittivity can be calculated (see "symbols"):

$$\xi' = [1 + \phi/12D\nu]^2 \tag{1}$$

The loss factor is obtained from the attenuation and permittivity as follows:

$$\xi'' = \xi' \tan \delta \tag{2}$$

$$\tan^2 \delta + 1 = \left(\left[\frac{dB/cm}{1.286 \nu \sqrt{\xi'}} \right]^2 + 1 \right)^2 \tag{3}$$

If $\tan \delta \ll 1$,

$$\tan \delta = \frac{1.0994}{\nu \sqrt{\xi'}} \frac{dB}{cm} \tag{4}$$

Equations (1) through (4) are well known in electromagnetic theory. Curves for various snow samples show the dependence of ξ' and ξ'' on frequency. For convenience, curves are also included for decibels per centimeter and $\tan \delta$, although these evidently are related to ξ' and ξ'' by the preceding equations.

The following three equations are essentially empirical relations. Theoretical considerations served as a guide, but the equations are justified mainly on the basis that they are in agreement with measured values. Analysis of the calibration curves, test runs, and checks for internal consistency resulted in the following relations:

$$\frac{dB}{cm} = W_v (0.045[\nu - 4] + 0.066[1 + a]) \tag{5}$$

$$\xi'_{calc} = 1 + 2\rho + bW_v^{3/2} \tag{6}$$

and

$$b = 5.87 \times 10^{-2} - 3.10 \times 10^{-4} (\nu - 4)^2 \tag{7}$$

DISCUSSION OF SNOW WETNESS EQUATIONS

Because of its importance, the reliability of equations (5), (6), and (7) is discussed first.

For any snow sample (wetness known by calibration or unknown), the attenuation and phase shift were measured simultaneously by the instrumentation. From the phase shift, the permittivity can be obtained using equation (1). This is the measured value and involves no assumptions.

A calculated value of the permittivity is obtained from equation (6). The data required are the density, wetness, and the value of the parameter b defined in equation (7) as a function of frequency. The wetness can be obtained from calibration data or alternatively from equation (5) with measured attenuation.

Thus, for any snow sample, the measured and calculated values of permittivity can be compared by plotting one value along the X axis and the other along the Y axis. The distribution of such points relative to the 45° line shows how well the two sets of values agree. This is done in figure 4, for 17 samples; some samples were dry, the rest had wetnesses that were obtained from calibration data or from measured attenuation. Various snow densities and grain sizes were included in the samples and various frequencies, in the 4 to 12 GHz range, were used. The excellent agreement of the measured and calculated values of the permittivity provides assurance that equations (5), (6), and (7) are trustworthy.

Equation (5), which gives a semiempirical relation, is now considered. In the frequency range of 4 to 12 GHz, the attenuation per unit length is proportional to the volume-percent wetness and is a linear function of frequency. For convenient reference, equation (5) is plotted in figure 5 for selected frequencies and in figure 6 for selected values of wetness. The value of the constant a is taken to be zero, unless stated otherwise. Equation (5) is based on the data shown in figures 1-3 and is additionally checked by the permittivity comparisons discussed above.

Equation (6), as pointed out earlier, permits a calculation of the permittivity from measurements of the density, attenuation, and frequency. For dry snow, the first two terms in equation (6) yield a value for permittivity that is in close agreement (within a few percent) with the measured permittivity from equation (1), obtained from phase shift in the GHz region. The upper limit to density to be used in equation (1) is 0.6 gm/cm^3. Because the sample dimensions used in these measurements are so large, any crystal orientation effects are presumed to be insignificant because of the randomization produced in the sampling processes. Approximately 50 samples were measured. Our permittivity measurements are about 6% higher than those of Cumming (ref. 1) for snow densities of about 0.2 gm/cm^3, and are in excellent agreement with his densities near 0.5 gm/cm^3.

The third term in equation (6) must be included for the case of wet snow. The factor b is obtained from equation (7) since the frequency is known. The volume-percent wetness for equation (6) can be obtained from calibration data or from equation (5), using measured attenuation per centimeter.

An interesting check on the validity of equations (4) through (7) is obtained by comparing predictions derived from them with the measurements of Cumming (ref. 1). His stated values of wetness, density, and frequency were substituted into the equations of this paper; the results and Cumming's curves are plotted in figure 7. Good agreement is obtained for the density of 0.38 gm/cm^3. For the density of 0.76 gm/cm^3, agreement is evident for the lower portions of the curves, but further investigation is necessary to identify the disparity at the upper portions of the curves.

A comparison of the data of Sweeny and Colbeck (ref. 2) and the equations of this paper is shown in figure 8 for the dielectric constant (or permittivity) and in figure 9 for the loss factor. Because the data of reference 2 include the effect of snow density, three curves are shown in figure 8 for the snow densities, respectively, of 0.5, 0.6, and 0.7 gm/cm^3. Some of the data

points of Sweeny and Colbeck agree with the curve for 0.7 gm/cm^3, but most are higher than this curve. According to the equations in this paper, the snow density plays a minor role in regard to the loss factor, and so only the curve for density 0.6 gm/cm^3 is shown in figure 9. The curves for 0.5 and 0.7 gm/cm^3 would essentially coincide with the curve shown. The loss factors of Sweeny and Colbeck are larger than values obtained from our equations.

SNOW DRAINAGE CHARACTERISTICS

The instrumentation measures the wetness of the snow sample without disturbing it in any way because the intensity of the beam is too low to produce measurable melting. The volume-percent wetness versus drain time for three samples is shown in figure 10. Sample A consisted of small grains (0.1 to 0.3 mm diam) and had an initial wetness of about 9.6 volume percent. The wetness was remeasured after drain times totaling 2, 6, and 20 hr. The sample was kept in a 0° C environment at all times. Sample C had large grains (1 mm diam) and an initial wetness of about 9.4 volume percent. Its wetness decreased more rapidly than that of sample A.

Sample B had large grains (1 mm diam) and was treated as follows: (1) the snow was completely immersed in 0° C water for about 10 min then removed and allowed to drain; (2) after 3 hr, it was placed in the test cell and measured; and (3) a second measurement was made at the 18-hr drain time.

Similar data are presented in figure 11. The curves A and C are the same as those shown in figure 10; they are repeated in figure 11 for reference. Curve D shows the drainage characteristic for a sample that initially had a wetness of 6.2 volume percent. Curve E shows that a snow sample having only 4.0 volume-percent wetness did not drain at all, within the measurement capability of the instrumentation, up to a time of 9 hr.

The water speed in "ripe snow" (i.e., grain size approximately 1 mm diam) was measured with sample C. After the drain time of 20 hr, a 3-cm-deep region was present at the bottom of the snow cell (where it did not produce beam attenuation); the region had a wetness of about 33 volume percent and the appearance of "slush." The snow cell was inverted to bring the slush layer above the beam axis; this caused the accumulated water to drain down through a distance of about 21 cm to the beam region. Based on the time for maximum beam attenuation to occur, the measured speed of travel of the pulse of water for the stated conditions was about 20 cm/min. Further measurements of water passage under a variety of conditions are planned.

APPLICATIONS OF TECHNIQUE

The techniques discussed here may have many applications. We plan to measure the permittivity and attenuation of a variety of snow samples that differ in density, grain size, wetness, age, etc. Those measurements will be made to confirm and to extend the data base. After an adequate number of samples is available, the technique can be used for "snow truth;" that is, from

the electrical measurements of samples one can determine the physical characteristics of the snowpack.

The equations presented here permit the rapid determination of snow wetness in the field by measurement of the attenuation of samples. Since phase measurements are not required, the instrumentation can be quite simple and inexpensive, such as a system having microwave horns, an oscillator that can be swept in frequency, and a detector. Details will be published in a later paper.

An important application is the prediction of the time and rate of snow melt for reservoir management in the western United States. Sets of microwave horns located at snow courses can provide data on a daily basis to assess the wetness of the snowpack. Several pairs of horns located at different heights above Earth and immersed in the snow can indicate the liquid-phase content. The instrumentation may be capable of predicting when the snowpack will reach its maximum water-holding state and thus be primed for runoff. Another potential application, although somewhat speculative at present, is the measurement of snowpack wetness for avalanche warning.

The techniques can be employed to measure water speed in snow, earth, and similar materials by observing the changes in attenuation caused by wetness. A version of the technique may permit the _in situ_ measurement of soil moisture on a repetitive basis, to provide index-station data.

SUMMARY OF RESULTS AND CONCLUSIONS

Samples of natural snow have been calibrated by the addition of known amounts of water, thoroughly mixed, and measured in large-volume cells. The phase shift and the attenuation were measured in the frequency range of 4 to 12 GHz. Density was obtained for each snow sample.

Empirical equations were developed from the data, giving the relation between attenuation per unit length and volume-percent wetness. Additional equations were developed for the calculation of permittivity from snow density, attenuation per unit length, and frequency. A comparison showed excellent agreement between the permittivity obtained from phase-shift measurement and that calculated from the equations.

The water passage in snow having a typical grain diameter of 1 mm was measured and found to be about 20 cm/min. Curves are given to show the dependence of snow wetness on drain time for a variety of samples.

It is concluded that "ripe snow" (grain diameter of about 1 mm) can hold about 0.04 g of liquid-phase water per cubic centimeter of snow, equivalent to about 8 weight percent for snow whose density is 0.5 gm/cm^3. This represents the 24-hr drain time state, based on homogeneous samples. The presence of layers, ice lenses, and similar impediments to the flow of water may produce increased snow wetness.

SYMBOLS

ν	frequency, GHz, equal to 10^9 Hz
ξ'	permittivity, real part of dielectric constant relative to vacuum
ξ''	loss factor, imaginary part of dielectric constant relative to vacuum
tan δ	ratio of imaginary/real parts of dielectric constant
ϕ	phase shift, deg
ρ	density, gm/cm^3
D	thickness of sample, cm
a	average snow grain diameter (in mm) for the range of 0.1 to 1.0 mm
b	proportionality constant
W_V	wetness in volume percent, grams of water per cubic centimeter multiplied by 100
W_W	wetness in weight percent, grams of water per gram of snow multiplied by 100

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FIGURE CAPTIONS

FIG. 1. Calibration curve for dry snow.

FIG. 2. Calibration curves for wet snow (2.51 volume percent).

FIG. 3. Calibration curves for wet snow (6.24 volume percent).

FIG. 4. Comparison of measured and calculated permittivities.

FIG. 5. Variation of attenuation with snow wetness at selected frequencies.

FIG. 6. Variation of attenuation with frequency at selected snow wetnesses.

FIG. 7. Comparison of Cumming's results (tan δ vs snow wetness) with corresponding results of present paper.

FIG. 8. Comparison of Sweeny and Colbeck's results, permittivity vs wetness, with corresponding results of present paper.

FIG. 9. Comparison of Sweeny and Colbeck's results, loss factor vs wetness, with corresponding results of present paper.

FIG. 10. Variation of snow wetness with drain time for initially saturated snow.

FIG. 11. Variation of snow wetness with drain time for different degrees of initial wetness.
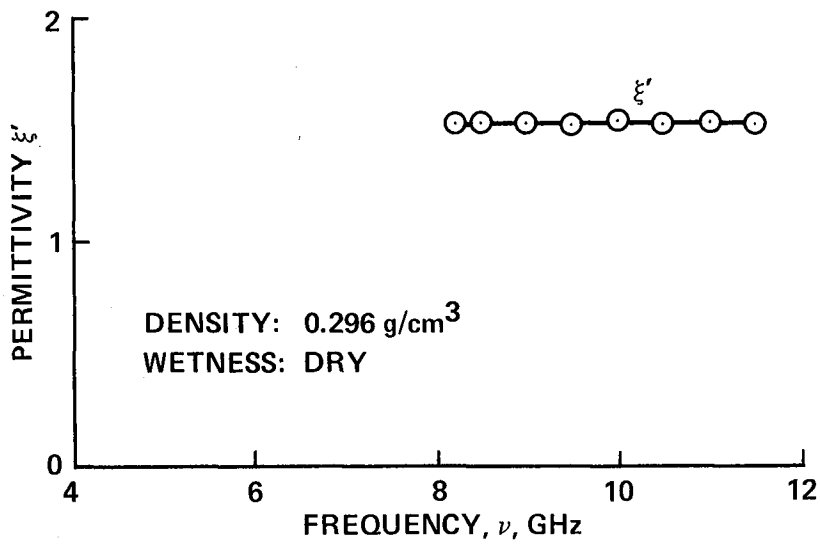
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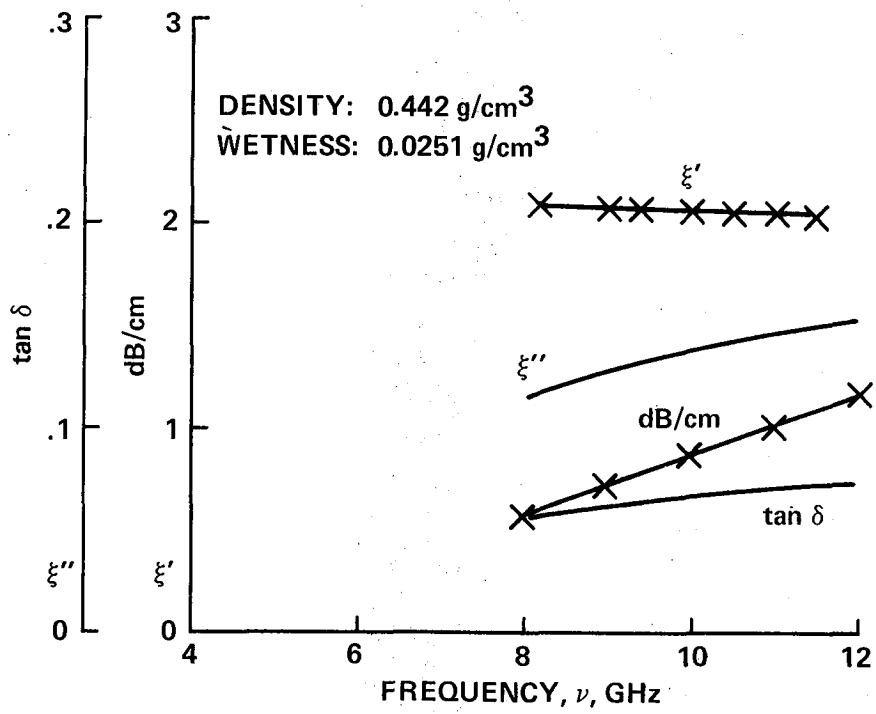
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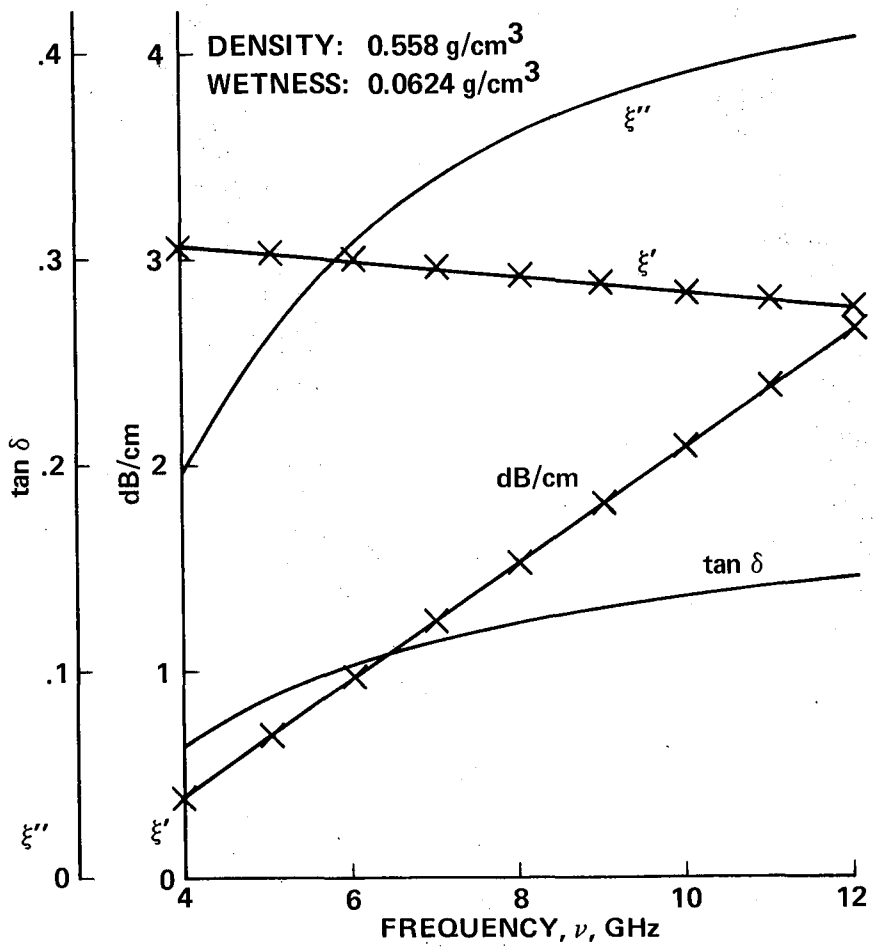
Figure 3

Figure 4

Figure 5

Figure 6

C: CUMMING, FIG. 5
   J. APPL. PHYSICS, 23, 768 (1952)
L: LINLOR, THEORETICAL CURVE.

9.375 GHz

$\rho = 0.76$ g/cm^3

$\rho = 0.38$ g/cm^3

tan δ

SNOW WETNESS, W_w, weight percent

Figure 7

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FIG. 8

Figure 8

Points shown are from Sweeny & Colbeck, 6 GHz

Curves: Linlor "Snow Equations"
For density values of 0.5 to 0.7 grams/cm^3,
the curves coincide on scale shown.

LOSS FACTOR

WETNESS (W$_v$) IN VOLUME PERCENT

FIG. 9

Figure 9

Figure 10

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Figure 11

117
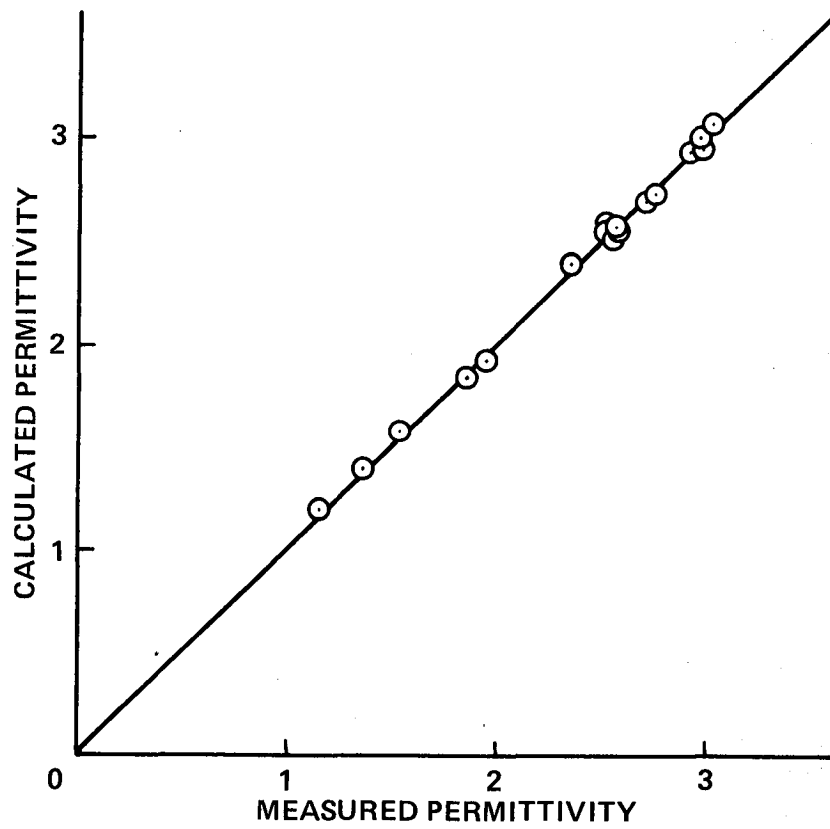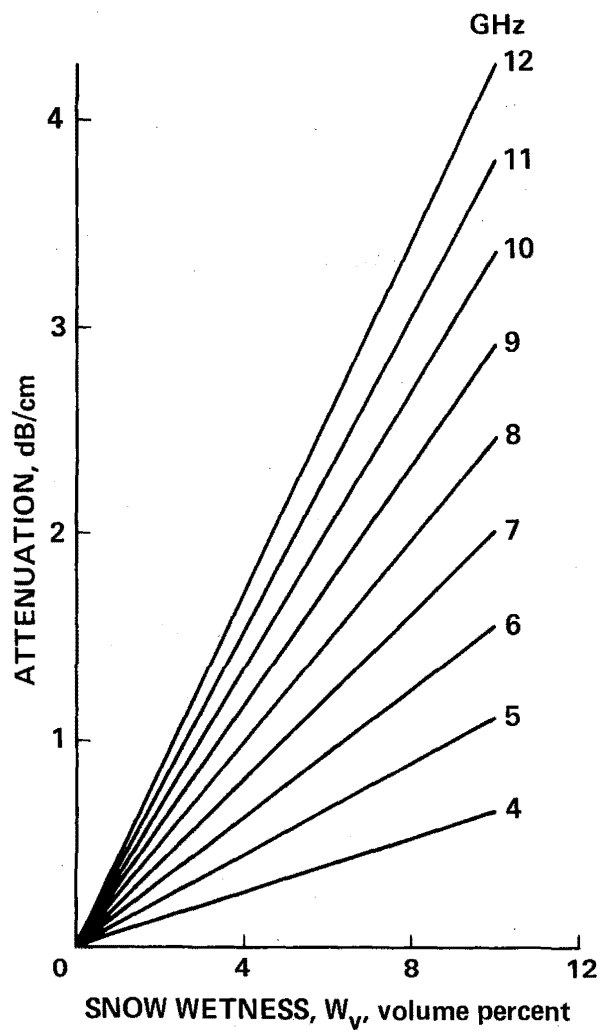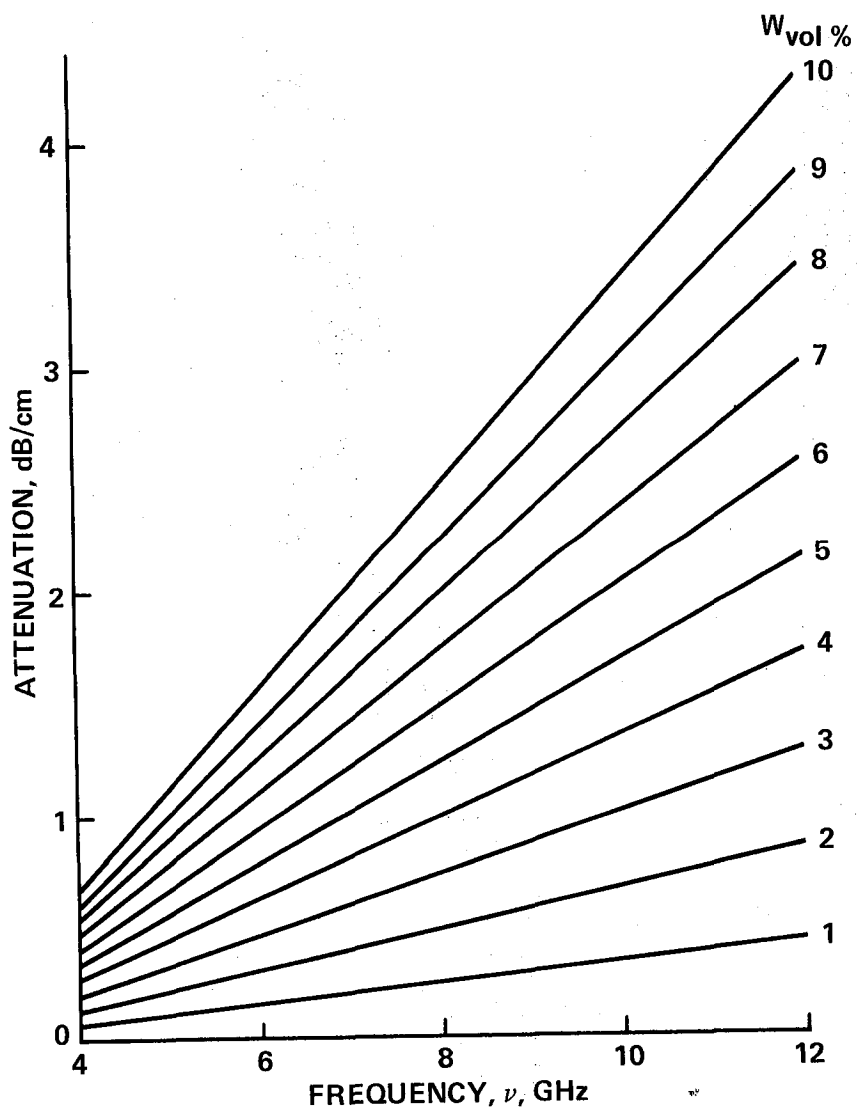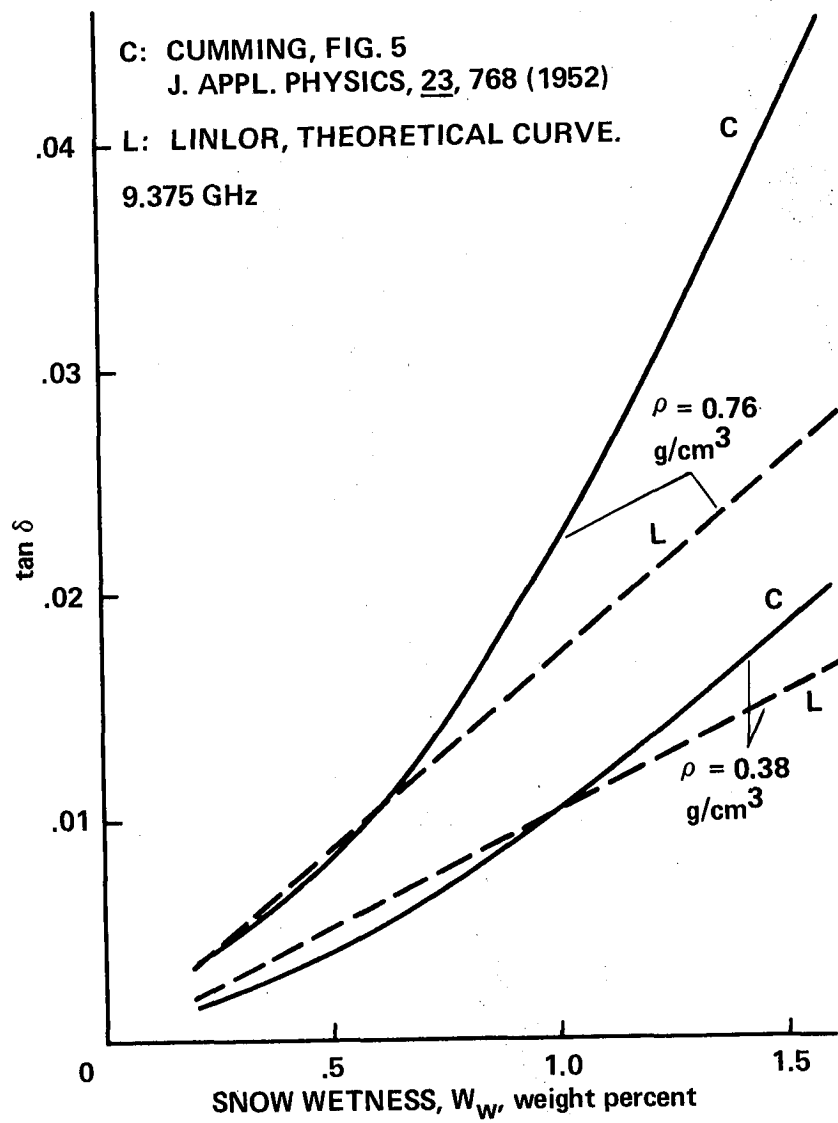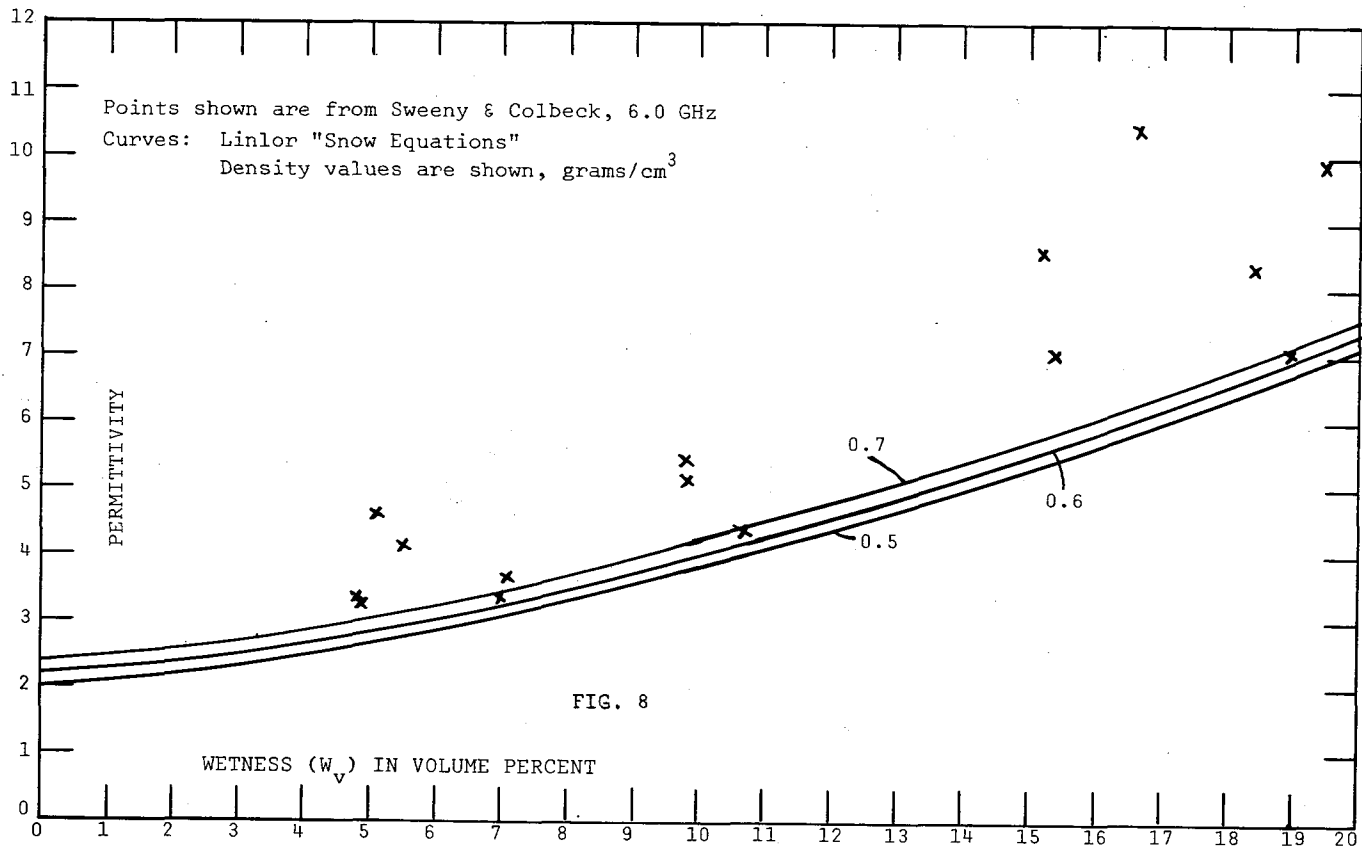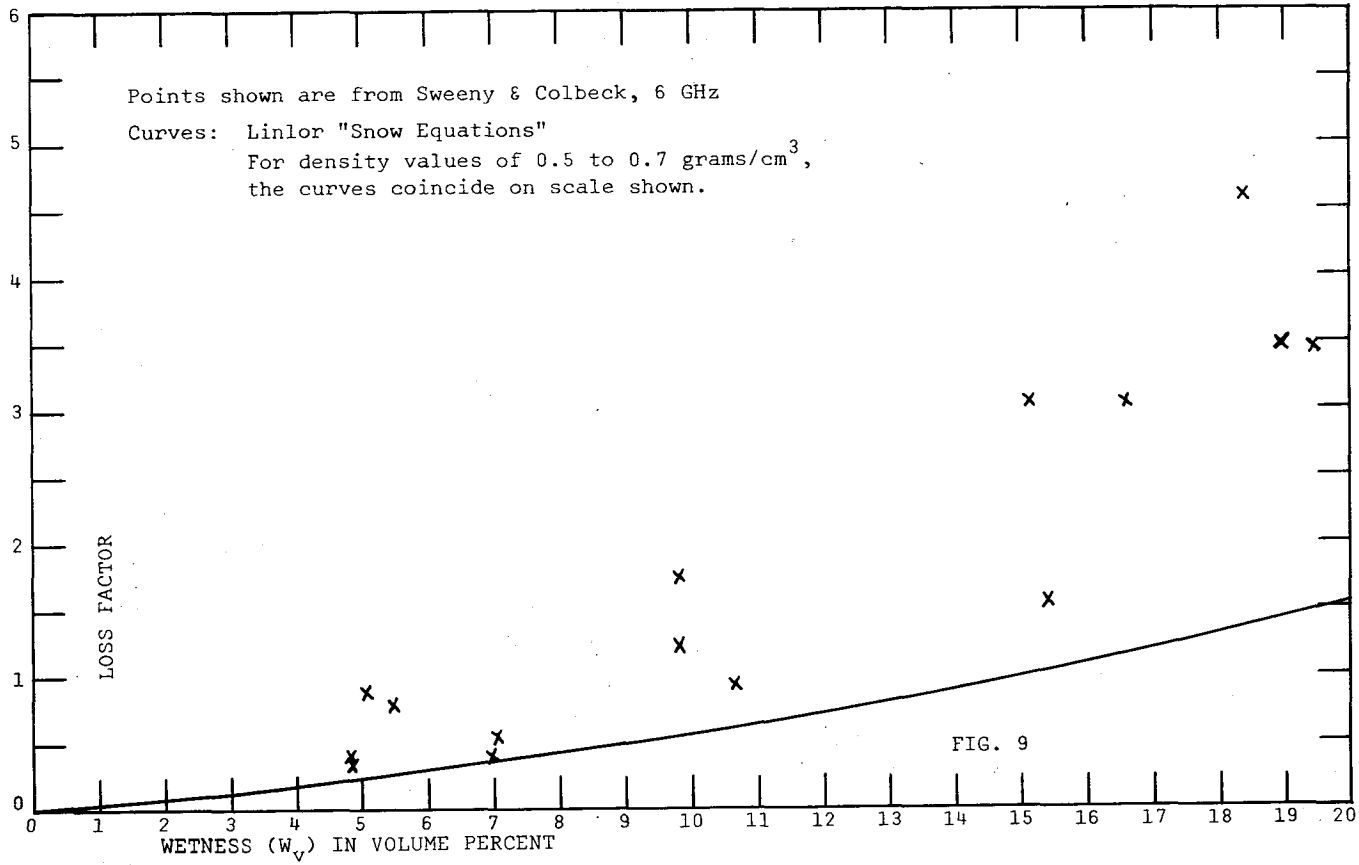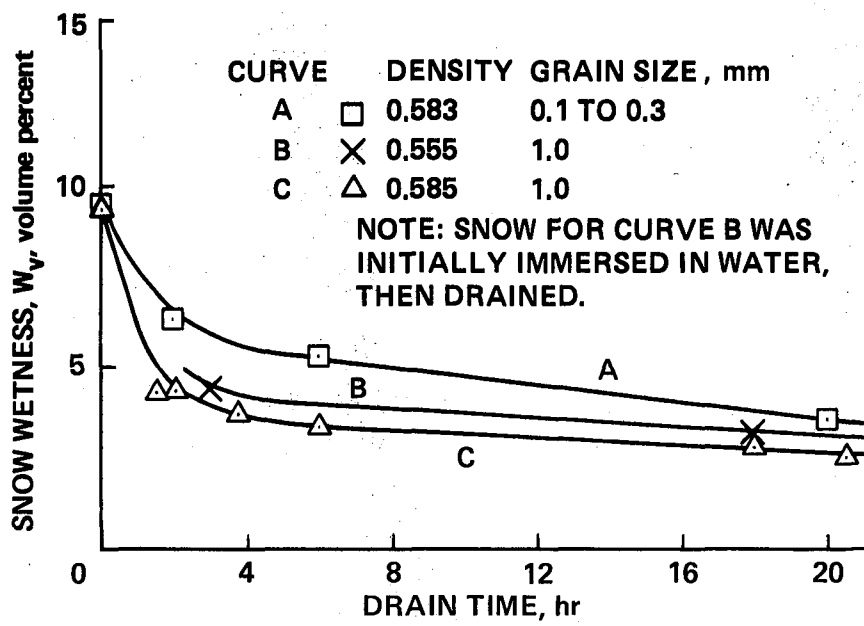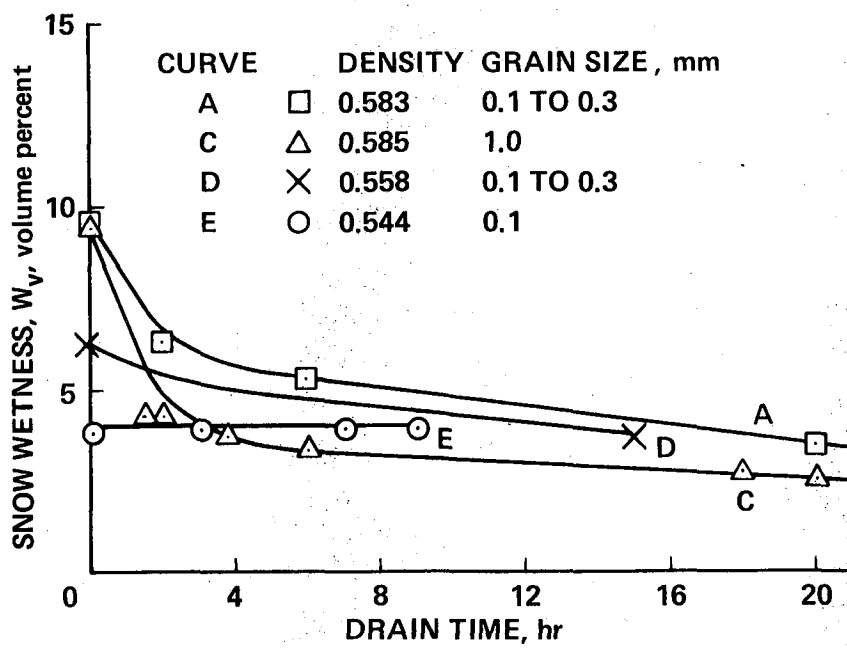
ACTIVE MICROWAVE WATER EQUIVALENCE

MEASUREMENTS*

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Fort Collins, CO

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ABSTRACT

Measurements of water equivalence using an active FM-CW microwave system have been conducted over the past three years at various sites in Colorado, Wyoming and California. The measurement method will be described. Measurements of water equivalence and stratigraphy will be compared with ground truth. A comparison of microwave, federal sampler and snow pillow measurements at three sites in Colorado will be described.

INTRODUCTION

Snowpack in mountain watersheds in the western United States accounts for 65% to 80% of the total available water supply. Prediction of annual water storage in snowpack is, therefore, important in estimating total annual runoff. The U.S.D.A. Soil Conservation Service (ref. 1) and the California State Department of Water Resources (ref. 2) have been coordinating a cooperative snow survey and water supply forecasting program for more than 50 years to serve the needs of a variety of water users. Traditionally, manual measurements of some 2000 snowcourses situated in watersheds throughout the western United States have provided information on snow depth and water content once a month from January to June. These measurements serve as input data to snowmelt runoff forecasting models.

More recently, the SCS has installed over 500 sites to collect hydrometeorological data from remote sites throughout the mountains of the West. Snowpack water equivalence, total precipitation and ambient air temperature are collected twice daily and transmitted to a data collection and processing center via a meteor burst communication system known as SNOTEL (ref. 3).

Each remote site has a snow pillow to measure snowpack water equivalence and a total precipitation gauge. In most cases, the site is situated close to an existing snow course so that the SNOTEL data can be intercompared with

manual measurements made with a federal sampler, to determine any systematic difference which could affect the snow runoff model. A monthly manual measurement is made around the periphery of the snow pillow as an additional performance check on the SNOTEL system.

Recently, an electromagnetic technique has been developed which is capable of probing the snowpack to determine stratigraphic layering and snowpack water equivalence (ref. 4). The technique involves the transmission of electromagnetic radiation through the pack and observing the reflected coherent backscatter from the snow surface, the stratigraphic layers within the snow and the ground surface. The amplitude of the reflected signal is displayed on a spectrum analyzer as a function of frequency. The frequency is, in turn, proportional to the snowpack depth. The detailed analysis of the system, known as an FM-CW radar, is given in reference 5.

This paper describes the intercomparison of the federal sampler, the snow pillow and the radar measurements of snowpack water equivalence at three different SNOTEL sites in the Colorado Rocky Mountains having a range of snow depth from 140 cm to 385 cm.

MEASUREMENT PROCEDURE

Comparison between the radar system and the snow pillow was made by suspending the radar antennas over the pillow (Figure 1) and taking measurements at three selected points above the pillow system. The average water equivalence given by these measurements was compared with the pillow manometer pressure. Finally, the federal sampler was calibrated by comparing its measurements with a glacier sampler measurement taken in a pit which was dug at the site. This last comparison is necessary because the federal sampler is known to over-estimate the actual water content by 6% to 12% depending on the average snow density (ref. 6). Five federal sampler measurements were averaged in this comparison.

Measurements were made at three SNOTEL sites in northern Colorado-- Willow Creek, Columbine, and Tower. The water equivalence at these sites was 39 cm, 65 cm, and 130 cm, respectively. The Columbine site is being used as an experimental site where various types of pillow and pillow configurations are being examined. Three different pillows at this site, all with approximately the same area, were used for this experiment. At the Tower site, a direct comparison of the glacier sampler, the federal sampler and the FM-CW radar system was made in addition to the pillow comparison described above. Figure 2 shows the stratigraphy, type of snow and temperature profile at all three sites. The snowpack classification used is described in reference 7. Figures 3, 4, and 5 show examples of the radar response of the snowpack at Willow Creek, Columbine, and Tower respectively.

COMPARISON OF RESULTS

The measurement results are given in Table 1. Columns 1, 3, 5 and 6 are the water equivalence measurements using the glacier sampler, federal sampler

(corrected), snow pillow, and radar system, respectively. The accuracy associated with a single measurement of the radar system has been determined to be ± 5% (ref. 4). Since 3 measurements over each pillow were averaged, the accuracy of the FM-CW measurement is ± 2.8%. The accuracy of a single federal sampler measurement is more complex since it is a combination of random and systematic errors. Assuming that the systematic error is accounted for in the comparison with the glacier sampler, the random error for a single measurement is ± 5%. The average of eight measurements around the pillow reduces this error to ± 1.8%. Snow pillow manometer errors are assumed to be ± 5%.

Columns 7, 8, and 9 show the percent deviations for the federal sampler vs FM-CW, and pillow vs FM-CW measurement respectively. The comparisons between the pillow and federal sampler and the pillow and FM-CW radar show a definite trend in which the pillow tends to overestimate the water content by an average of 8.5%. The deviations between the corrected federal sampler and the FM-CW radar, however, are statistically insignificant.

CONCLUSIONS

The measurement comparison supports previous measurements of water equivalence with the FM-CW radar (ref. 4) and demonstrates the potential of the system as an operational measurement device. However, additional work is needed in both instrumentation and signal analysis to bring the system to operational capability.

The measurements also support previous reports that the uncorrected federal sampler and pillow measurements tend to overestimate the snowpack water equivalence by about the same amount (ref. 1).

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TABLE 1

Water Equivalence Measurement Comparison

Snotel Site	Glacier Sampler (cm)	Federal Sampler Uncorrected (cm)	Federal Sampler Corrected (cm)	Average Depth (cm)	Pillow Manometer Corrected (cm)	FM-CW Radar (cm)	Percent Deviation Fed. Sampler and Pillow	Percent Deviation Fed. Sampler and FM-CW	Percent Deviation Pillow and FM-CW
Willow Creek									
a) Pit	39.3	41.9	39.3	142.2	--				
b) Pillow		39.9	37.4	142.7	35.3	37.1	-5.6	0.8	-4.8
Columbine									
a) Pit	64.8*	72.4	67.5	200.7	--			--	
b) Cal. Pillow		75.2	69.3	207.6	76.5	64.3	9.4	4.4	7.1
c) Hypalon		72.4	66.8	205.7	74	64.2	9.7	3.9	13.2
d) SNOTEL 76		73.2	67.5	199.4	80.3	69.1	15.9	-2.4	13.9
Tower									
a) Pit	130.6	142.2	130.6	383.5	--	127.1		2.7	
b) Pillow		139.7	128.2	384.8	129.7	123.0	1.1	4.1	5.2

* Because of the differences in depth between the pit and the federal sampler, average comparisons are based on density rather than water equivalence.

Figure 1 Radar Antennas Suspended over the Snow Pillow at the
          Columbine Site

Figure 2 Caption on Figure

Figure 3 Radar Response at Willow Creek, Scale: 1 KHz=30 cm.
          Reflections between the snow surface and the ground are due to
          stratigraphy in the pack

Figure 4 Radar Response at Columbine
          Scale: 1 KHz=30 cm.

Figure 5 Radar Response at Tower
          Scale: 1KHz=30 cm.
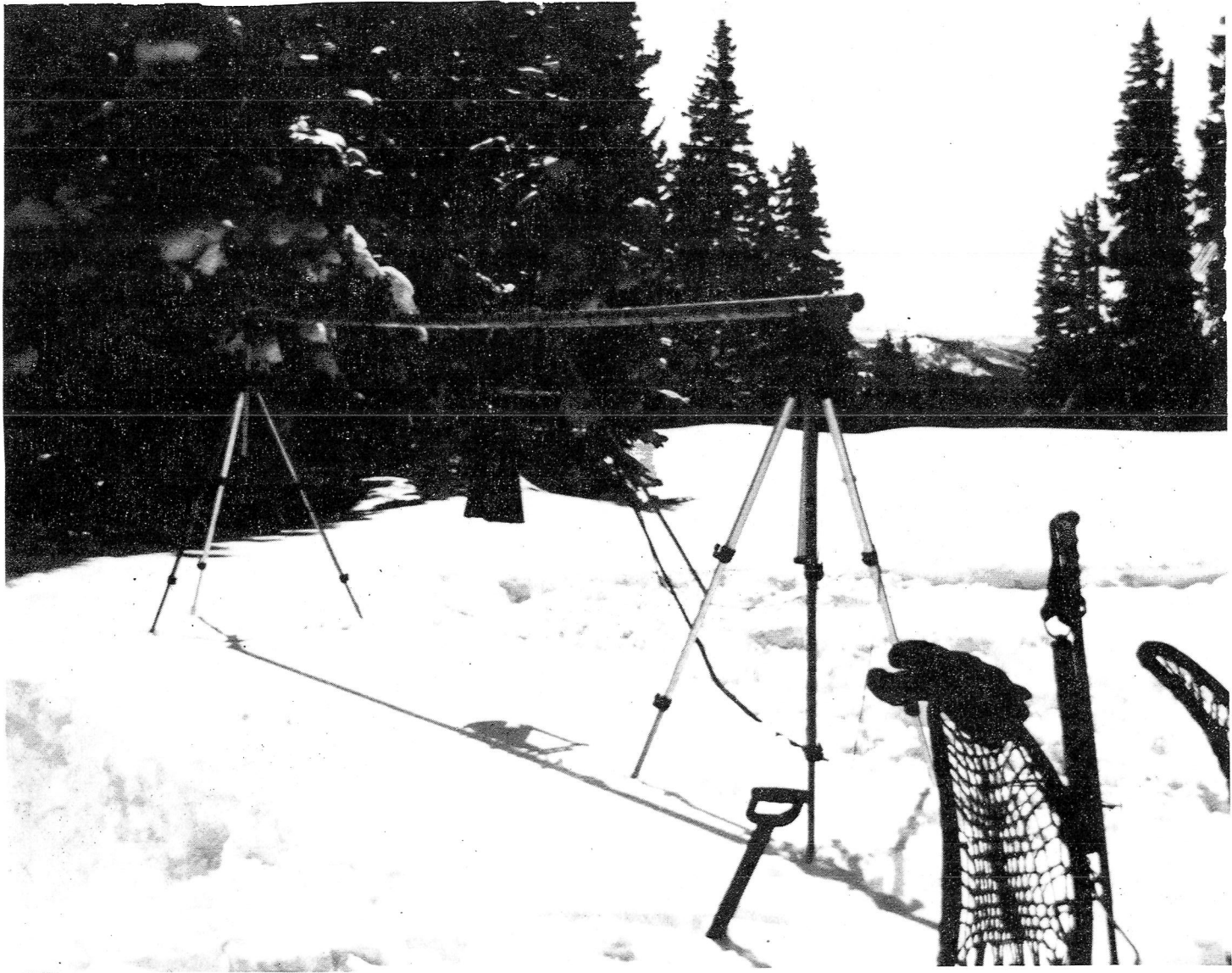
Figure 1

FIGURE 2

STRATIGRAPHY AND TEMPERATURE PROFILE
OF SNOWPACK

Figure 2

Figure 3

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Figure 4

BUFFALO PASS TOWER

Figure 5

129
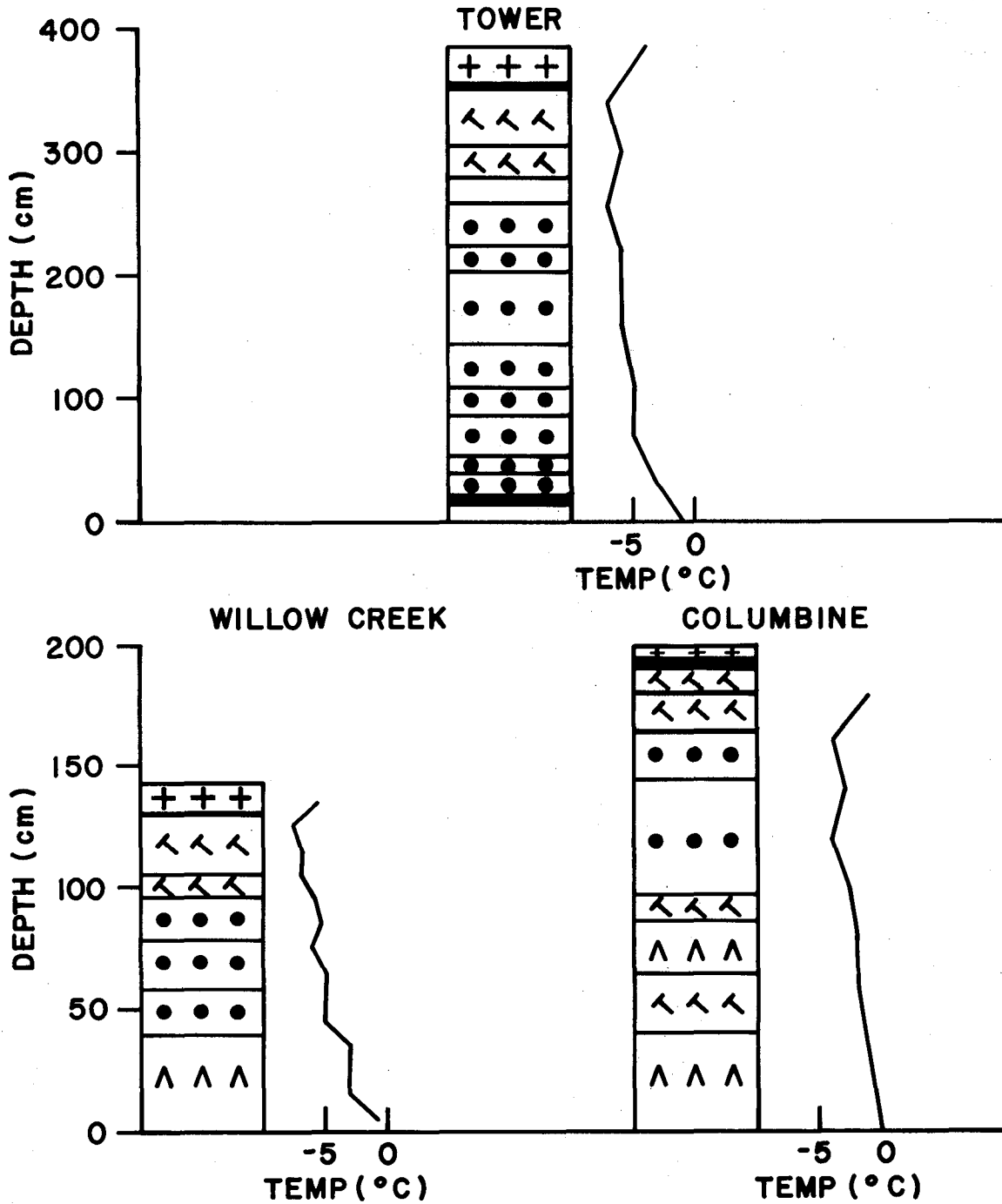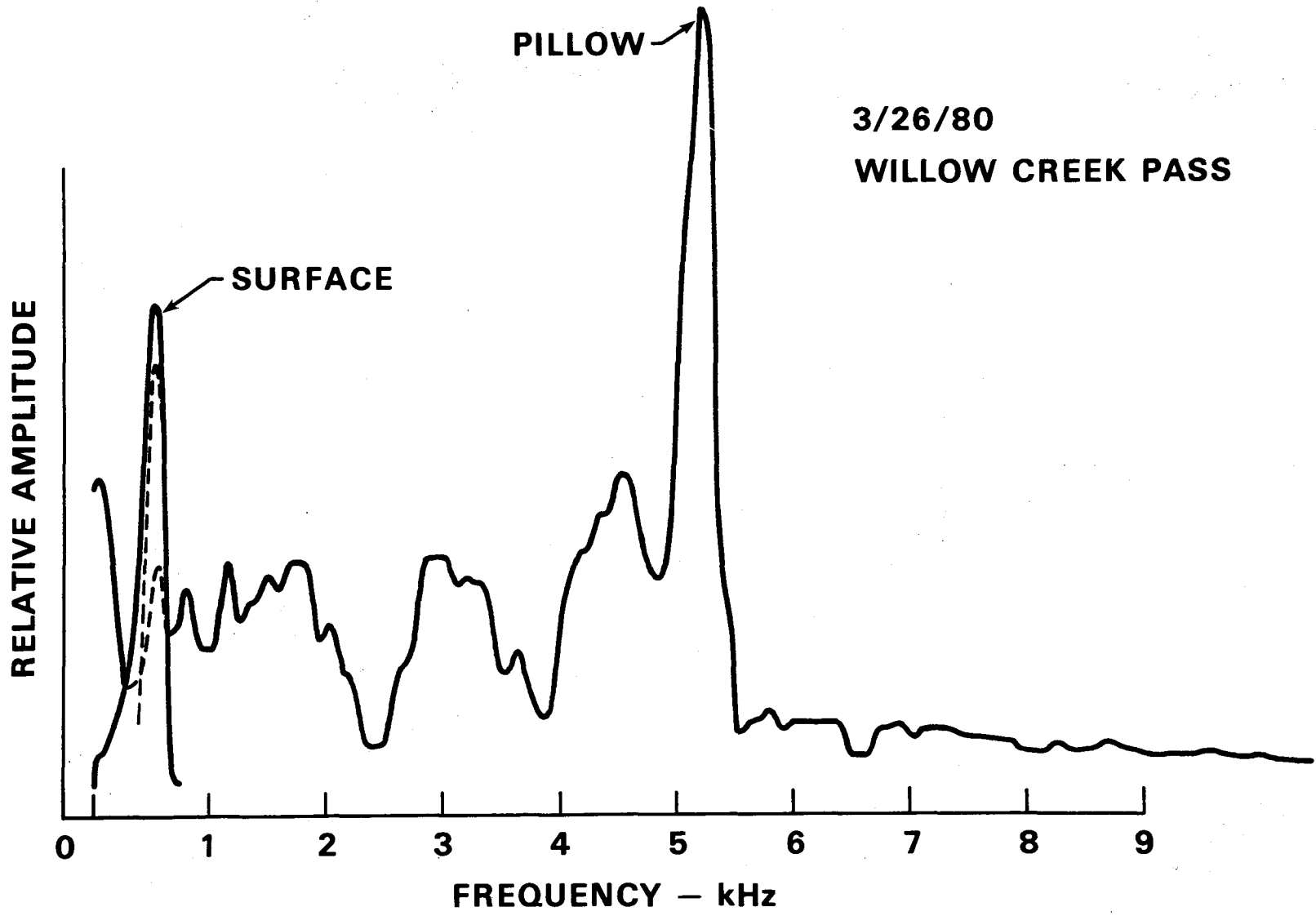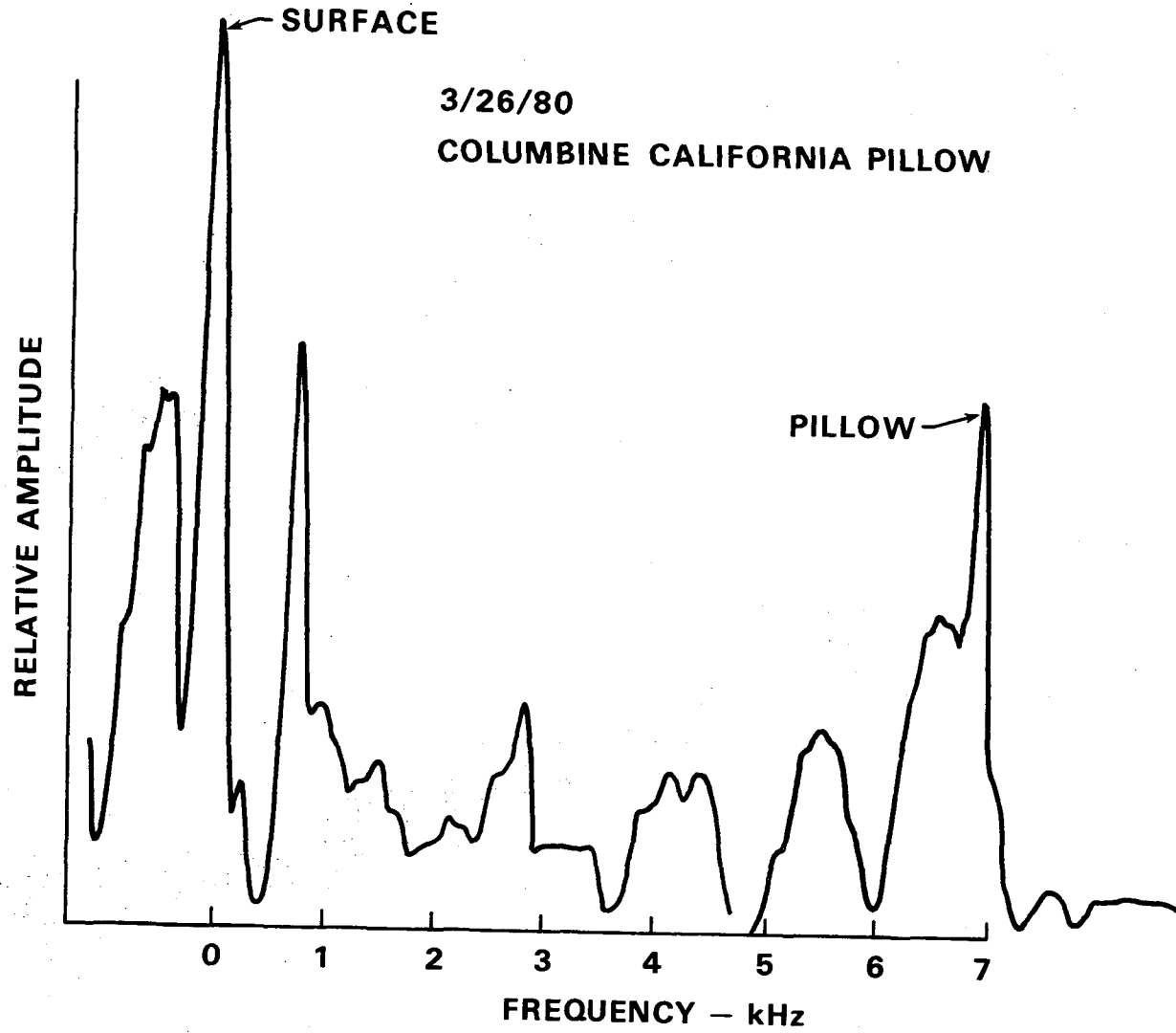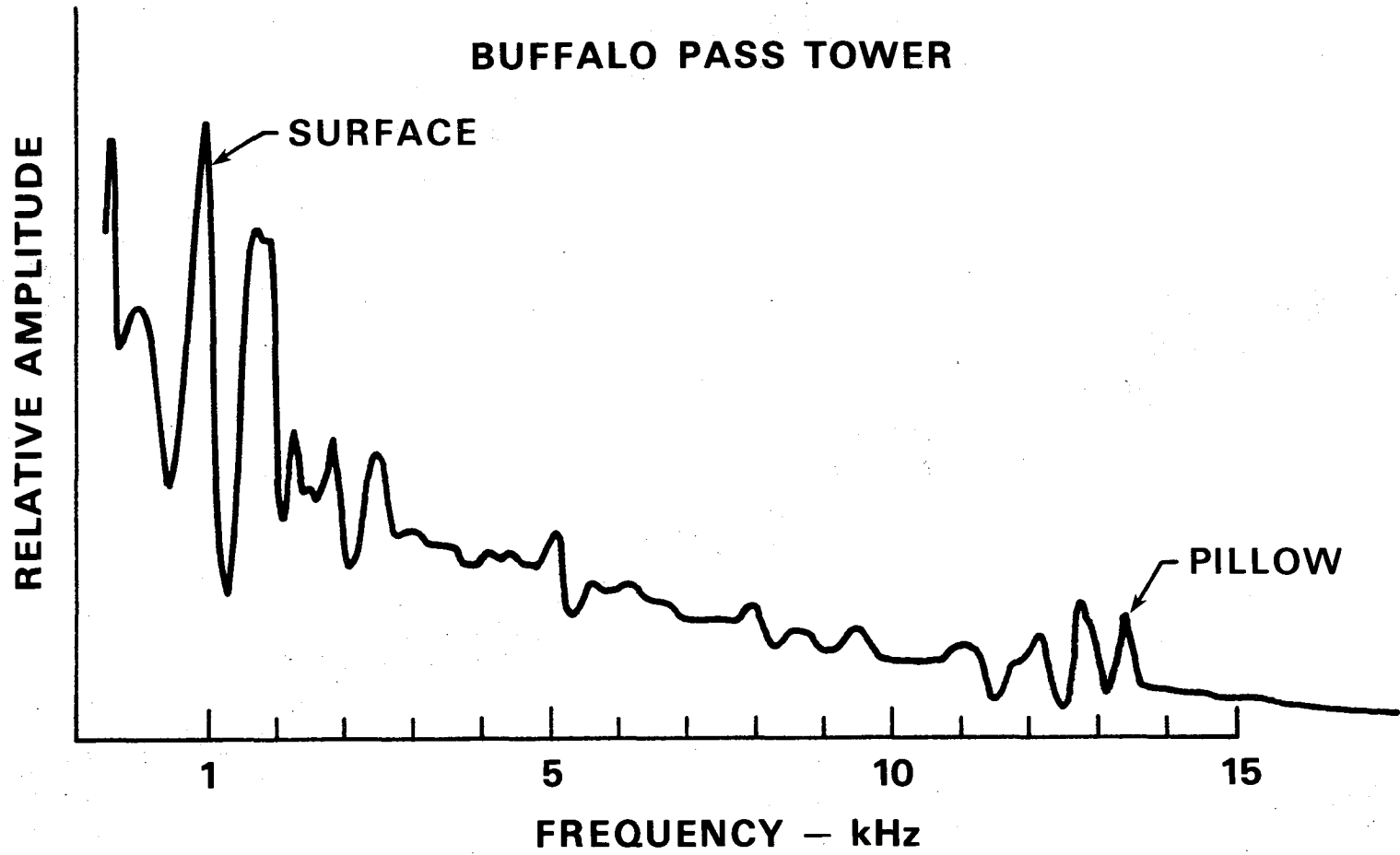
RADAR OBSERVATIONS OF SNOWPACKS*

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ABSTRACT

Radar observations of snowpacks were made at test sites in Kansas, South Dakota and Colorado using truck-mounted scatterometers covering the 1-18 GHz frequency range (30 cm to 1.7 cm in wavelength) and the atmospheric window frequency of 35 GHz (0.86 cm wavelength). Experiments were conducted as a function of snow depth, wetness and surface roughness. The acquired data were used to model the backscattering coefficient in terms of snow and underlying soil parameters. The results indicate that the radar return a) increases with increasing water equivalent, b) decreases with increasing wetness, c) is sensitive to the snow surface roughness only when the snow is wet, d) is sensitive to the state (frozen or thawed) of the underlying soil if the snow is dry, and e) is repetitive from one site to another and from one season to the next. Additionally, the measurements indicate that multifrequency observations or day-night observations may potentially provide the means for monitoring snow water equivalent, snow wetness and the soil state.

INTRODUCTION

Radar reflectivity, usually called the backscattering coefficient σ^o, is analogous to the optical reflectivity measured by an optical radiometer or scanner, except that with radar the observation direction is the same as the illumination direction. After factoring out the functional dependence on the radar system and propagation parameters, the remaining tonal variations in a radar image represent the variations of σ^o of the imaged scene. Hence, the first step one should make in evaluating the potential of radar for mapping snowpack parameters is to determine the response of σ^o to these parameters. This determination is the subject of the present paper.

Whereas the optical reflectivity is governed by the surface layer of the snowpack, the radar backscattering coefficient generally is governed by scattering from the entire snowpack volume and may be influenced by contributions from the underlying ground surface. The degree of penetration of microwaves into the snowpack is a function of the microwave frequency (or wavelength) and the snow wetness. The purpose of the present investigation is to conduct spectral measurements of σ^o for snow under a variety of conditions in

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order to propose a radar design configuration that can remotely sense snow water equivalent and an "effective" value of the snowpack wetness. The results obtained to date show that this objective is realizable, although additional research is needed.

Truck-mounted platforms are especially suitable for the acquisition of accurate σ^0 data, particularly in the basic research phase, where detailed ground-truth information is needed. The results given in this paper are based on measurements conducted at test sites in Kansas, South Dakota and Colorado. Some of these results have been reported elsewhere [1-4], along with detailed descriptions of the ground-truth data acquisition procedures and the Microwave Active Spectrometer (MAS) radar systems. A list of the major snow, soil and weather parameters recorded in conjunction with the radar measurements is given in Table 1. The key MAS specifications also are included.

MODELING THE RADAR BACKSCATTER FROM SNOW

The backscatter from a snow scene consists of surface scattering from the snow-air interface, volume scattering from the snow layer and surface scattering from the ground-snow interface (if the snow attenuation loss is small). The backscattering from the snow volume is due to the dielectric inhomogeneity, which is related to the dielectric constant of the ice crystals (relative to air) and to their size and spatial distributions, relative to the wavelength of the signal in the snow medium. The backscattered energy consists of a direct component that accounts for the direct backscatter from the individual scatterers (crystals) and a diffuse component that accounts for multiple scatter. The radiative transfer model [5], which originally was developed for treating the emission problem, recently was used for modeling the backscatter case [6,7]. Due to the complexity of these models, however, a simpler semi-empirical approach is used in this paper to facilitate the discussion of the microwave observations. Ignoring diffuse scattering and multiple reflections between the two interfaces within which the snow layer is bounded, σ^0 may be written in the following form:

$$\sigma^0(\theta) = T_{sa}^2(\theta) \left\{ \frac{\eta}{2\kappa_e} \left(1 - \frac{1}{L^2(\theta')} \right) \cos\theta + \frac{\sigma_{soil}^0(\theta')}{L^2(\theta')} \right\} \qquad (1)$$

where

$T_{sa}(\theta)$ = power transmission coefficient of the snow-air boundary, dimensionless

θ = angle of incidence from nadir

η = volume reflectivity of snow, cm^{-1}

κ_e = volume extinction coefficient of snow, cm^{-1}

$\sigma_{soil}^0(\theta')$ = backscattering coefficient of the underlying ground medium

θ' =   angle of refraction, related to θ by Snell's Law

$L(\theta')$ =   loss factor of the snow layer

      = $\exp(\kappa_e\, d\, \sec\theta')$, where d is the snow depth in cm.

Equation (1), which was adapted from a similar expression developed for a vegetation canopy [8], does not include the backscatter contribution of the snow-air surface, and therefore it does not apply for σ^0 near nadir, nor can it account for changes in snow surface roughness.

To evaluate the dependence of σ^0 on snow parameters, measurements were conducted as a function of frequency and angle for a variety of snowpack conditions. Typical examples of the angular and spectral behavior are shown in Figures 1 and 2. In each case, comparison is made between σ^0 for dry snow and for wet snow, where the latter is characterized by the volumetric snow wetness (m_v) of the top 5 cm snow layer. It is observed that the sensitivity to m_v increases with increasing frequency; for σ^0 in dB, the difference, $\Delta\sigma^0 = \sigma^0(\text{wet}) - \sigma^0(\text{dry})$, increases from about 1 dB at 1 GHz to 15 dB at 35.6 GHz. A difference of 15 dB corresponds to a ratio of 50 for $\sigma^0(\text{dry})/\sigma^0(\text{wet})$ when σ^0 is expressed in natural units (m^2/m^2). A more detailed discussion of the σ^0 dependence on m_v is given in the next section.

BACKSCATTER RESPONSE TO SNOW WETNESS

A convenient method for studying the radar response to snow wetness variations is by monitoring σ^0 over a diurnal cycle, simultaneously with frequent sampling of snow wetness for several snow layers, particularly the top surface layer. Several diurnal experiments of this type were conducted during the winter seasons of 1977-1980. An example of the results of a two-day experiment conducted at a site near Brookings, South Dakota in 1979 is shown in Figure 3. During the first day, characterized by completely cloudy, overcast conditions and below 0°C air temperature, σ^0 (at 17 GHz) remained essentially constant (within the system measurement precision) as a function of time for all angles of incidence θ. The measured snow wetness m_v was zero throughout the first day. The combination of clear sky conditions and warm daytime air temperatures (as high as 4.1°C) on the second day resulted in the large wetness variation observed in Figure 3. The values of m_v are for the top 5 cm layer of the snowpack. In response to this rapid change in m_v, the radar backscattering coefficient exhibited correspondingly large variations in magnitude at all angles except nadir (0°). The weak dependence of σ^0 on m_v at nadir is not well understood at the present time. The dependence at the higher angles is governed by the ratio η/κ_e. At 17 GHz, κ_e increases rapidly with m_v, resulting in large values for L, even for values of m_v as small as 1 percent by volume [9]. For L large, the soil backscatter contribution (second term in Equation (1)) becomes negligible in comparison to the first term, and $1/L^2$ may be neglected in favor of unity in the first term. With these simplifications, Equation (1) reduces to:

$$\sigma^0(\theta) = \frac{T_{sa}^{2}(\theta) \, \eta \, \cos\theta}{2\kappa_e} \tag{2}$$

The transmission coefficient $T_{sa}(\theta)$ is weakly dependent on m_v (through the snow dielectric constant) while the ratio η/κ_e decreases rapidly with increasing wetness due to the strong dependence of the absorption coefficient on wetness. Moreover, the rate of increase of the absorption coefficient with wetness is directly proportional to the microwave frequency. Hence, the sensitivity to wetness increases with frequency.

WETNESS RESPONSE

In the general case, σ^0 represents backscatter contributions integrated over the snowpack depth, as well as backscatter from the underlying soil medium. Since, in general, m_v is not necessarily uniform with depth, it is not possible to formulate a simple, direct relationship between σ^0 and m_v. Such a formulation requires the availability of detailed information on the m_v profile with a vertical resolution of about 1 cm, particularly for the top 10 cm layer. Nonetheless, as a first step towards understanding the radar response to snow wetness, the wetness of the top 5 cm layer will be used to represent the wetness of the snowpack. Electromagnetically, this assumption is justified by the fact that the top layer exhibits the greatest influence on the radar response, and at the higher microwave frequencies, the attenuation by the top layer (when wet) masks the contributions from the lower layers. The data shown in Figure 3 are plotted in Figure 4 in the form of σ^0(dB) as a function of m_v. The σ^0 response is linear with m_v up to about 4 or 5 percent by volume, beyond which a saturation-like behavior is observed.

The diurnal behavior observed in Figure 3 is typical of all the diurnal experiments conducted at all test sites; at nadir, σ^0 is insensitive to snow wetness variations and at $\theta > 20°$, σ^0 decreases rapidly with m_v. The sensitivity to m_v is frequency-dependent. Figure 5 shows the diurnal response observed at a test site near Steamboat Springs, Colorado in 1977. The σ^0 curves correspond to five microwave frequencies, all at $\theta = 50°$. The magnitude of the observed dip in σ^0 (in response to m_v) increases from about 1 dB at the lowest frequency of 1.2 GHz to 15 dB at 35.6 GHz.

BACKSCATTER RESPONSE TO WATER EQUIVALENT

For a snowpack with a density profile $\rho(z)$ defined over the range $0 \leqslant z \leqslant d$, the snow water equivalent is

$$W = \int_0^d \rho(z) \, dz \tag{3}$$

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If ρ is in g/cm^3 and d is in cm, W is in cm of water. For a homogeneous snowpack with a constant density ρ, $W = \rho d$.

The extinction coefficient κ_e may be expressed in terms of the mass extinction coefficient κ_e' as follows:

$$\kappa_e = \kappa_e' \, \rho \tag{4}$$

The above conversion leads to the following expression for the loss factor $L(\theta')$:

$$L(\theta') = \exp(\kappa_e \, W \, \sec\theta') \tag{5}$$

An expression similar to Equation (4) may be defined for the volume reflectivity η, namely $\eta = \eta'\rho$, but since η in Equation (1) is divided by κ_e, the ratio η/κ_e is unaffected by the conversion to mass coefficients. Using Equation (5), Equation (1) may be rewritten in the abbreviated form:

$$\sigma^o(\theta) = A(\theta) - B(\theta) \exp\left[-C(\theta') \, W\right] \tag{6}$$

where

$$A(\theta) = \frac{T_{sa}^{\,2}(\theta) \, \eta \, \cos\theta}{2\kappa_e} \tag{7}$$

$$B(\theta) = A(\theta) - T_{sa}^{\,2}(\theta) \, \sigma_{soil}^o(\theta') \tag{8}$$

$$C(\theta') = 2\kappa_e' \, \sec\theta' \tag{9}$$

For a given antenna polarization and fixed values of the angle of incidence θ and frequency f, the quantities $A(\theta)$, $B(\theta)$ and $C(\theta')$ are constants if the snow properties are constant with depth. Under these conditions, the dependence of σ^o on W is explicitly defined by Equation (6). To evaluate this dependence experimentally, snow was piled up in horizontal layers that were 20-30 cm thick each, up to a depth of 170 cm. Dry, newly-fallen snow was used throughout, thereby maintaining approximately uniform snow density and crystal size distribution as a function of height between the soil surface and the snow surface. The results of this experiment are shown in Figure 6, together with empirical fits that were generated on the basis of the form of Equation (6). As would be expected, the attenuation coefficient (represented by $C(\theta')$) increases with frequency, thereby causing σ^o to saturate faster at 16.6 GHz in comparison to 9 GHz. This observation implies that for estimating

snow water equivalent with radar, 9 GHz is superior to 16.6 GHz and if the range of snowpack depths of interest is greater than 1.5 meters, frequencies lower than 9 GHz are needed.

In addition to providing information on the response of σ^o to W, the above experiment also provides partial information on the sensitivity of σ^o to snow crystal sizes. The majority of snow crystals (in the freshly-fallen snow used in the snowpile experiment) were smaller than 0.5 mm in diameter. In contrast, the natural snowpack contained crystals that were as large as 5 mm in diameter, primarily in the layers close to the bottom of the snowpack. Comparison of the σ^o values obtained from Figure 6 for W = 10.5 cm with the values of σ^o shown in Figure 2 (at 9 and 16.6 GHz) for the dry natural snowpack case shows that the difference is within 1.5 dB. Thus, in the 9-16.6 GHz range, σ^o appears to be relatively insensitive to snow "type."

TEMPORAL BEHAVIOR

During the 1977 investigation, the microwave observations were made (irregularly) over a six-week period. Figure 7 shows the temporal variation of σ^o at 8.6 GHz for $\theta = 50°$. Also shown is the temporal variation of the water equivalent W and wetness m_v (in the top 5 cm layer).

The following observations are noted:

(a) For the period prior to 3/17/77, σ^o is driven by the snow wetness of the top 5 cm layer. During this period, whenever the top layer was dry (m_v = 0), the entire snowpack was dry.

(b) For the period after 3/17/77, some of the microwave observations were made for snowpack conditions characterized by a dry surface layer over slightly wet lower layers. With the top layer dry, the attenuation through it is small, and therefore σ^o is influenced by the wetness of the lower layers, which explains the observed behavior of σ^o after 3/17/77.

(c) The dashed lines in Figure 7 represent the envelope of the σ^o variation prior to 3/17/77. It is observed that σ^o_{min} remains constant throughout this period. That is, the limiting value of σ^o for wet snow is independent of the water equivalent W. This is to be expected because when the top layer is substantially wet, it masks the contributions from the remainder of the snowpack and from the underlying soil.

σ^o_{max}, corresponding to dry snowpack conditions, increases monotonically with time, in response to the increase of W from about 6 cm (26 cm depth) at the beginning of the observation period to 13.5 cm (52 cm depth) at the end.

CONCLUDING REMARKS

Experimental observations of the backscattering coefficient (σ^0) of snowpacks show that the two snow parameters that exercise the greatest influence over σ^0 are snow wetness of the top layer (or lower layers if the top layer is dry) and snow water equivalent. Sensitivity to both parameters increases with increasing microwave frequency. For snow depths of 50 cm or less, σ^0 is dominated by the backscatter contribution of the underlying ground medium for frequencies below about 4 GHz. As frequency is increased, the depth of snow (or water equivalent) at which the snowpack appears electromagnetically semi-infinite in depth decreases. Above 15 GHz, σ^0 is very sensitive to W over the range $0 \leqslant W \leqslant 20$ cm, but loses sensitivity (saturates) as W is increased beyond 20 cm. As a compromise between minimizing the influence of the underlying ground medium and simultaneously maintaining good sensitivity to variations in W, a frequency in the neighborhood of 9 GHz is proposed. A superior configuration is one consisting of a multiple-frequency system, such as a low frequency in the 1-4 GHz range whose purpose, primarily, would be to monitor variations in the state (frozen or thawed) of the underlying ground medium; an intermediate frequency around 9 GHz for monitoring W; and two additional frequencies at 17 GHz and 35 GHz which, together with the lower frequencies, may be used for estimating the snow wetness profile of the snowpack.

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TABLE 1. List of Measured Ground-Truth Parameters
and Radar System Specifications.

SNOW PARAMETERS

Depth

Density Profile

Water Equivalent

Wetness by Volume of
Top 5 cm Layer

Temperature Profile

Stratification

SOIL PARAMETERS

Temperature Profile

Moisture Content

ATMOSPHERIC PARAMETERS

Air Temperature

Barometric Pressure

Relative Humidity

OTHER

Solar Radiation

Cloud Cover

MICROWAVE ACTIVE SPECTROMETER SPECIFICATIONS

Frequency: 1-18 GHz in approximately 1 GHz steps, and 35.6 GHz

Angle of Incidence Range: 0° (nadir) - 70°

Polarization Configurations: HH, HV, VV

Calibration Method: Internal - Signal Injection

External - Luneberg Lens Reflector
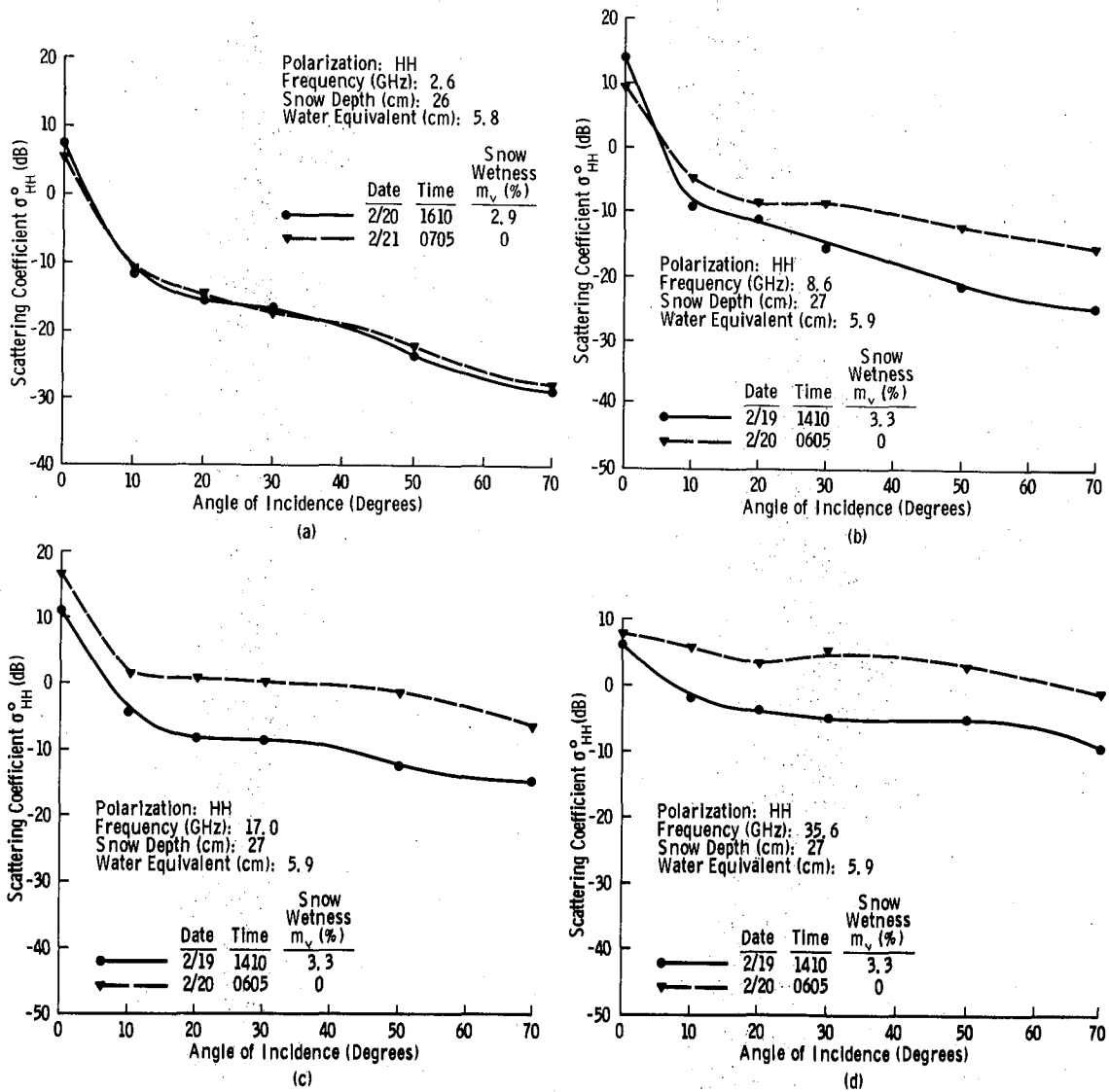
Figure 1. Angular Response of σ° to Wet and Dry Snow at: (a) 2.6 GHz, (b) 8.6 GHz, (c) 17.0 GHz, and (d) 35.6 GHz.
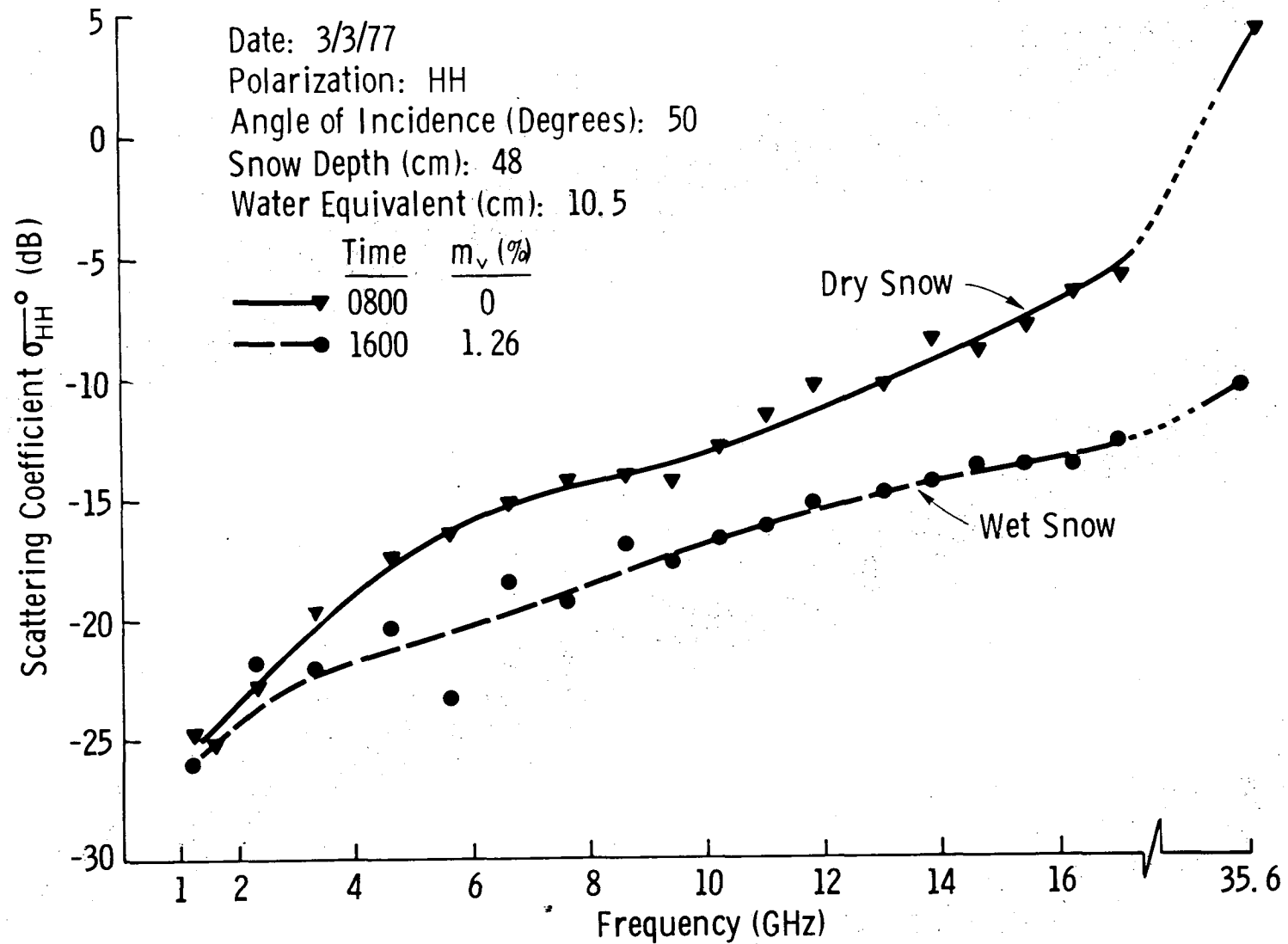
Figure 2. Spectral Response of σ^o at 50° Angle of Incidence to Wet and Dry Snow
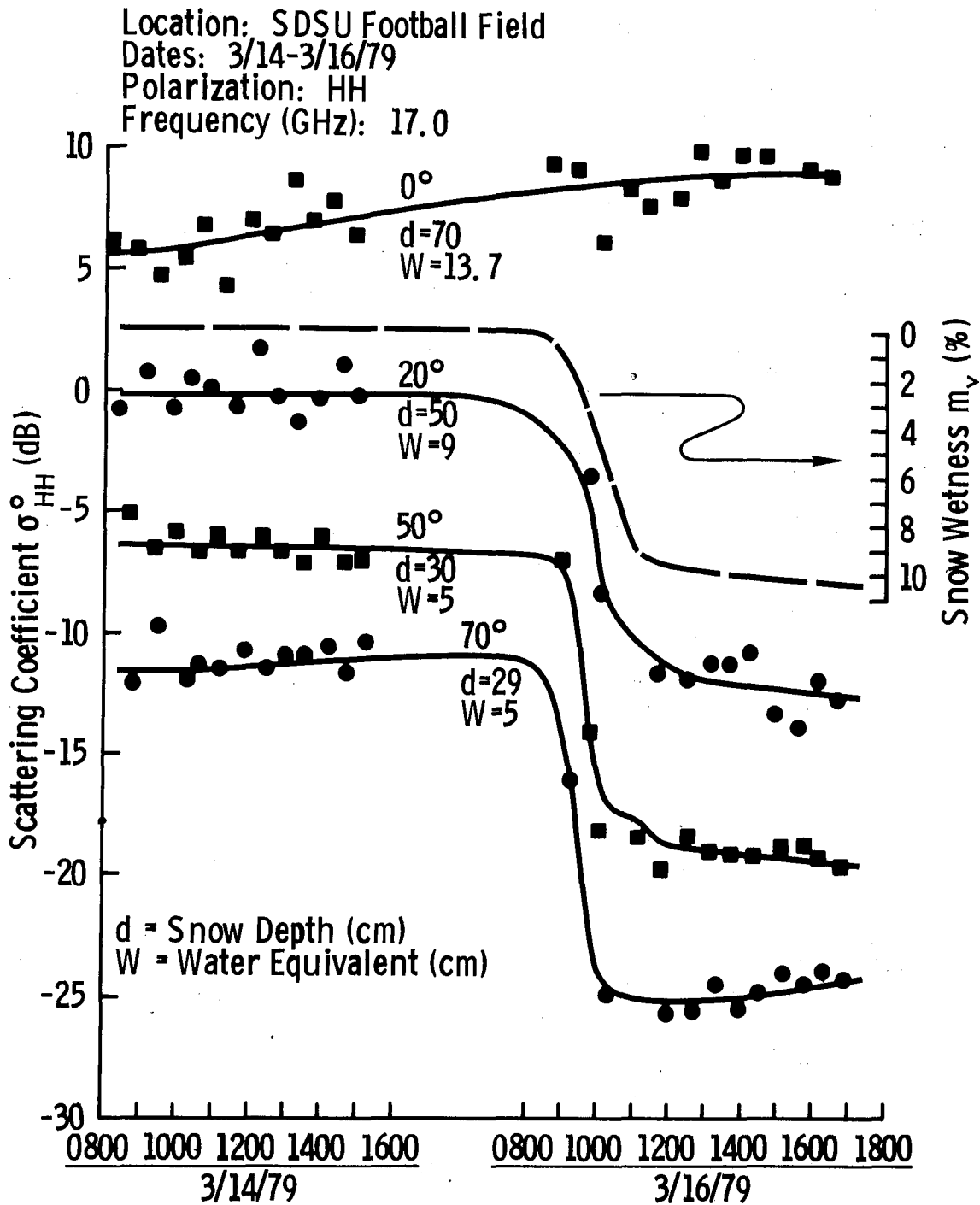
Figure 3. Diurnal variation of σ^o for two days. The first
day (3/14/79) was characterized by completely
overcast conditions and below 0°C air temperature,
while the second day (3/16/79) was characterized
by above 0°C air temperature and clear sky
conditions leading to snow wetness values as high
as 10% by volume. The target was the South Dakota
State University football field.
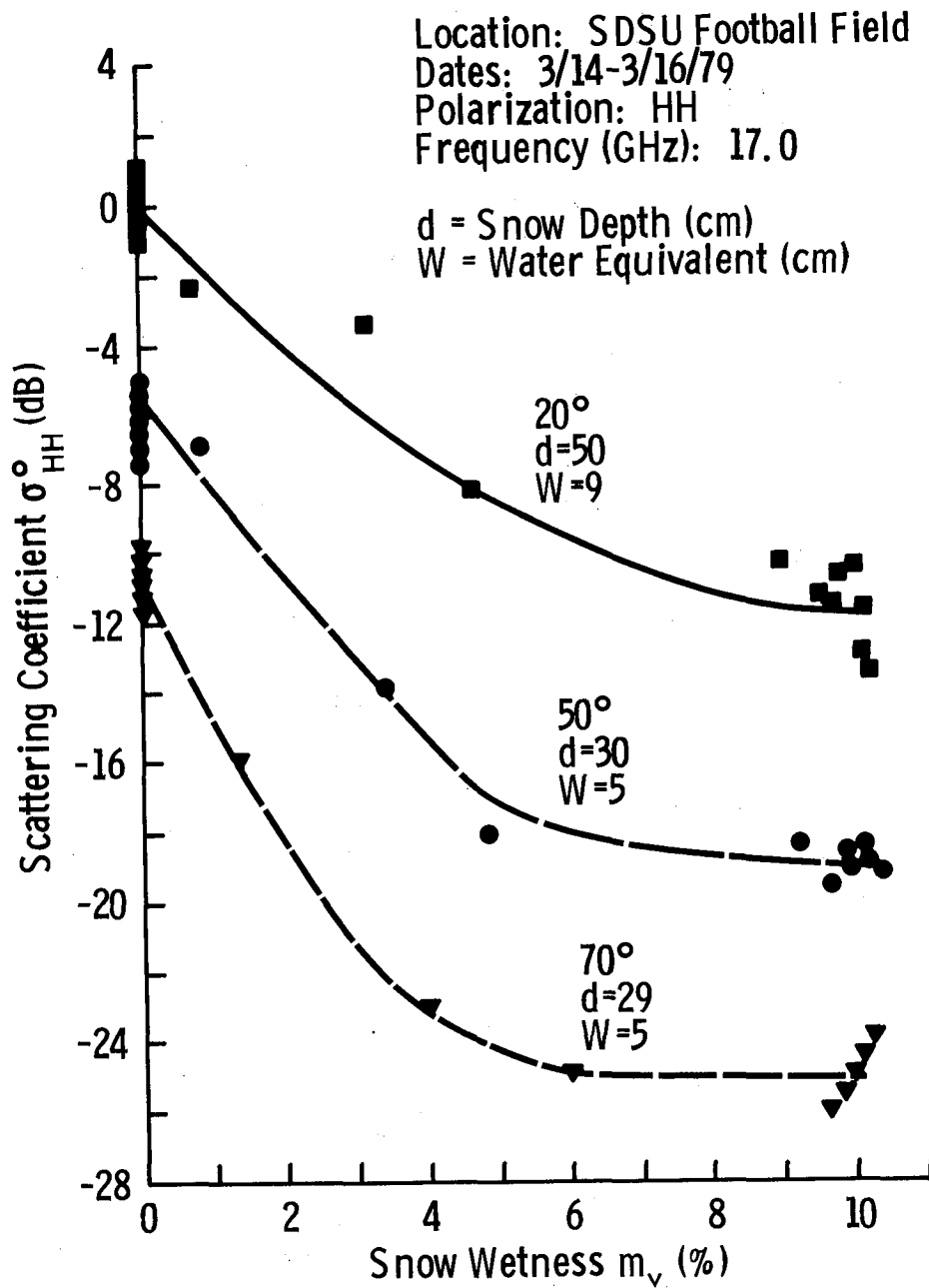
Figure 4. Response of σ^0 to snow wetness at 17 GHz.
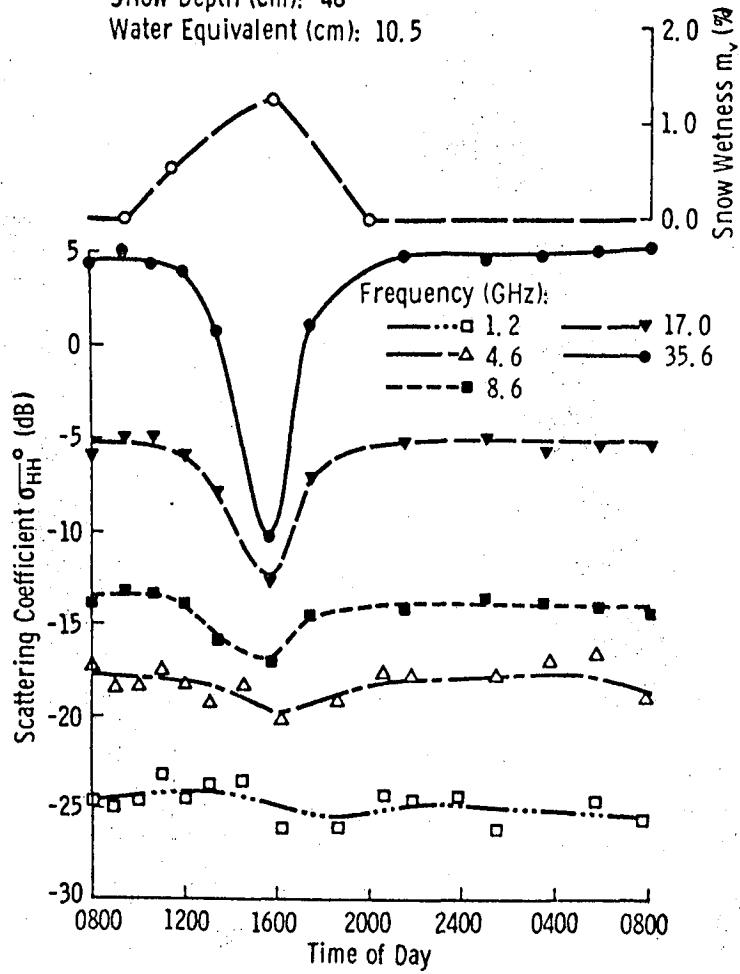
Figure 5. Diurnal variation of σ^o at five microwave frequencies, and of snow wetness in the top 5 cm snow layer.
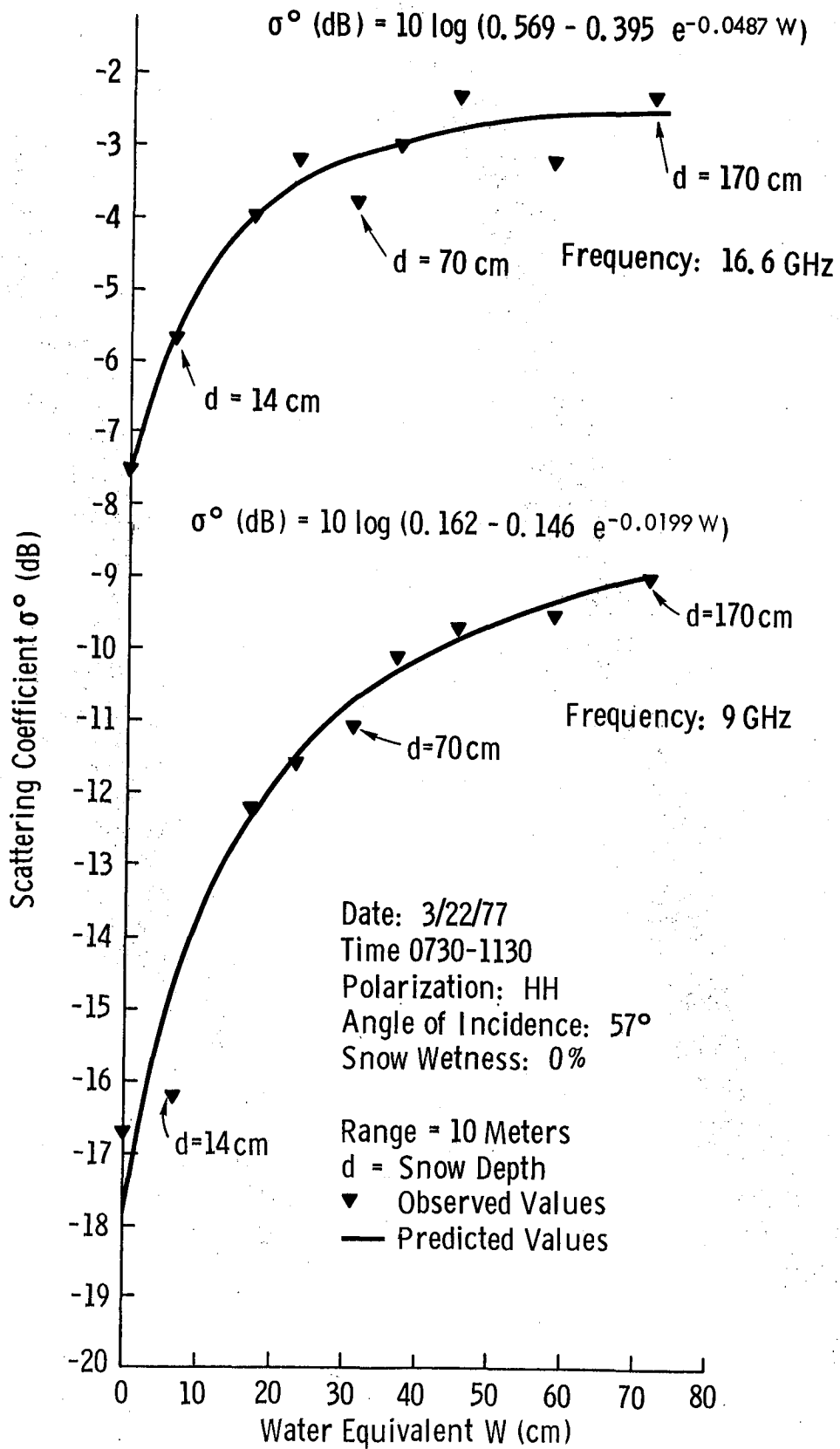
Figure 6. Scattering coefficient response to snow water equivalent
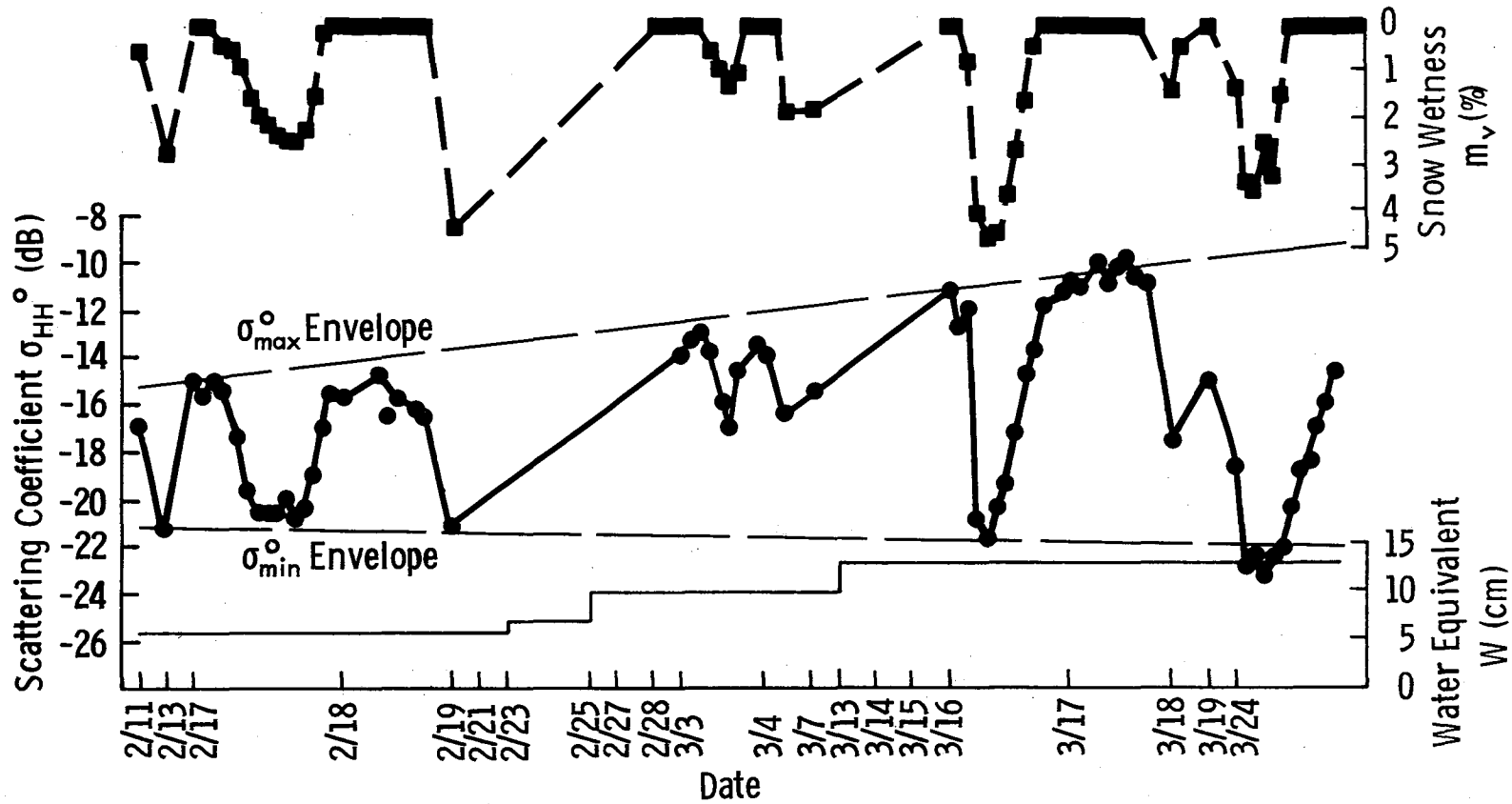
Figure 7. Temporal Variation of σ° Value Over the Experiment Duration at 8.6 GHz and 50° Angle of Incidence.

THEORETICAL MODELS FOR MICROWAVE SNOW RESPONSE AND

APPLICATIONS TO REMOTE SENSING*

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ABSTRACT

In both the active and passive microwave remote sensing of snowpacks, volume scattering effects due to medium inhomogeneities play a dominant role in the determination of the radar backscattering cross sections and the brightness temperatures. Two theoretical models have been developed to characterize snowpacks: (1) a random medium with a variance, a horizontal correlation length, and a vertical correlation length and, (2) a homogeneous dielectric containing discrete scatterers. The earth terrain is then modelled as layers of such scattering media bounded by air above and a homogeneous half-space below. The development of the theoretical approach is guided by the motivation that data set obtained in a field and plotted as functions of frequency, angle, and polarization must be matched with same set of parameters characterizing the same field. In matching the theoretical results with experimental data collected from snow-ice fields, we summarize the following findings: (1) For snow-ice fields the horizontal correlation length is no less than the vertical correlation length signifying a more laminar structure. The correspondence between the continuum random medium and the discrete spherical scatterer model can be verified when the vertical correlation length is equal to the horizontal correlation length. (2) The vertically polarized backscattering cross section σ_{vv} is always greater than the horizontally polarized backscattering cross section σ_{hh} for half-space scattering media and may become smaller for a two-layer model. (3) To account for diurnal change exhibited by snow fields in both the active and passive remote sensing cases, a three-layer model with a thin top layer caused by solar illumination must be used.

INTRODUCTION

In developing theoretical models for both active and passive microwave remote sensing of snowpacks, we account for the volume scattering effects by modelling snowpacks as a random medium [1-7] or a homogeneous medium containing discrete scatterers [8-11]. The random medium has an average permittivity ε_1 and its random part is characterized by a correlation function with variance δ, horizontal correlation length ℓ_ρ and vertical correlation length ℓ_z. In the discrete-scatterer model, we characterize the scatterers as spheres of radius a, permittivity ε_s and effective fractional volume f imbedded inside a homogeneous medium with permittivity ε_1. With the earth terrain modelled as layers of such scattering media bounded by air above and half-space earth below, we then calculate the radar backscattering cross-sections for active remote sensing and the radiometric brightness temperatures for passive remote sensing.

THEORY

For the random medium model, radar backscattering cross-sections are calculated with an iterative procedure to the integral equations for the scattered intensities which gives rise to a Born series that converges quickly for small albedo. The first order terms yield the backscattering cross-section σ_{vv} for the vertically polarized return and σ_{hh} for the horizontally polarized return. For a layer of random medium with thickness d_1, the results are relatively simple and they are given as follows [5]:

$$\sigma_{hh} = \frac{\delta k_1'^4 \ell_z \ell_\rho^2}{4} \frac{|X_{10i}|^4}{|D_{2i}|^4} \left|\frac{k_{ozi}}{k_{1zi}}\right|^4 e^{-k_o^2 \ell_\rho^2 \sin^2\theta_{oi}}$$

$$\left\{ \frac{(1 - e^{-4k_{1zi}'' d_1})(1 + |R_{12i}|^4 e^{-4k_{1zi}'' d_1})}{2k_{1zi}''(1 + 4k_{1zi}'^2 \ell_z^2)} \right.$$

$$\left. + 8d_1 |R_{12i}|^2 e^{-4k_{1zi}'' d_1} \right\} \tag{1}$$

$$\sigma_{vv} = \frac{\delta k_1'^4 \ell_z \ell_\rho^2}{4} \frac{|Y_{10i}|^4}{|F_{2i}|^4} \left|\frac{k_{ozi}}{k_{1zi}}\right|^4 e^{-k_o^2 \ell_\rho^2 \sin^2\theta_{oi}}$$

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$$\left\{ \frac{(1 - e^{-4k''_{1zi}d_1})(1 + |S_{12i}|^4 e^{-4k''_{1zi}d_1})}{2k''_{1zi}(1 + 4k'^2_{1zi}\ell_z^2)} \left| \frac{k^2_{1zi}}{k_o^2} + \sin^2\theta_{oi} \right|^2 \right.$$

$$\left. + 8d_1|S_{12i}|^2 e^{-4k''_{1zi}d_1} \left| \frac{k^2_{1zi}}{k_o^2} - \sin^2\theta_{oi} \right|^2 \right\} \qquad (2)$$

where the subscript i indicates the incident direction and the correlation function is taken to be gaussian horizontally and exponential vertically. The other symbols in (1)-(2) are defined as follows

$$R_{\ell m} = \frac{k_{\ell z} - k_{mz}}{k_{\ell z} + k_{mz}} \qquad\qquad S_{\ell m} = \frac{\varepsilon_m k_{\ell z} - \varepsilon_\ell k_{mz}}{\varepsilon_m k_{\ell z} + \varepsilon_\ell k_{mz}}$$

$$X_{\ell m} = 1 + R_{\ell m} \qquad\qquad Y_{\ell m} = 1 + S_{\ell m}$$

$$D_2 = 1 + R_{01}R_{12} e^{i2k_{1z}d_1} \qquad F_2 = 1 + S_{01}S_{12} e^{i2k_{1z}d_1}$$

$$k_{\ell z} = (k_\ell^2 - k_\perp^2)^{1/2}$$

where $k_o^2 = \omega^2\mu_o\varepsilon_o$, $k_1^2 = \omega^2\mu_o\varepsilon_1$, $k_2^2 = \omega^2\mu_o\varepsilon_2$, $k''_{1z} = \text{Im}(k_{1z})$, and the subscripts ℓ and m denote 0, 1, and 2.

As the layer thickness d_1 becomes very large, the terms involving $\exp[-4k''_{1zi}d_1]$ can be omitted and we see that the bracket terms in (1) and (2) becomes simply $[2k''_{1zi}(1 + 4k'^2_{1zi}\ell^2)]^{-1/2}$. The scattering cross-sections decay as $\exp[-k_o^2\ell_\rho^2\sin^2\theta_{oi}]$ and they are proportional to $|X_{10i}|^4$ for σ_{hh} and $|Y_{10i}|^4$ for σ_{vv}, which implies that σ_{vv} is always larger than σ_{hh} for a half-space random medium. The presence of the bottom boundary is thus very important in order to account for experimental data with $\sigma_{hh} > \sigma_{vv}$. The second order terms

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which give rise to depolarization effects [6], and the results for backscattering by multilayer random media [4] are too complicated to be given here.

In the case of passive remote sensing, we can calculate brightness temperatures by using the radiative transfer theory. The radiative transfer equation inside a scattering medium which can be either a random medium or a homogeneous medium containing discrete scatterers takes the form

$$\cos\theta \; \frac{d}{dz} \; \bar{I}(\theta, z) = K_a \, C_1 T_1 - K_e \, \bar{I}(\theta, z)$$

$$+ \int_0^\pi d\theta' \; \sin\theta' \; \bar{\bar{P}}(\theta, \theta') \cdot \bar{I}(\theta', z) \tag{3}$$

where for $0 < \theta < \pi$

$$\bar{I}(\theta, z) = \begin{bmatrix} I_v(\theta, z) \\ I_h(\theta, z) \end{bmatrix} ,$$

I_v is the vertically polarized intensity, I_h is the horizontally polarized intensity. $K_e = K_a + K_s$ with K_a denoting absorption loss and K_s the scattering loss, $C_1 = K\varepsilon_1'/\varepsilon_o \lambda^2$ with K denoting the Boltzmann constant, and $\bar{\bar{P}}(\theta, \theta')$ is the scattering function matrix for the scattering medium. For the random medium model, K_e in general takes the form of a 2×2 diagonal matrix.

For a layer of scattering medium with boundaries at $z = 0$ and $z = -d_1$, the boundary conditions are, for $0 < \theta < \pi/2$,

$$\bar{I}(\pi - \theta, z = 0) = \bar{\bar{R}}_{10}(\theta) \cdot \bar{I}(\theta, z = 0) + \bar{\bar{T}}_{01}(\theta_o) \cdot \bar{I}_{sky}(\theta_o)$$

$$\tag{4}$$

$$\bar{I}(\theta, z = -d_1) = \bar{\bar{R}}_{12}(\theta) \cdot \bar{I}(\pi - \theta, z = -d_1) + \bar{\bar{T}}_{21}(\theta_2) \cdot \bar{I}_2(\theta_2)$$

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where $\bar{I}_{sky}(\theta_o)$ is the intensity corresponding to the sky temperature, $\bar{I}_2(\theta_2)$ is the intensity corresponding to the temperature in the medium $z < -d_1$, $\bar{\bar{R}}_{10}(\theta)$ is the coupling matrix at the boundary $z = 0$, $\bar{\bar{R}}_{12}(\theta)$ is the coupling matrix at the boundary $z = -d_1$, $\bar{\bar{T}}_{01}(\theta_o)$ is the coupling matrix from air region to the random medium, and $\bar{\bar{T}}_{21}(\theta_2)$ is the coupling matrix from bottom layer to the random medium.

The radiative transfer equations can be solved numerically, using Gaussian quadrature method. We can replace the integrals in the radiative transfer equations by a Gaussian quadrature, an appropriately weighted sum over 2n intervals between the 2n zeros of the even-order Legendre polynomial $P_{2n}(\theta)$, and obtain a system of ordinary differential equations with constant coefficients. The system of equations can then be solved by finding the eigenvalues and eigenvectors and matching the boundary conditions [2,7-10].

The radiative transfer approach can also be applied to the solution of backscattering cros-sections [3,11]. The radiative transfer equations take the form

$$\cos \theta \, \frac{d}{dz} \bar{I}(\theta, \phi, z) = - K_e \bar{I}(\theta, \phi, z)$$

$$+ \int_0^{2\pi} d\phi' \int_0^{\pi} d\theta' \, \sin \theta' \, \bar{\bar{P}}(\theta, \phi; \theta', \phi')$$

$$\cdot \bar{I}(\theta', \phi', z) \tag{5}$$

where the intensity $\bar{I}(\theta, \phi, z)$ contains all four Stoke's parameters,

$$\bar{I}(\theta, \phi, z) = \begin{bmatrix} I_v(\theta, \phi, z) \\ I_h(\theta, \phi, z) \\ U(\theta, \phi, z) \\ V(\theta, \phi, z) \end{bmatrix} \, .$$

The boundary conditions are, for $0 < \theta < \pi/2$,

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$$\bar{I}(\pi - \theta, \phi, z = 0) = \bar{\bar{T}}_{01}(\theta_o) \cdot \bar{I}_{oi}(\pi - \theta_o, \phi_o) + \bar{\bar{R}}_{10}(\theta)$$

$$\cdot \bar{I}(\theta, \phi, z = 0)$$

$$\bar{I}(\theta, \phi, z = -d_1) = \bar{\bar{R}}_{12}(\theta) \cdot \bar{I}(\pi - \theta, \phi, z = -d_1). \qquad (6)$$

The incident beam in region 0, $\bar{I}_{oi}(\pi - \theta_o, \phi_o)$, assumes the form

$$\bar{I}_{oi}(\pi - \theta_o, \phi_o) = \bar{I}_{oi} \, \delta(\cos\theta_o - \cos\theta_{oi}) \, \delta(\phi_o - \phi_{oi}) \qquad (7)$$

where the use of Dirac delta function is made.

The radiative transfer equations can be solved with iterative and numerical approaches. The iterative approach gives closed form solutions which are valid when the effect of scattering is small (small albedo). The radiative transfer equations and the boundary conditions are first cast into integral equation forms. Then an iterative process is applied to solve the integral equations to both the first and second orders in albedo. Again the depolarization of the backscattered intenstiy is shown to be a second order effect. The numerical approach provides a solution which is valid for both small and large albedos. A Fourier series expansion in the azimuthal direction can be used to eliminate the azimuthal ϕ-dependence from the radiative transfer equations. We let

$$\bar{I}(\theta, \phi, z) = \bar{I}^O(\theta, z) + \sum_{m=1}^{\infty} [\bar{I}^{mc}(\theta, z) \cos m(\phi - \phi_i)$$

$$+ \bar{I}^{ms}(\theta, z) \sin m(\phi - \phi_i)] \qquad (8)$$

$$\bar{\bar{P}}(\theta, \phi; \theta', \phi') = \bar{\bar{P}}^O(\theta, \phi') + \sum_{m=1}^{\infty} [\bar{\bar{P}}^{mc}(\theta, \theta') \cos m(\phi - \phi')$$

$$+ \bar{\bar{P}}^{ms}(\theta, \theta') \sin m(\phi - \phi')] \qquad (9)$$

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where superscript m indicates the order of harmonics in the azimuthal direction, and supercripts c and s indicate the cosine and sine dependence. We can substitute (8) and (9) into the radiative transfer equations and carry out the $d\phi'$ integration. Then, by collecting terms with the same cosine or sine dependence, we obtain, for $m = 0$

$$\cos\theta \frac{d}{dz} \bar{I}^O(\theta, z) = - K_e \bar{I}^O(\theta, z) + 2\pi \int_0^\pi d\theta' \sin\theta' \bar{\bar{P}}^O(\theta, \theta')$$

$$\cdot \bar{I}^O(\theta', z) \tag{10}$$

and for $m \geq 1$

$$\cos\theta \frac{d}{dz} \bar{I}^{mc}(\theta, z) = - K_e \bar{I}^{mc}(\theta, z) + \pi \int_0^\pi d\theta' \sin\theta'$$

$$\{\bar{\bar{P}}^{mc}(\theta, \theta') \cdot \bar{I}^{mc}(\theta', z) - \bar{\bar{P}}^{ms}(\theta, \theta') \cdot \bar{I}^{ms}(\theta', z)\} \tag{11}$$

$$\cos\theta \frac{d}{dz} \bar{I}^{ms}(\theta, z) = - K_e \bar{I}^{ms}(\theta, z) + \pi \int_0^\pi d\theta' \sin\theta'$$

$$\{\bar{\bar{P}}^{ms}(\theta, \theta') \cdot \bar{I}^{mc}(\theta', z) + \bar{\bar{P}}^{mc}(\theta, \theta') \cdot \bar{I}^{ms}(\theta', z)\}. \tag{12}$$

The incident intensity $\bar{I}_{oi}(\pi - \theta_o, \phi_o)$ can be expanded into the Fourier series.

$$\bar{I}_{oi}(\pi - \theta_o, \phi_o) = \bar{I}_{oi} \delta(\cos\theta_o - \cos\theta_{oi})$$

$$\left[\frac{1}{2\pi} + \frac{1}{\pi} \sum_{m=1}^\infty \cos m(\phi_o - \phi_{oi}) \right] \tag{13}$$

Using the above equation we can obtain the boundary conditions for each harmonic. Then, the set of radiative transfer equations without the ϕ-dependence can be solved with the method of

Gaussian quadrature [11].

DATA MATCHING

In the exercise of data matching with the theoretical re-
sults, the basic requirement is to come up with one single set of
parameters that matches all experimental data plotted as func-
tions of angle, frequency and polarization and collected from
the same field at the same time. In Fig. 1 we illustrate the
angular match of active remote sensing data obtained from a
snow field for vertically like-polarized backscattering cross
sections using random medium model. The thickness of the snow
is obtained from ground truth data. In Figs. 2-6, we show the
result of data interpretation using the random medium model for
the brightness temperatures measurements obtained as a function
of angle at frequencies of 5, 10.7, 18, and 37 GHz [12]. In
Fig. 2, the brightness temperatures are plotted as a function of
frequency for viewing angle of 33° and matched with the result
of scattering layer with $\ell_\rho = 0.2$ cm, $\ell_z = 0.2$ cm, and $\delta = 0.056$. The permittivities of the snow and the ground are taken
to be $\varepsilon_1 = (1.5 + i0.00375)\varepsilon_o$ and $\varepsilon_b = (6.0 + i0.6)\varepsilon_o$, re-
spectively. The angular dependence of the brightness tempera-
tures at the four different frequencies are matched with the
same theoretical model and are shown in Figs. 3, 4, 5, and 6. In
Fig. 7 we show the data matching using the discrete scatterer
model of the same brightness temperature measurements in order to
show the correspondence between the random medium model and the
discrete scatterer model. Since the data can be matched with a
random medium model having $\ell_\rho = \ell_z = 0.2$ cm, we show that a
homogeneous medium of the same background dielectric of $\varepsilon_1 = (1.5 + i0.00375)\varepsilon_o$ containing spherical scatterers also match
the data. It is noted that the effective fractional volume f is
by no means equivalent to the actual volume occupied by the dis-
crete scatterers. It is well known that the physical volume of
ice particles constituting snow leads to overestimation of scat-
tering effect and a rigorous theory which includes the homogene-
ous dielectric medium as a special case can only be obtained from
a multiple scattering theory for closely packed scatterers[13,14].

In the passive remote sensing of snow field, we encounter
the phenomenon of diurnal change where the brightness tempera-
ture decreases as a function of frequency in the morning and
increases in the afternoon [15]. In order to explain this
phenomenon, we have to resort to a three layer model where we
assume that in the afternoon due to sun light illumination, a
thin layer with higher loss tangent is created so that it pro-
vides more emission and masks the scattering effects at higher
frequencies. In Fig. 8 we show a spectral plot of T_B in the
morning and in Fig. 9 we show the spectral dependence T_B in
the afternoon. Notice that in the thick snow layer, all parame-
ters remain the same so that sunlight only affects the top

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7 cm.

In Fig. 10, vertically and horizontally polarized back-scattering cross sections σ_{vv} and σ_{hh} are plotted as functions of the thickness of the scattering layer at the frequency of 16 GHz and at the incident angle of 60 degrees. Here we illustrate the very interesting point that at a shallow depth, σ_{hh} is higher than σ_{vv} and at a greater depth, σ_{vv} is higher than σ_{hh}. We can attribute this phenomena to the bottom layer. When the thickness of the scattering layer is large and the effect of the bottom layer is small, σ_{vv} is greater than σ_{hh} because more intensities are being transmitted to be scattered in the vertical polarization. However, at the interface of the bottom layer, more intensities in the horizontal polarization are being reflected to be scattered by the scatterers in the backward direction. Thus for more shallow depths, we may have more intensities in the horizontal polarization to be scattered by the scatterers.

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FIGURE CAPTIONS

Figure 1   Vertically polarized backscattering cross sections at 10 GHz as a function of angle for a 59 cm snow layer.

Figure 2   Brightness temperature of 66 cm snow layer as a function of frequency.

Figure 3   Brightness temperature as a function of angle at 5 GHz.

Figure 4   Brightness temperature as a function of angle at 10.7 GHz.

Figure 5   Brightness temperature as a function of angle at 18 GHz.

Figure 6   Brightness temperature as a function of angle at 37 GHz.

Figure 7   Brightness temperature of 66 cm snow layer as a function of frequency.

Figure 8   Brightness temperature of 215 cm snow layer as a function of frequency in the morning.

Figure 9   Brightness temperature of 215 cm snow layer as a function of frequency in the afternoon.

Figure 10   Vertically and horizontally polarized backscattering cross sections as a function of thickness of the scattering layer at 16 GHz and at $\theta = 60°$.
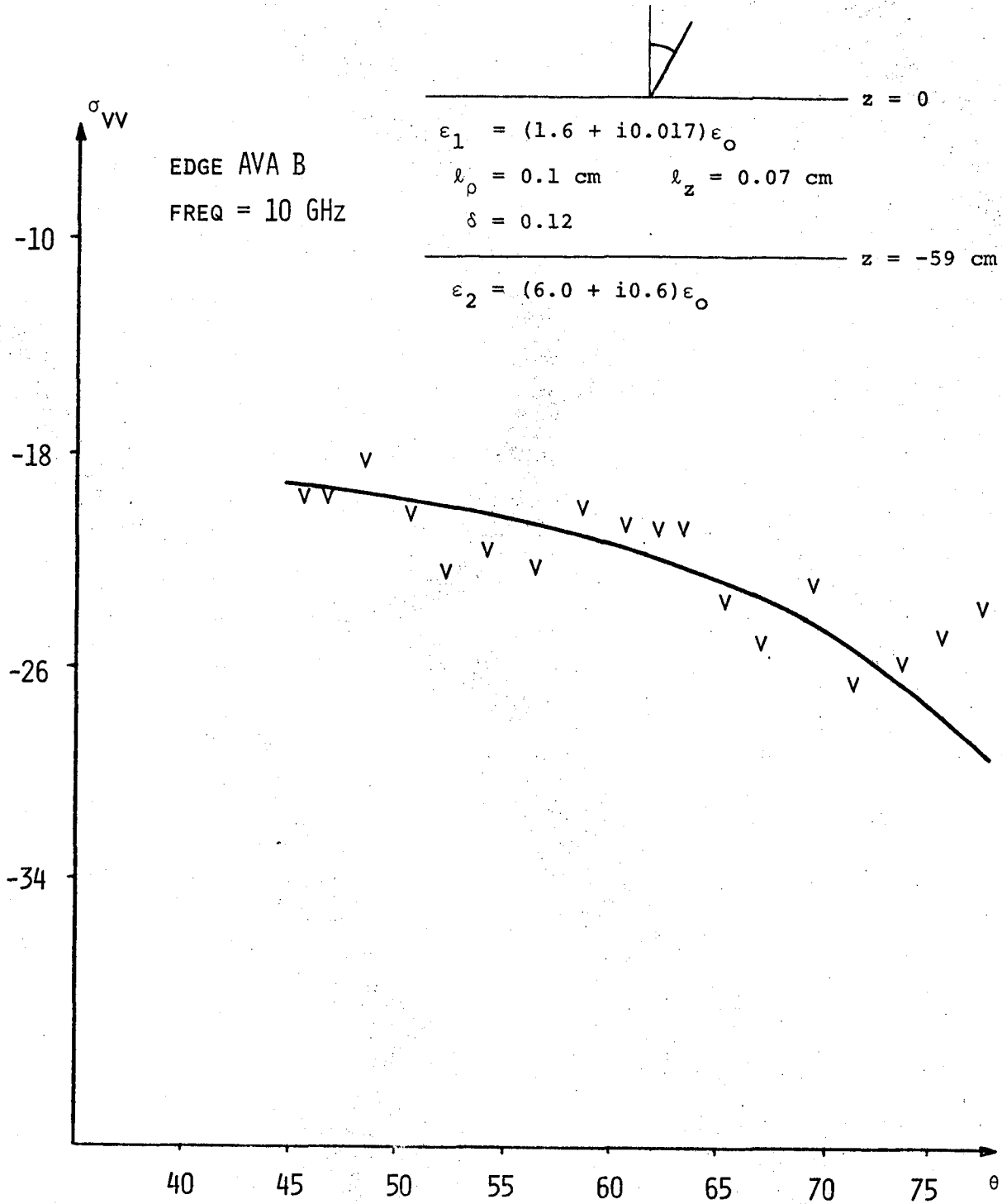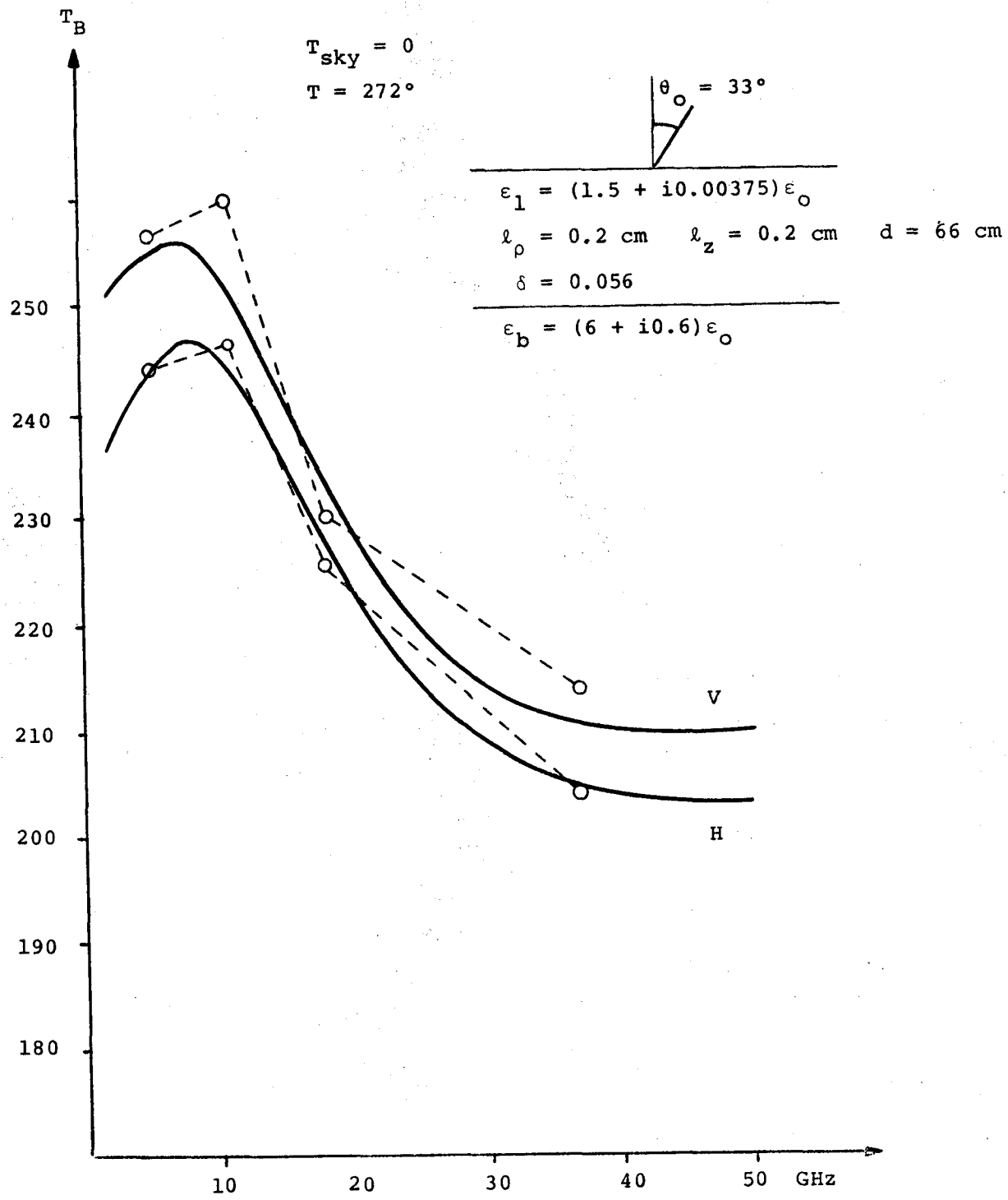
Figure 1

$$T_{sky} = 0$$
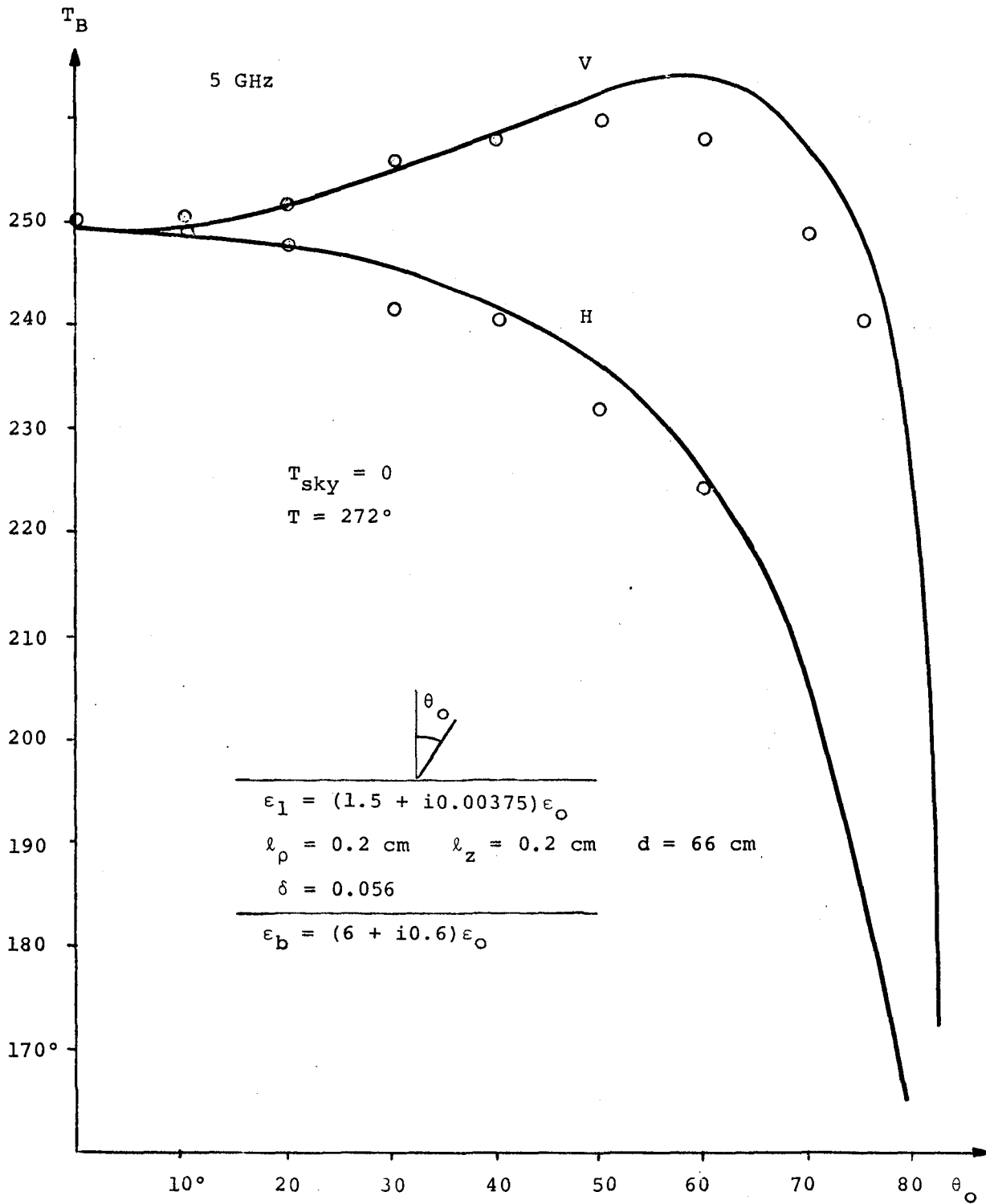$$T = 272°$$

$$\theta_o = 33°$$

$$\varepsilon_1 = (1.5 + i0.00375)\varepsilon_o$$
$$\ell_\rho = 0.2 \text{ cm} \qquad \ell_z = 0.2 \text{ cm} \qquad d = 66 \text{ cm}$$
$$\delta = 0.056$$
$$\varepsilon_b = (6 + i0.6)\varepsilon_o$$

Figure 2

Figure 3

160

Figure 4

Figure 5

18 GHz

T_B

$T_{sky} = 0$
$T = 272°$

θ_o

$\varepsilon_1 = (1.5 + i0.00375)\varepsilon_o$

$\ell_\rho = 0.2$ cm $\ell_z = 0.2$ cm $d = 66$ cm

$\delta = 0.056$

$\varepsilon_b = (6 + i0.6)\varepsilon_o$

H

V

230

220

210

200

190

180

170

10°   20   30   40   50   60   70   80 θ_o

Figure 6

163
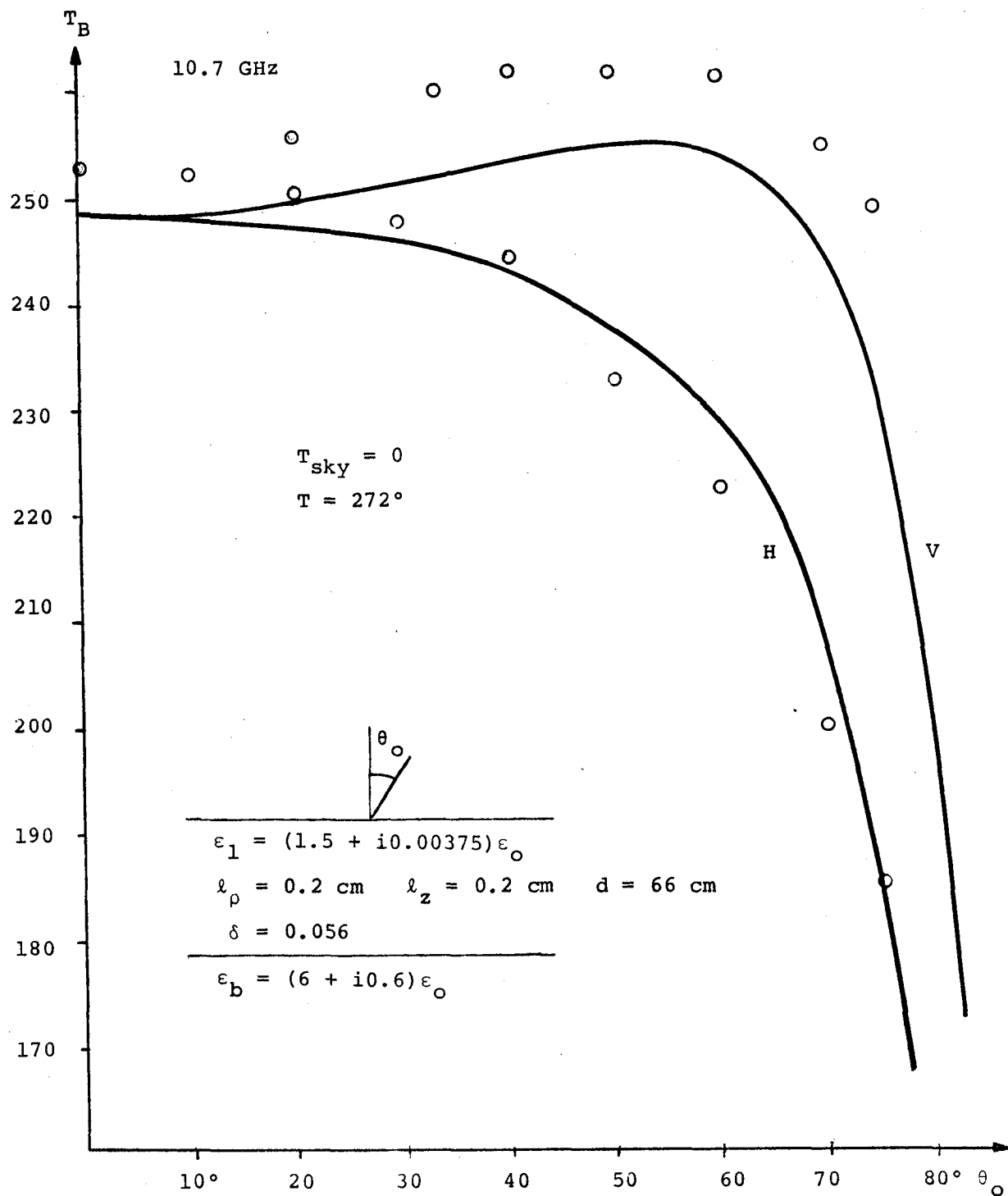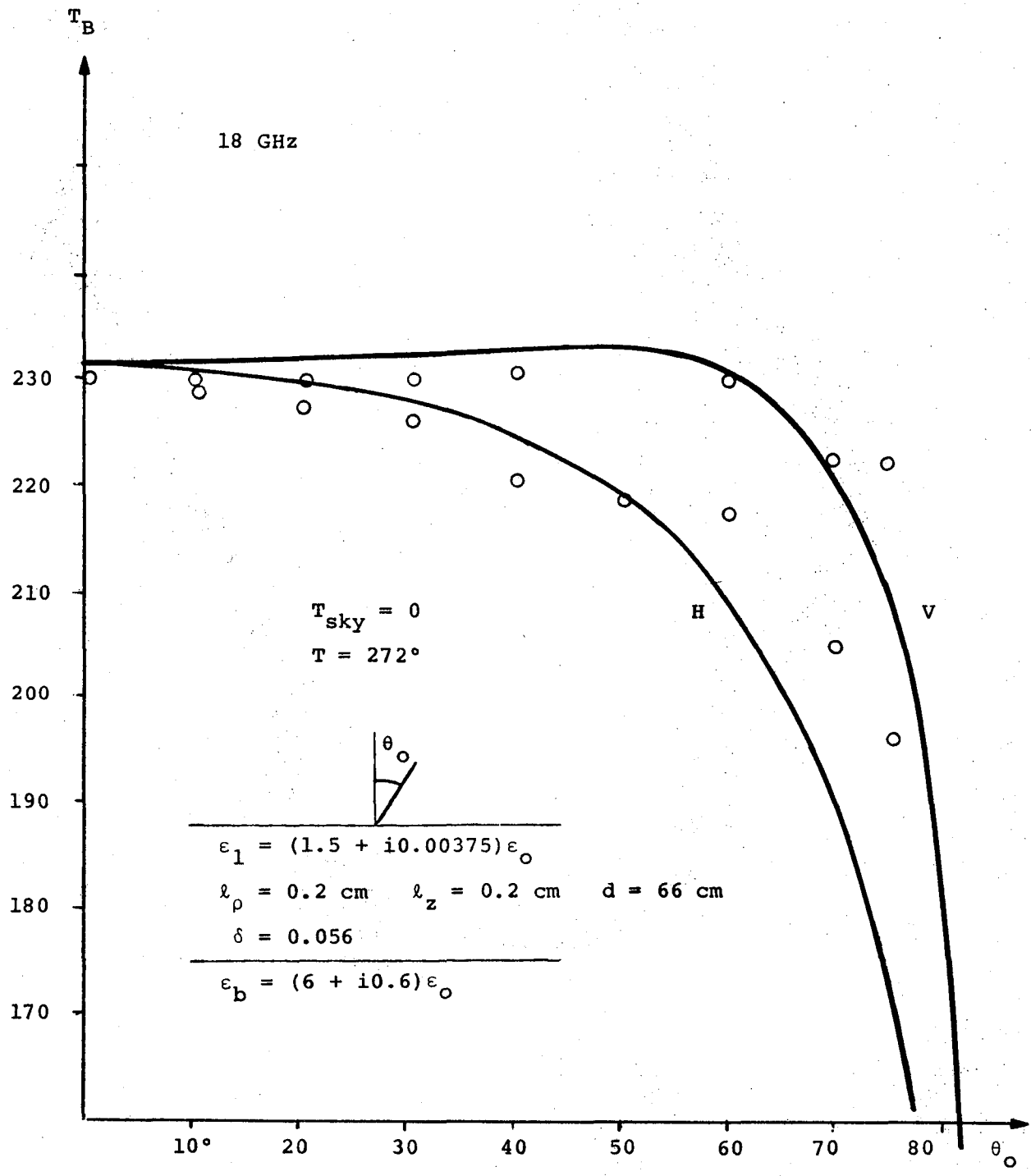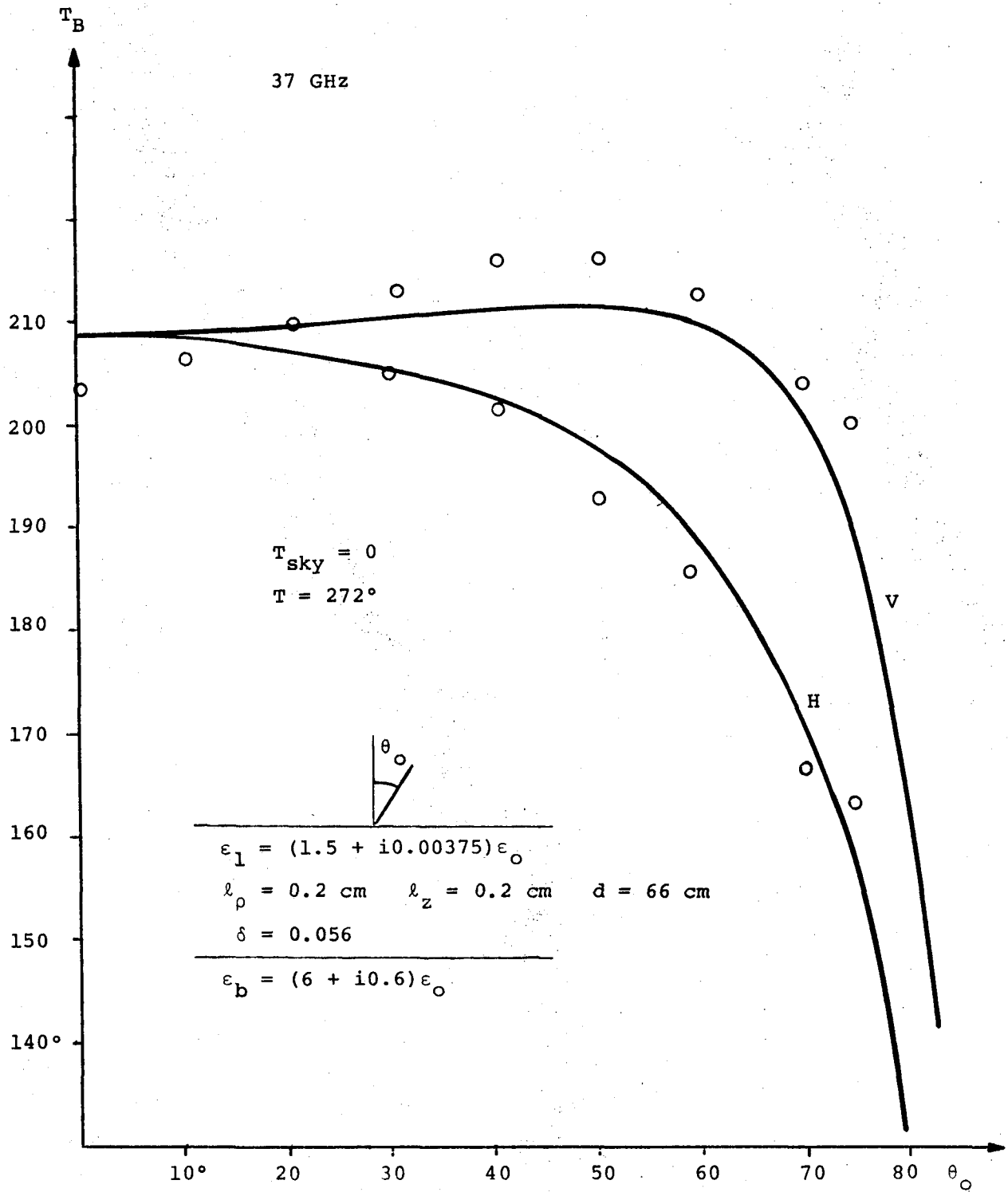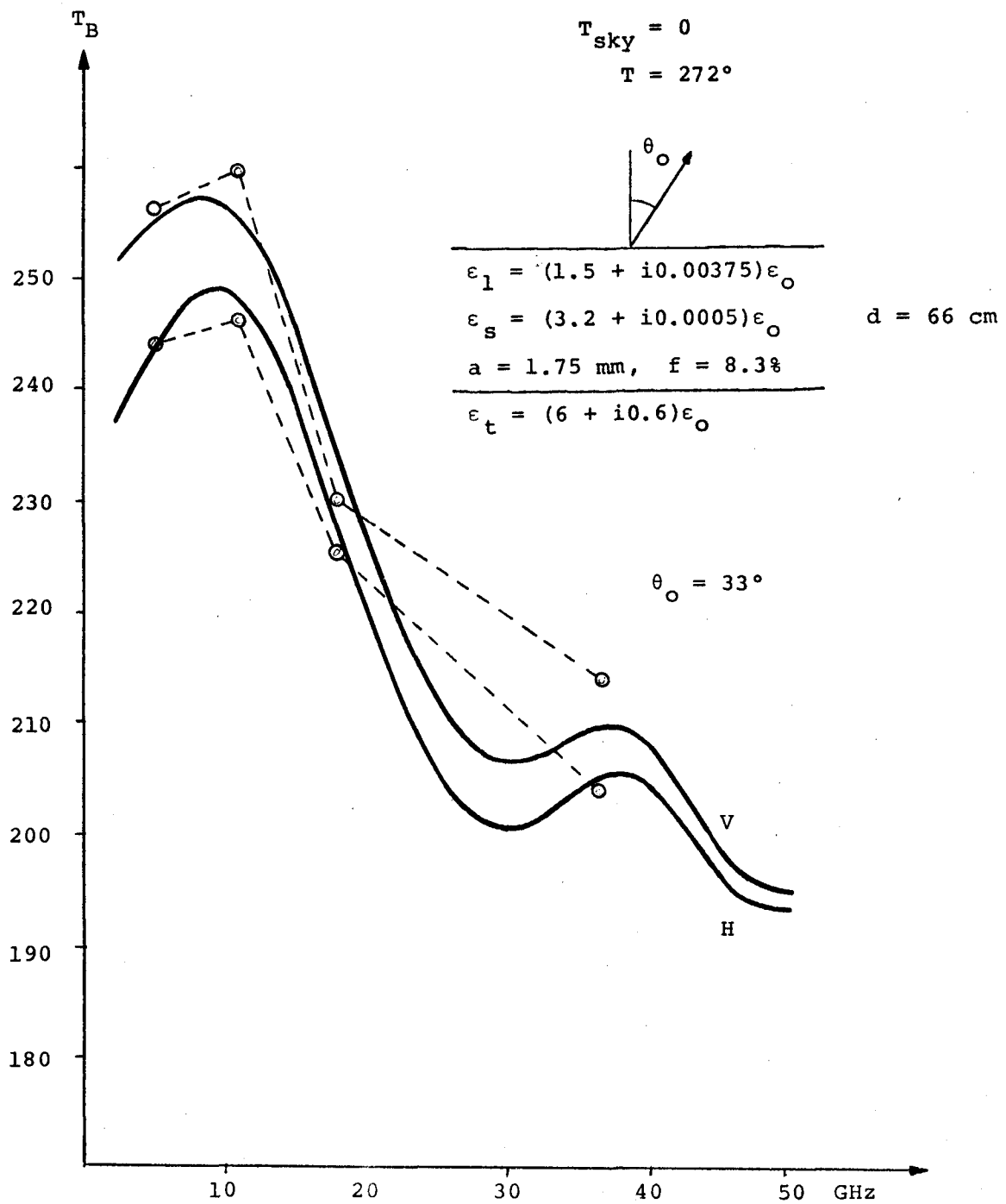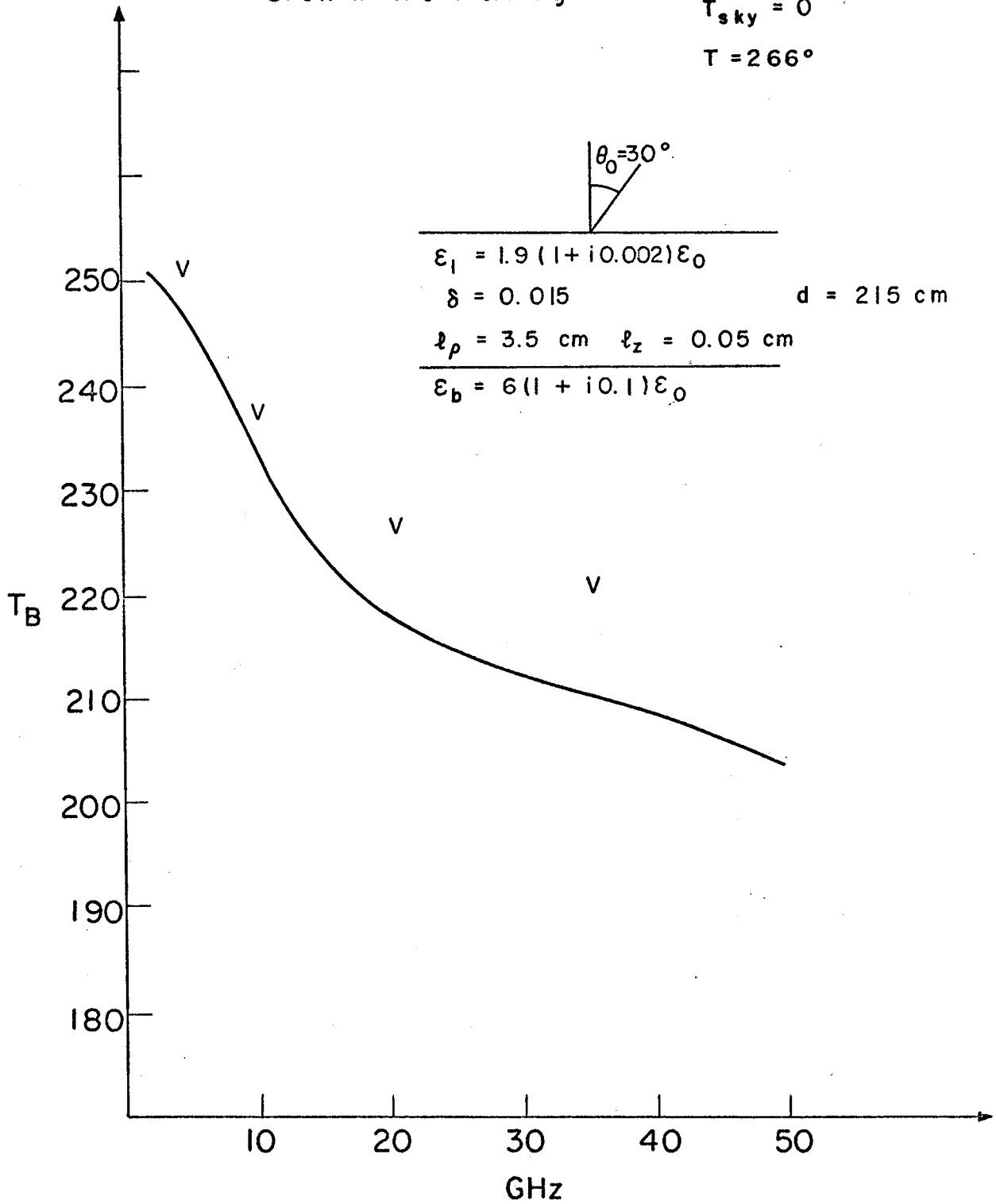
Figure 7

Figure 8

Snow in the afternoon

$\theta_0 = 30°$

$\varepsilon_1 = 1.9 (1 + i0.025) \varepsilon_0$ $d_1 = 7\,cm$

$\varepsilon_2 = 1.9 (1 + i0.002) \varepsilon_0$

$\ell_\rho = 3.5\ cm$ $\ell_z = 0.05\ cm$ $d_2 = 208\,cm$

$\delta = 0.015$

$\varepsilon_b = 6(1 + i0.1)\varepsilon_0$ $T_b = 268°$

$T_1 = 272°$

$T_2 = 268°$

$T_{sky} = 0$
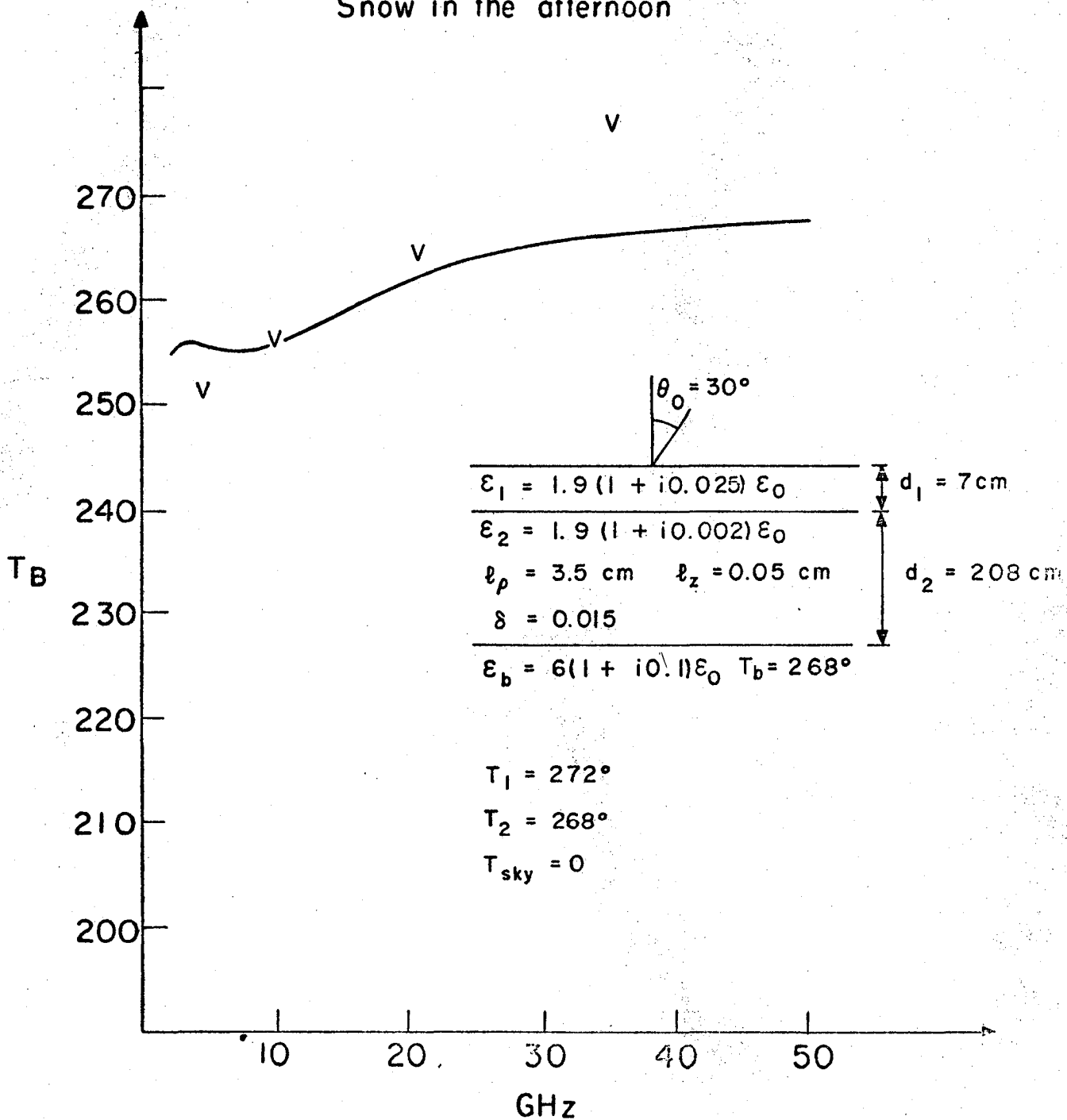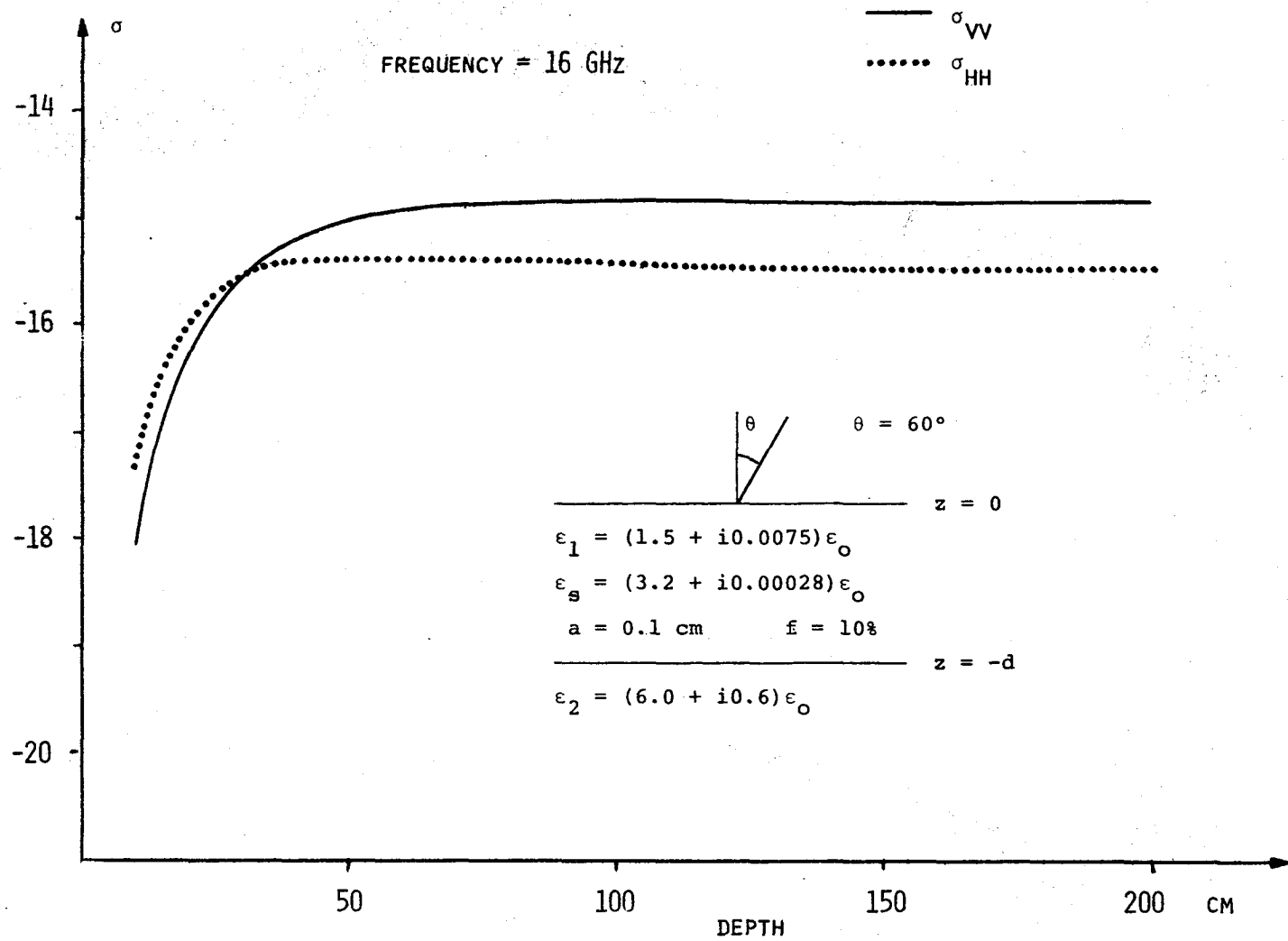
Figure 9

Figure 10

REMOTE SENSING OF SNOW PROPERTIES BY PASSIVE MICROWAVE

RADIOMETRY: GSFC TRUCK EXPERIMENT

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ABSTRACT

Recent results indicate that microwave radiometry has the potential
for inferring the snow depth and water equivalent information from snowpacks.
In order to assess this potential for determining the water equivalent of a
snowpack, it is necessary to understand the microwave emission and scattering
behavior of the snow at various wavelengths under carefully controlled condi-
tions.

Truck-mounted microwave instrumentation was used to study the microwave
characteristics of the snowpack in the Colorado Rocky Mountain region during
the winters of 1977-78 and 1978-79.   The spectral signatures of C, X, K_u, and
K_a band radiometers with dual polarization were used, together with measure-
ments of snowpack density, temperature and ram profiles, liquid water content,
and rough characterization of the crystal sizes.   These data compared favorably
with calculated results based on recent microscopic scattering models.

INTRODUCTION

Runoff from melting snow provides more than 65 percent of the streamflow
for most of the mountainous western United States.   Timely and accurate
prediction of the amount of runoff would allow more efficient management of
the scarce water supply for hydropower generation, irrigation, and domestic
and industrial water use.

In order to monitor the snow resources to predict runoff, it is necessary
to measure the water equivalent, liquid water content, and area covered by
snow.   In addition to the snow properties (e.g., temperature profile, density,
grain size), the condition of the underlying soil surface is especially
important for estimating the amount of snow-melt water that will reach the
stream channel as runoff.   Presently, data needed for runoff predictions are
obtained from in-situ measurements of snow depth, density, and water equivalent
along preselected snow courses.   Measurements using these methods are difficult
to obtain in severe weather conditions; hence, data for snowmelt models in
watershed runoff forecasting are frequently insufficient.

Remote sensing techniques may provide data more suitable to model calculation because of their capability of making measurements over the entire watershed area in a relatively short time period. Estimates of snow-covered areas from satellite-borne visible and infrared images for several test watersheds correlate well with the runoff yields (Rango et al., 1975). This seems to be a new and promising technique for runoff forecasts. However, the capabilities of the visible and infrared images are limited by the penetration of these short wavelengths through clouds and snowpacks.

Microwaves are largely unaffected by clouds and can penetrate through snow, the depth of penetration depending on the wavelength. Therefore, the development of a technique for the remote sensing of snow-water equivalent over large areas would seem feasible. Microwave sensing is one of the most promising techniques because of the volume scattering properties of snow grains at microwave wavelengths. Recent studies (Chang et al., 1976 and Chang and Gloersen, 1975) showed that multifrequency microwave measurements could be used to infer these interesting snow parameters.

A truck-mounted multifrequency microwave radiometer system (C, X, K, and K_a band) was used to conduct field experiments in the Rocky Mountains of Colorado during the winters of 1977-78 and 1978-79. The truck mobility allowed experimenters to move from one test site to another with relative ease and speed. This system was integrated by personnel of the National Aeronautics and Space Administration/Goddard Space Flight Center in cooperation with the National Bureau of Standards.

EXPERIMENTAL APPARATUS AND PROCEDURES

Four radiometers (C, X, K, and K_a band) were mounted on a metal-framed enclosure that could be controlled by an aerial lift. The truck-mounted hydraulically operated boom had a maximum length of 14 m when all three sections were fully extended. The boom could be moved in both elevation and azimuth. Also, the instrument unit could be rotated in an elevation plane. The rotation, coupled with the elevation movement and telescoping of the boom, allowed experimenters to vary the incidence angle and location for the measurements.

All four radiometers measured both vertically and horizontally polarized electromagnetic waves. The antennas were corrugated horns with low sidelobes. The nominal 3-dB beam widths were 6° except the C band unit, which was 15°. The radiometers were comparison or Dicke types with square waves for modulation and synchronous detection. The noise-equivalent brightness temperature (temperature sensitivity) was about 1 K with 0.1-sec integration time. In this experiment, more than 60 samples of brightness temperature measurements were taken and averaged for each single data point to improve measurement precision. The radiometers were calibrated internally every 16 seconds by switching the inputs from the antenna to a "hot" load and a liquid-nitrogen "cold" load. The hot-load temperature and the physical temperature of the antenna were also monitored. External calibration was achieved by viewing the clear sky and an ambient-temperature Eccosorb target.

Most brightness-temperature data were obtained by scanning the instrument unit. Two different types of scanning procedures were used in measuring the brightness temperature as a function of incidence angle. In "swath" scanning the radiometer antenna is scanned in a vertical plane gradually from nadir (normal incidence) until it is almost perpendicular to the nadir (Figure 1). Meanwhile, the antenna remains in a fixed position. Therefore, the antenna actually views at different spots along a radial "swath" as the incidence angle changes. Under this condition, any inhomogeneity of the snowfield could modify the characteristic of the angular dependence. In order to remove the potential field inhomogeneity effect, the experimenters scanned the radiometer antenna so that it always viewed the same "spot" while the incidence angle changed (Figure 2). In addition to the scanning measurement, several time-sequence measurements were made to study the diurnal effect on microwave snow signatures.

The instrument package was usually located about 5 meters above the snow surface so the reflection of the instrument package had virtually no effect on the measured brightness temperature. The influence of the reflected atmospheric brightness due to the water vapor and cloud liquid water on the measured brightness temperature was very small because of the high altitude of the test sites (about 2,700 meters above mean sea level) and the low air temperatures.

The physical characterization of the snowpack "ground truth" was also documented with the microwave measurements of snow density and temperature (Chang et al., 1979). The relative hardness and strength for each layer of snow was measured by a ram penetrometer, and visual inspections were made on the average grain size at various depths. The liquid water content was measured by centrifuge separation and freezing calorimetry.

The magnitude of the brightness temperatures for the four radiometers was determined by comparing the radiation received by the antenna with an internal "hot" load at about 310 K and a liquid nitrogen "cold" load at about 77 K. The calibration of each radiometer system was checked by aiming the antenna at targets whose brightness temperatures could be calculated. These targets were an Eccosorb absorber and the cold sky. The results of a typical check for the calibration targets are listed in Table 1. The calculated sky temperature included the effect of a dry atmosphere and used the snow temperature of 273 K (0°C), observed by the thermometer. The temperatures for the Eccosorb were those measured by the thermocouple inserted inside the microwave absorber. In this case the observed values generally differed less than 10 K from the calculated brightness temperature. For the range of brightness observed in this experiment, the error should be comparable to the difference shown in Table 1.

RADIOMETER EXPERIMENTAL RESULTS

The experiment was conducted at three test sites covering both shallow, uniform snowpacks in a valley, and deeper snowpacks in a high elevated mountain pass. Figure 3 shows the measured brightness temperatures T_B versus incidence angle for both horizontal and vertical polarization for a set of spot scan

data taken on February 16, 1978. Due to the leakage of liquid nitrogen from the calibration load dewar, measurements were made for incidence angles greater than 30 degrees. The air temperature was approximately -10°C when the test data were taken, hence, no liquid water was assumed to be present in the snowpack. The snow depth was 70 cm, which consisted of about 30 cm of new powdered snow and 40 cm of depth hoar which has a slightly larger crystalline structure. The metamorphosis for the bottom 10 cm was more advanced and the ram-hardness measurement increased from approximately 0 to 10 kg. No noticeable ice layer was observed within the snowpack. The under-lying ground surface was frozen soil sparsely covered with dried-up stalk cover. Under these conditions, the brightness temperature contribution from the ground surface was closely related to each of the four test frequencies. The C, X, and K band measured brightness readings were closely related, and these readings strongly suggest that the scattering effect of snow is relatively small for these frequencies. The brightness temperature of K_a band was about 40 K lower than the other frequencies. This difference showed that the scattering effect is a dominant factor affecting the measured brightness temperature at this frequency.

The measurement set of January 10, 1979 was carried out with air temperatures at -6°C (see Figure 4). The snow depth was about 60 cm, with snow condition similar to the previous case. The depth hoar layer was 25 cm as compared with 40 cm the previous year. The measured X and K band brightness temperatures closely related to the previous measurements, while the K_a band showed considerable difference because of the depth hoar layer thickness. The C band results were quite different for the two years. More study will be conducted to pinpoint the problem area.

The data set of March 23, 1978 represented a measurement for a very wet snow case as active melting was taking place in the isothermal snowpack (see Figure 5). The liquid water content was about 15 to 20 percent by weight. Figure 5 shows the measured brightness temperatures at C, X, and K_a band. The brightness temperatures of K_a band show a slight angular variation for a vertical polarization and horizontal polarization for an incidence angle between 0° and 50°. The brightness temperature of the snow was nearly identical to its physical temperature.

RADIATIVE TRANSFER EQUATION ANALYSIS

The microwave radiation emitted from a snowpack is dependent on the physical temperature, crystal size, and density of the snowpack. The basic relationship between the properties of the snowpack and the emitted radiation can be derived by using the radiative transfer approach.

An insight into the microwave emission from snow fields has been provided by a macroscopic volume scattering model by England (1974). This model specifically involves a parameter called the volume scattering albedo, ω_0, which is the ratio of the volume scattering coefficient to the total extinction coefficient. The extinction coefficient includes both the resistive and scattering losses. The analysis involves a value for ω_0 which is used to compute the brightness temperature or emissivity based on the ω_0 parameter.

The model may be used to calculate the emissivity or the brightness temperature of finite slabs of snow and ice with varying compositions.

A snow particle scattering model was developed by Chang, et al. (1976) using the microscopic approach. This model assumed that the snow field or snow cover consisted of randomly spaced scattering spheres which did not scatter coherently. Since the snow fields of interest generally consisted of nonspherical particles which were not well separated, two assumptions were required to apply the theory. Firstly, it was assumed that the scattering particles were spherical; secondly, that the particles scattered incoherently and independently of the path length between scatters. These assumptions, however, were not expected to influence the quantitative nature of the test results. The Mie theory was then used to calculate the extinction and scattering cross sections of the individual particles as a function of particle radius and the complex index of refraction for given wavelengths. Subsequently, these quantities were used to solve the radiative transfer equation within the snow medium and to calculate the radiative emission from the model snowfield surface.

For the case of the melting ice sphere, it was assumed that the sphere consisted of a central core of ice and a surrounding shell of water. The solution of scattering of electromagnetic waves from these concentric spheres were solved by Aden and Kerker (1951). In this study, the thickness of the water layer is set according to the measured liquid water content. The index of refraction for water is calculated according to the results of Lane and Saxton (1952). The refractive index of ice is taken to be 1.78 + 0.0024 for this study (Cumming, 1952).

The radiative transfer equation for an axially symmetric inhomogeneous medium in which all interactions are linear can be written in the form of an integro-differential equation as stipulated by Grant and Hunt (1969):

$$\mu \frac{dI\,(x,\mu)}{dx} = -\sigma(x)\,I\,(x,\mu) + \sigma(x) \left\{ [1 - \omega(x)]\,B(x) \right.$$

$$\left. + \frac{1}{2}\omega(x) \int_{-1}^{1} p(x,\mu,\mu')\,I\,(x,\mu')\,d\mu' \right\} \tag{1}$$

where the radiation intensity $I(x,\mu)$ is at depth x traveling in the direction making an angle whose cosine is μ with the normal toward the direction of increasing x.

The functions $\sigma(x)$, $\omega(x)$, $B(x)$, and $p(x,\mu,\mu')$ are prescribed functions of their arguments. They are referred to as the extinction per unit length, the single scattering albedo, the source, and the phase function, respectively. For a nonuniform medium these functions are generally piecewise continuous functions of depth subject to the conditions.

$$B(x) \geqslant 0, \quad \sigma(x) \geqslant 0, \quad 0 \leqslant \omega(x) \leqslant 1, \quad p(x,\mu,\mu') \geqslant 0. \tag{2}$$

In the present work, the following normalization for the phase function will be used:

$$\frac{1}{2} \int_{-1}^{1} p\,(x,\mu,\mu')\ d\mu' = 1 \tag{3}$$

for all values of x. Instead of working with depth x, one generally works with a dimensionless depth variable called optical depth τ, defined in differential form as:

$$d\tau = \sigma(x)\ dx. \tag{4}$$

In terms of optical depth, equation (1) reduces to:

$$\mu \frac{dI(\tau,\mu)}{d\tau} = -I(\tau,\mu) + [1 - \omega(x)]\ B(x)$$

$$+\frac{1}{2}\ \omega(\tau)\int_{-1}^{1} p(\tau,\mu,\mu')\ d\mu I(\tau,\mu'). \tag{5}$$

The equation of radiative transfer was solved numerically by the invariant imbedding technique by Chang and Choudhury (1978). By using the Mie scattering phase function and the boundary conditions, the brightness temperature emerging from the snowfield can be calculated. To calculate the microwave radiation emitted from the snowpack, it is necessary to know its physical temperature, snow density, and mean crystal radius within the snowpack. The calculated brightness for all four bands based on the measured snow parameters are shown in Figures 6, 7, and 8.

DISCUSSIONS

No obvious ice layer was detected within the snowpack during the test period in the vicinity of Fraser, Colorado. The measured brightness data corresponded with the smooth curves for the vertical and horizontal polarized data. These curves correspond to the Fresnel reflection characteristics for a dielectric media interface. The volume scattering effect became a dominant factor affecting the brightness temperature when higher frequencies were used which decreased the penetration depth. At K_a band (0.8cm), the scattering effect caused a decrease of 50 K and 20 K in the emerging brightness temperature for the first and second years over the entire measuring angular range for dry snowpack. The comprehensive change in the brightness temperature provides an opportunity to deduce the mean crystal radius within a snowpack by microwave measurement.

By comparing Figures 3 through 8, the calculated brightness temperatures generally agreed with the measured values. It was obvious that the scattering model simulates the real scattering behavior of the snow cover for the test

site near Fraser, Colorado. The fine structure variation within the snowpack, which was not included in the scattering model calculation, would have contributed to the deviation between the measurements and the calculated brightness temperatures. The input to the scattering model may be improved by collecting more information on the snowpack properties.

The presence of melt water in the snowpack drastically changes the microwave emission characteristics, resulting in as much as 50 K increases in the brightness temperature over a dry cross section snow condition because the extinction cross section for melting snow is an order of magnitude larger than that for dry snow. For a detailed quantitative analysis of snow wetness, new in situ techniques should be developed for measuring the snow parameters quickly and simultaneously with the microwave measurements.

SUMMARY AND CONCLUSIONS

The primary goal of the experiment was to study the feasibility of measuring snow depth, density, and nonuniform vertical grain-size distribution of a snowpack by multichannel microwave measurement. In the past year, snowfield data were collected at three different test sites for a total of approximately 30 days. In the second year data were collected for about 16 days.

Calibration of the radiometric data and compilation of all the ground truth data have been completed. Although the detailed analysis of these data has not been completed, our understanding of the microwave signatures of snowfields and how these signatures can be exploited to give us information we seek has been enhanced considerably.

By reviewing the radiometric data, it is obvious that those brightness-temperature data taken before any significant melting occurred match quite well with the calculated results from a microscopic scattering model (Chang et al., 1976). However, after the snowpack underwent freeze/thaw cycles, the measured brightness generally did not match very well with the results of the simple model. This is probably due to the layering effect generated by the refreezing of free water within the snowpack. In order to remove possible ambiguities caused by this effect, multichannel data are required to retrieve those snow parameters which are pertinent to the runoff-model prediction.

The measured 37-GHz brightness temperatures showed considerable effect of volume scattering by the snow grain. This effect is much less distinct for the 5-GHz when comparing the brightness temperature for a natural pack and a wind-drift pack. The 37-GHz brightness temperature for a wind-drift pack is generally about 40 K higher than the naturally compacted snowpack. We attributed the warmer brightness temperature of the wind-drift pack to its smaller average grain size (0.5 mm), which scatters less.

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The brightness temperature changed drastically when a small amount of liquid water existed within the snowpack. The K_a band brightness changes first because at this frequency most of the microwave radiation was emitted from a thin layer of snow exposed to the warmer snow-air interface. This increase of brightness temperature can be explained qualitatively by the calculations made by Chang and Gloersen (1975). Where there is more than 10 percent liquid water by weight within the snowpack, the measured brightness temperatures for all four frequencies were very close to the physical temperature of the snowpack. This characteristic may be used to determine the onset of melting of the snowpack.

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TABLE 1. — COMPARISONS OF OBSERVED AND CALCULATED BRIGHTNESS TEMPERATURES

| Target | | | | Brightness Temperature (K) | |
| | | | | Calculated | Difference (observed – calculated) |
		Observed			
Sky	5	V	2.6	4.9	-2.3
		H	5.4		0.5
	10	V	2.9	5.2	-2.3
		H	4.5		-0.7
	18	V	10.0	6.9	3.1
		H	10.0		3.1
	37	V	9.6	12.3	-2.7
		H	24.6		12.3
Eccosorb	5	V	288.8	288.6	0.2
		H	289.1	288.6	0.5
	10	V	287.1	288.1	-1.0
		H	286.8	288.1	-1.3
	18	V	279.8	288.1	-8.3
		H	284.3	288.1	-3.8
	37	V	285.1	288.1	-3.0
		H	287.4	288.1	-0.7
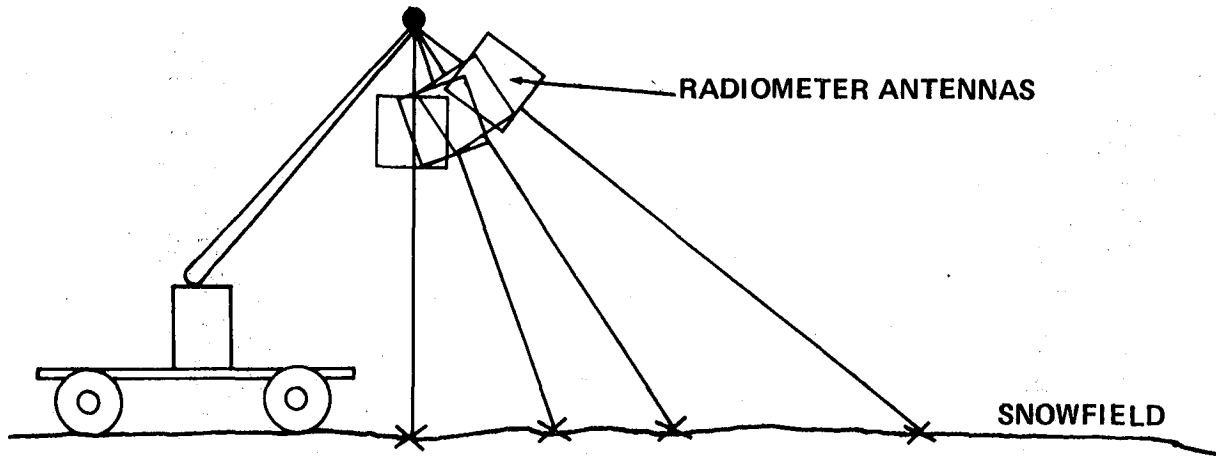
Figure 1. Swath Scan — The instrument package is directed to new test
spots as the incidence angle is changed.
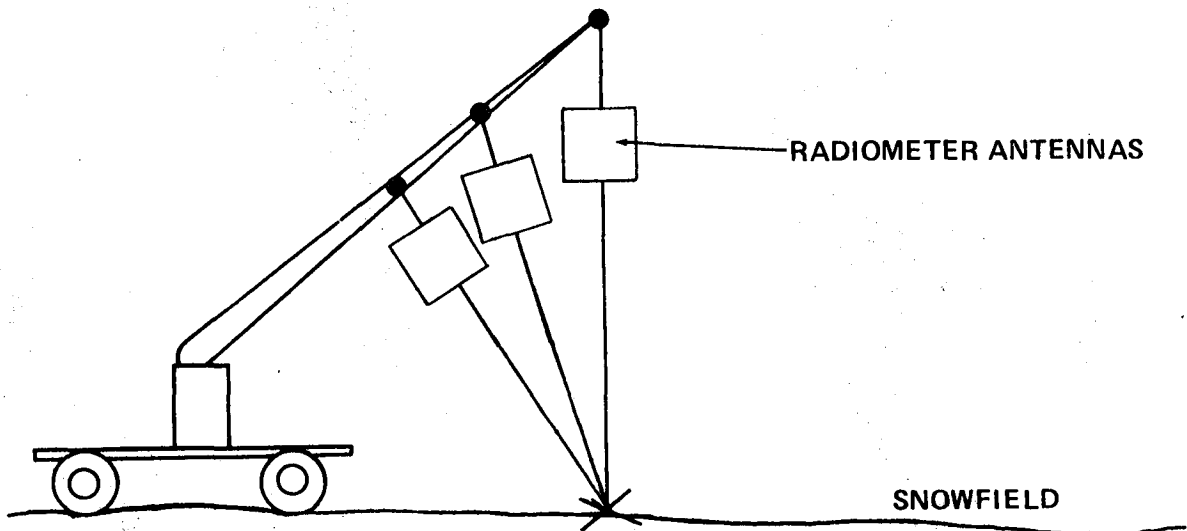
Figure 2. Spot Scan — The instrument package is always directed to the
same test spot while the boom's height is changed.
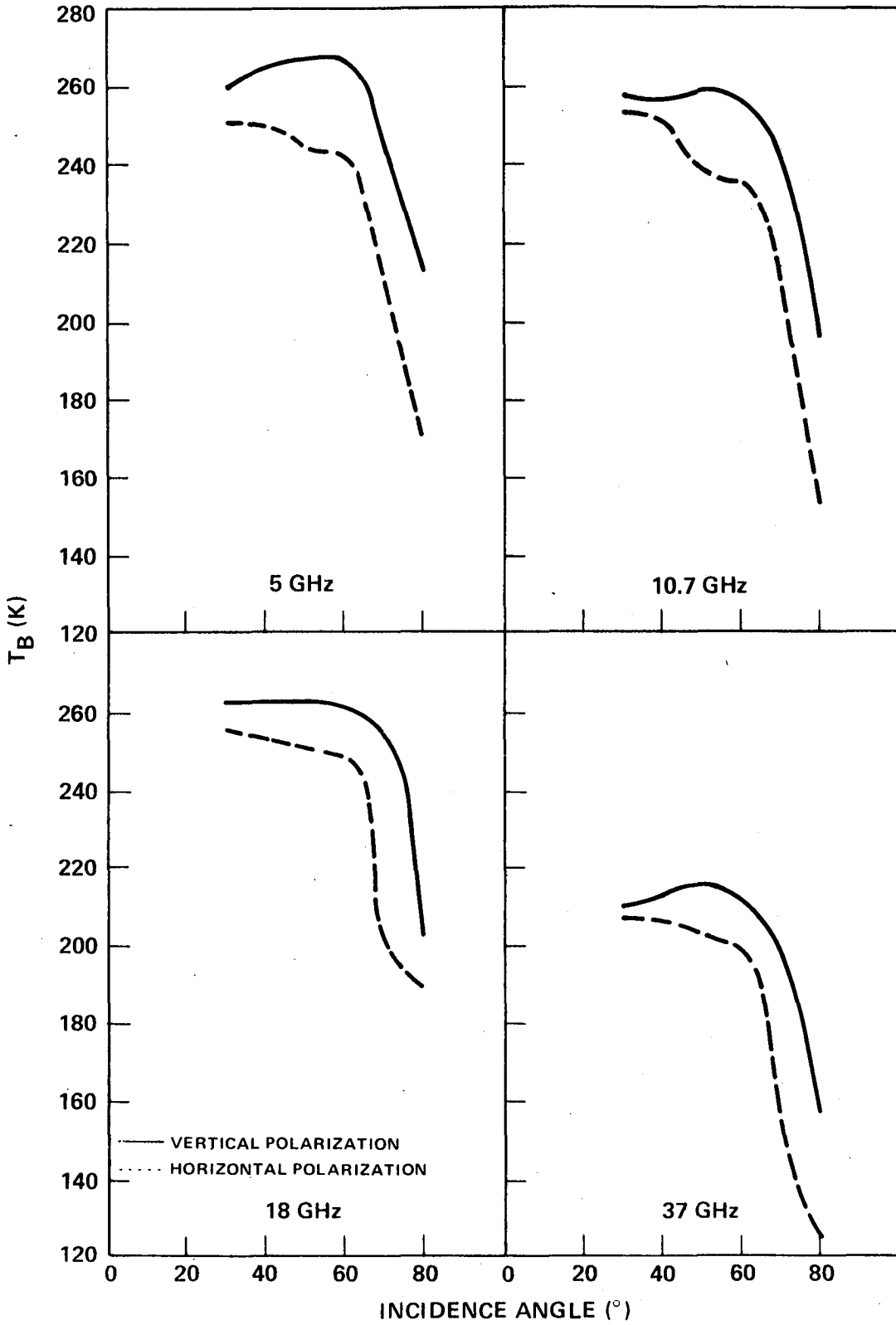
Figure 3.   February 16, 1978 — Brightness Temperature T_B Versus
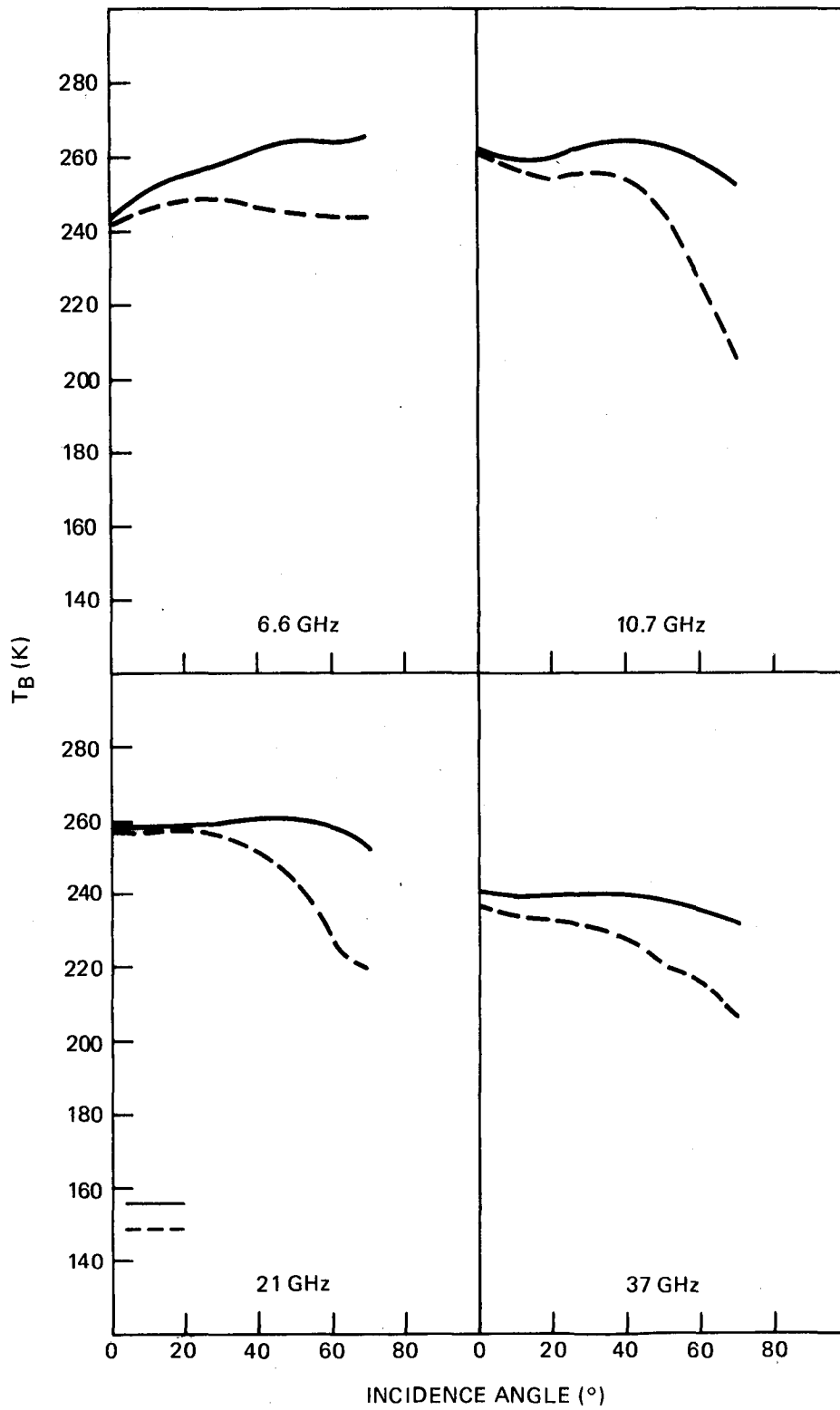Incidence Angle (Fraser, Colorado).

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Figure 4. January 10, 1979 — Brightness Temperature T_B
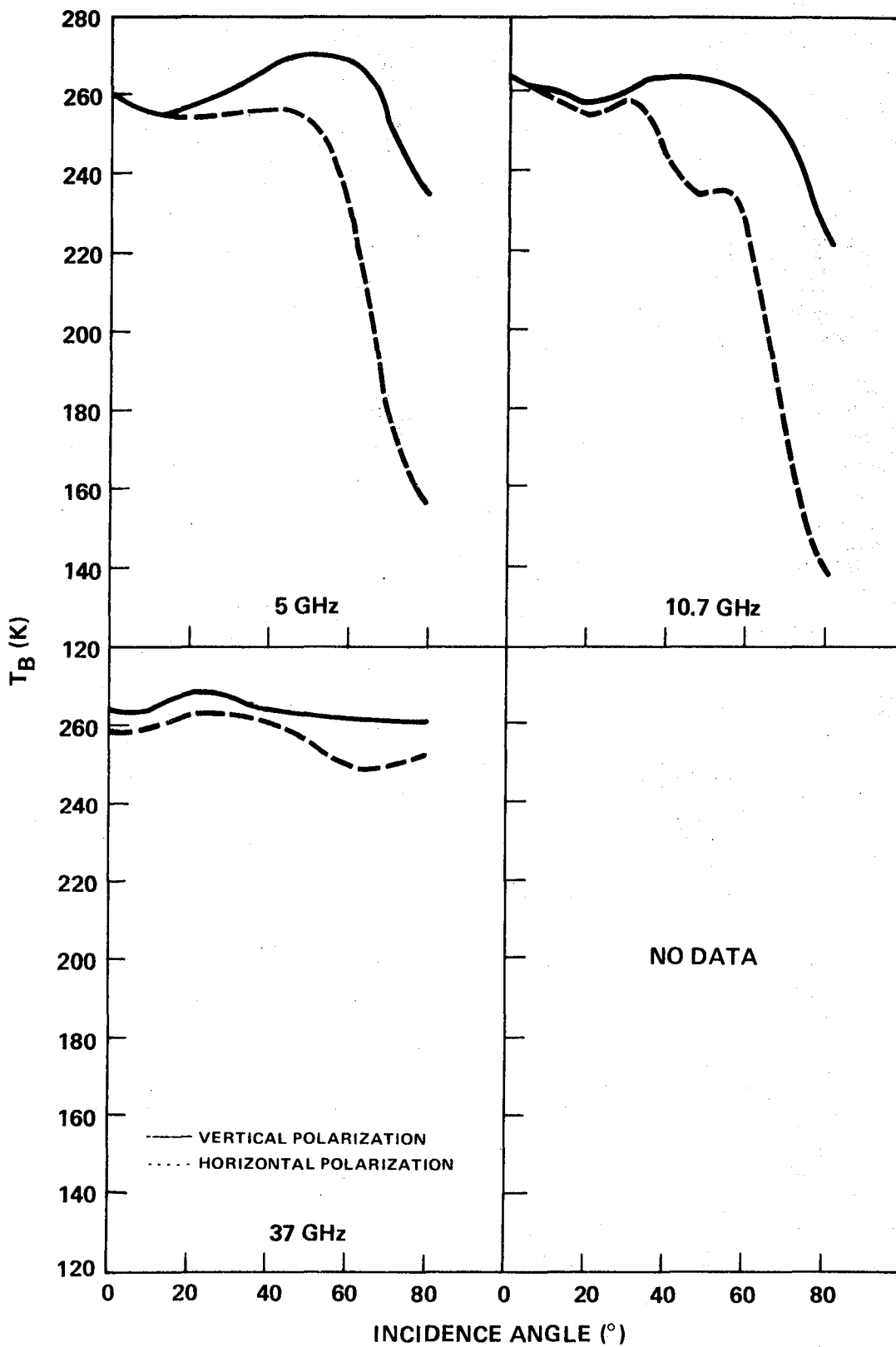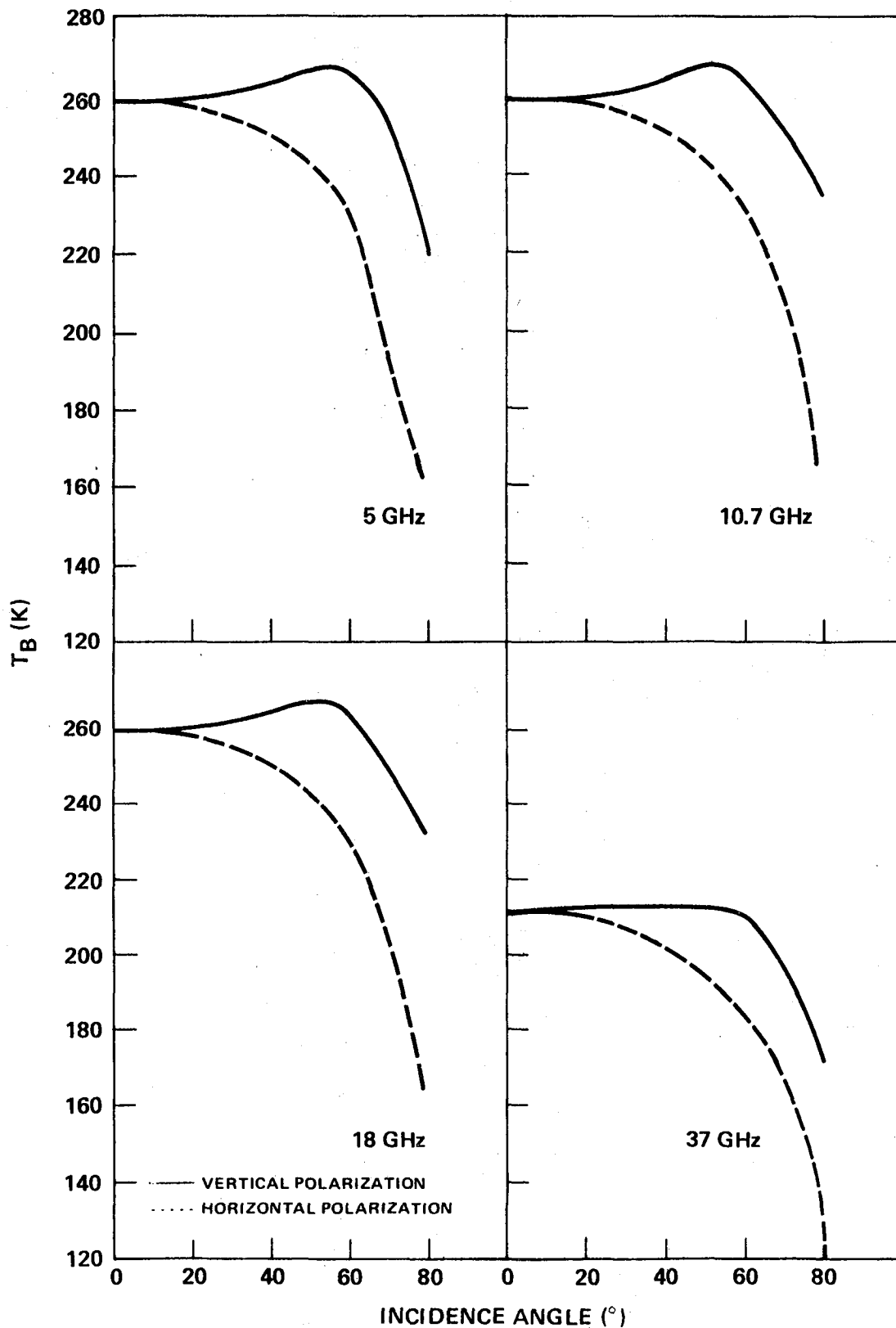Versus Incidence Angles (Fraser, Colorado).

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Figure 5. March 23, 1978 — Brightness Temperature T_B Versus
Incidence Angle (Fraser, Colorado).

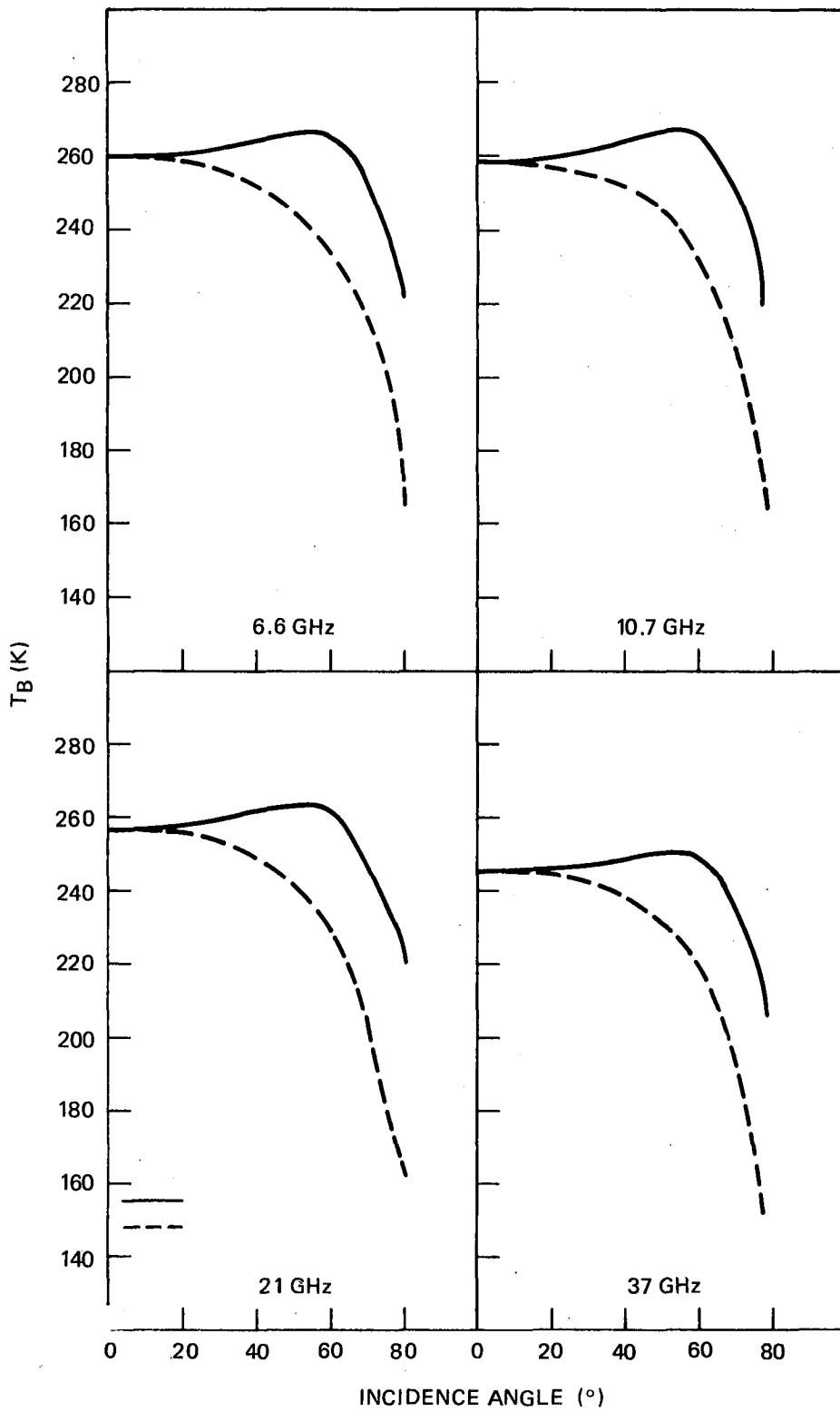Figure 6. Dry Snow — Calculated Brightness Temperature T_B Versus Incidence Angle.

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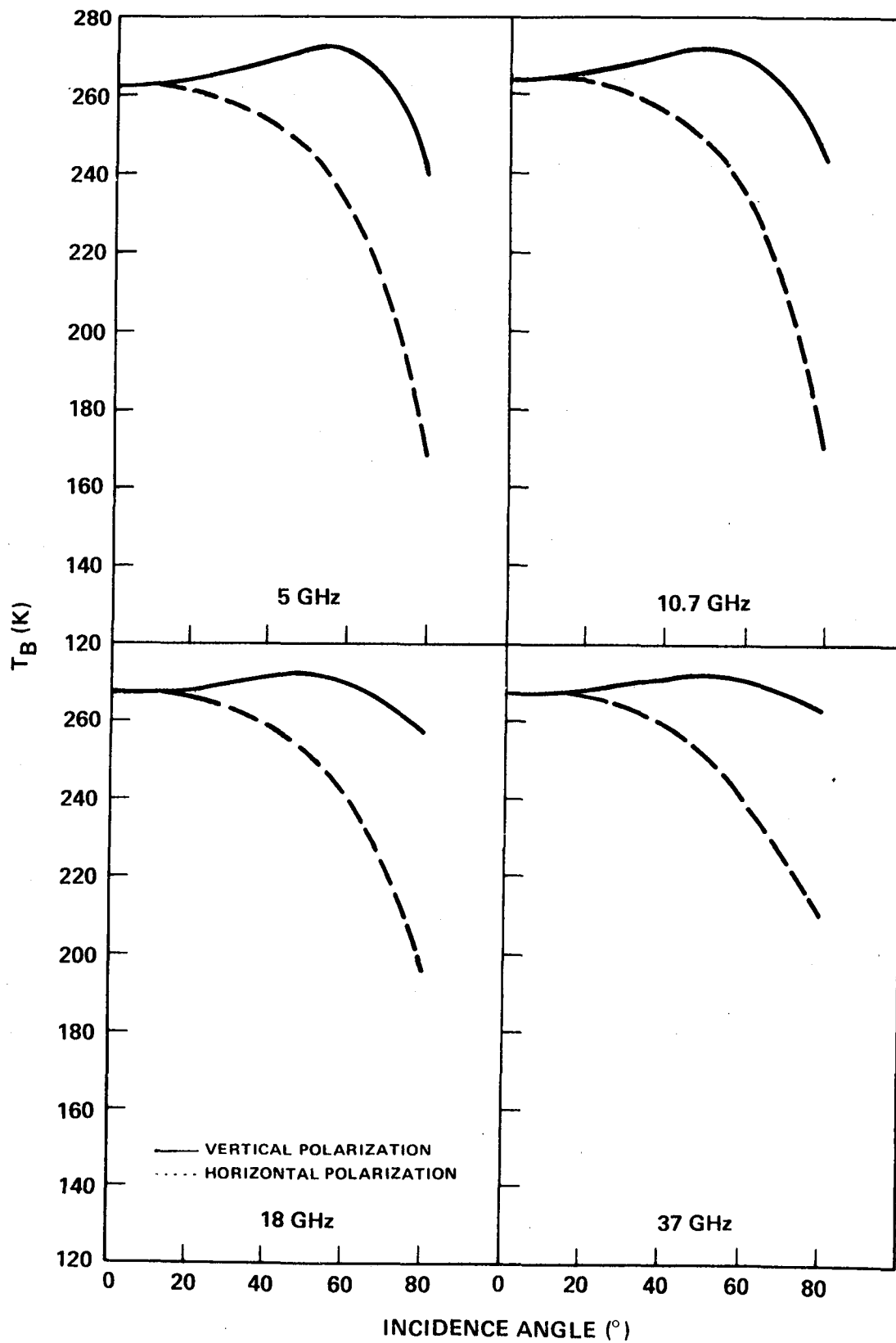Figure 7. Dry Snow Calculated Brightness Temperature T_e
Versus Incidence Angle.

Figure 8. Wet Snow — Calculated Brightness Temperature T_B Versus Incidence Angle.

MICROWAVE RADIOMETRIC OBSERVATIONS OF SNOWPACKS*

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ABSTRACT

Models for the microwave emission from snowpacks were generated on the basis of radiometric observations made at 10.7 GHz, 37 GHz and 94 GHz at a test site near Steamboat Springs, Colorado. In addition to conducting measurements on an approximately daily basis over a six-week observation period, measurements were made over several diurnal cycles during which the change in snow wetness was tracked by the microwave radiometers. Also, the variation in emissivity with snow water equivalent was examined, as was the sensitivity to changes in snow surface geometry. The microwave emissivity was observed to (a) decrease exponentially with snow water equivalent and (b) increase with snow wetness. Thus, the emission behavior is the reverse of the backscattering behavior observed by the radar. By fitting the models to the measured data, the variation of the optical depth with snow wetness was estimated.

INTRODUCTION

The paper by Stiles and Ulaby in this workshop Proceedings [1] provided a summary of the radar backscatter investigations that were conducted during the past four years. In one of these investigations (Colorado, 1977), microwave radiometric measurements also were made, simultaneously with the radar reflectivity measurements. Experimental investigations of the microwave radiometric response to snowpack parameters, specifically water equivalent W and wetness m_v, were conducted over a decade ago [2,3] by the Aerojet General Corporation. The observed behavior of the microwave emission from snow-covered terrain, characterized by the apparent temperature T_{ap}, has since been confirmed and extended by numerous ground-based experiments [4-11], airborne investigations [12,13] and satellite observations [14-15]. The key features of the present study are: (a) the observation period lasted about six weeks, during which time several diurnal experiments were conducted, (b) it is the only study conducted to date in which active and passive microwave data concerning snow were acquired simultaneously, which provides the opportunity for evaluating the potential advantages derived from the combined use of active and passive microwave sensors for estimating snow parameters of interest and (c) it provides a powerful tool for evaluating the applicability of theoretical models by simultaneously fitting radar reflectivity and radiometric emission models (based on the same basic assumptions about the snow volume and its air-snow and ground-snow interfaces) to the measured radar and radiometer data.

The data reported in this paper were acquired by radiometers operating at 10.69 GHz, 37 GHz and 94 GHz. Calibration was achieved through observations of the sky, representing a "cold" temperature source, and of microwave absorbing material, representing a perfect emitter. In addition to the acquisition of standard data sets consisting of radiometric temperature measurements as a function of angle between 0° (nadir) and 70°, several diurnal experiments at a select set of angles also were conducted. Further, the variation of apparent temperature with snow depth was investigated. In support of the microwave measurements, the ground-truth parameters listed in Table 1 of the previous paper [1] were monitored. More detailed information on ground-truth measurement techniques and radiometer system specifications is available in previous papers and reports [9-11].

THE APPARENT TEMPERATURE OF SNOW

The brightness temperature T_b of a snow layer over a ground surface characterizes the upward emission by the layer and the underlying ground. T_b is governed by the dielectric and geometrical characteristics of the snow volume and its interfaces and by the thermometric temperature profiles of the snow layer and the underlying medium. Since for a given condition, the total variation in the magnitude of the thermometric temperature between the ground surface and the snow surface is only a few degrees Kelvin, for modeling purposes it may be assumed to be a constant. For a constant snow (and ground) thermometric temperature T_o, the brightness is given by

$$T_b = \varepsilon\, T_o \tag{1}$$

where ε is the emissivity of the snowpack scene, which includes contributions by the snow layer as well as, in the general case, contributions by the underlying ground medium. The apparent temperature T_{ap} measured by a ground-based radiometer is related to T_b through

$$T_{ap} = T_b + r\, T_{sky} \tag{2}$$

where r is the effective reflectivity of the snowpack and T_{sky} is the sky radiometric temperature representing downwelling atmospheric radiation. As a first-order approximation, $r \simeq 1 - \varepsilon$, in which case T_{ap} becomes

$$T_{ap} = \varepsilon\, T_o + (1 - \varepsilon)\, T_{sky} \tag{3}$$

At a given microwave frequency f and for a given angle of incidence θ, T_{sky} may be calculated using atmospheric emission models or measured directly by pointing the radiometer towards the sky at the angle θ from the vertical. At 10.69 GHz, the magnitude of T_{sky} is only a few degrees Kelvin and the maximum observed value of $(1 - \varepsilon)$ is 0.2. Thus, the contribution of the second term in Equation (3) is negligible in comparison to the first term, and

therefore, it can be neglected (at 10.69 GHz). At the higher frequencies of 37 GHz and 94 GHz, the influence of T_{sky} on T_{ap} may be ignored in qualitative analyses of the T_{ap} response to snow parameters, but should be accounted for in comparisons of theory to measured values.

The emissivity of a homogeneous snow layer of depth h and density ρ (and therefore of water equivalent W) may be modeled by the approximate expression:

$$\epsilon = T_{sa} \left[(1 - \omega) \left(1 - \frac{1}{L} \right) + \frac{T_{gs}}{L} \right] \tag{4}$$

where T_{sa} and T_{gs} are respectively the transmission coefficients of the snow-air and ground-snow interfaces, ω is the snow albedo and L is its one-way loss factor,

$$\omega = \kappa_s' / \kappa_e' \tag{5}$$

$$L = \exp(\kappa_e' \, W \, \sec\theta') \tag{6}$$

κ_s' and κ_e' are respectively the volume mass scattering and extinction coefficients and θ' is the angle of refraction in the snow medium. At nadir, the magnitudes of T_{sa} and T_{gs} vary between about 0.90 and 0.98 depending on the wetness of the snowpack and the moisture content of the underlying soil medium. T_{gs} exhibits a weak angular dependence because the local angle of incidence at the snow-soil interface is smaller than the observation angle of incidence due to refraction at the snow-air interface.

In general, L increases with increasing microwave frequency f, snow wetness m_v, snow depth h and the angle of incidence θ. The albedo ω increases with increasing frequency f and decreases with increasing snow wetness m_v. Evaluation of the above model is discussed in later sections.

APPARENT TEMPERATURE RESPONSE TO SNOW WETNESS

The angular behavior of the passive microwave data is shown in Figure 1 for wet and dry snow conditions. At 10.69 GHz, the apparent radiometric temperature T_{ap} shows similar angular shapes for the wet and dry cases. This difference in T_{ap} between wet and dry snow conditions is very small compared with the large change (>100K) at 37 GHz. Similar behavior was noted in other investigations [3,4,5,7].

The effects of diurnal variations in snowpack parameters on T_{ap} were examined by acquiring data over four diurnal cycles. The general behavior was qualitatively similar for all four diurnal experiments [9], and therefore only representative examples are included here. Figures 2a and 2b show the diurnal variation of T_{ap} for nadir and 50°, respectively. The corresponding diurnal variation of the snow wetness m_v (of the top 5 cm snow layer)

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is shown in Figure 2b. As might be expected on the basis of the behavior noted earlier for Figure 1, the 37 GHz radiometer measured a much larger change in T_{ap} (in response to m_v) than did the 10.69 GHz radiometer. As m_v increases, the loss factor L increases and the albedo ω decreases. Since one of the terms in Equation (4) is proportional to $(1 - \omega)$, ε increases with increasing m_v. In contrast, the radar reflectivity is directly proportional to ω [1] and therefore σ^o decreases with increasing m_v.

The results of a similar diurnal experiment, which included measurements at 10.69 GHz, 37 GHz and 94 GHz, are shown in Figure 3. Of particular note is the rapid change in the magnitude of T_{ap} at 0900 hours, particularly for 37 GHz and 94 GHz.

The variation of T_{ap} with m_v is shown in Figure 4 for $\theta = 55°$. Also shown are empirical fits of the form:

$$T_{ap} = A - B\,e^{-cm_v} \qquad\qquad (7)$$

The data used in Figure 4 were acquired during one of the diurnal experiments. The spread in the magnitude of T_{ap} along the vertical axis (for $m_v = 0$) is a measure of the radiometer sensitivity to variations in the thermometric temperature of the snowpack over the diurnal cycle. Of course, the simple empirical expression given by Equation (7) does not account for variations in the thermometric temperature, nor does it account for wetness variations in the snowpack layer beneath the top 5 cm layer.

APPARENT TEMPERATURE RESPONSE TO SNOW WATER EQUIVALENT

During the six-week observation period, the snow depth of the test site varied from 26 cm to 50 cm. Since this is too narrow a range for evaluating the apparent temperature response to snow depth or water equivalent, an experiment was conducted in which dry snow was piled up in 20-30 cm increments up to a depth of 170 cm and T_{ap} was measured after the addition of each new step. This experiment was conducted using the 37 GHz and 94 GHz radiometer at $\theta = 57°$. Similar experiments also were conducted at other angles and frequency combinations.

In conjunction with the T_{ap} measurements, the thermometric temperature T_o of the snowpile was monitored, and after completion of each snowpile experiment, the sky temperature T_{sky} was measured by pointing the radiometer towards the sky. Using Equation (3), the above measurements were used to compute the "measured" emissivity ε as a function of h, the snowpile depth, or W, the snowpile water equivalent (the snow density of each layer also was measured). Figure 5 shows plots of ε as a function of W for 10.69 GHz, 37 GHz and 94 GHz. The lower two frequencies include plots at $\theta = 27°$ and 57° while at 94 GHz, only 57° data were measured.

Returning to Equation (4), for constant snow wetness (in this case zero), T_{sa} and ω are constant at a given frequency, and for constant ground

conditions, T_{gs} is constant. Thus, the only variable in Equation (4) is the loss factor L since it is a function of W. Rewriting Equation (4) in the form:

$$\epsilon = T_{sa}(1 - \omega) + T_{sa}(T_{gs} + \omega - 1) \exp(\kappa_e^\prime W \sec\theta^\prime) \tag{8}$$

and using the abbreviations

$$A_1 = T_{sa}(1 - \omega) \tag{9}$$

$$B_1 = T_{sa}(T_{ga} + \omega - 1) \tag{10}$$

$$C_1 = \kappa_e^\prime \sec\theta^\prime \tag{11}$$

ϵ becomes:

$$\epsilon = A_1 + B_1 \exp(-C_1 W) \tag{12}$$

The form of the above expression appears to provide an excellent fit to the measured data shown in Figure 5.

SEASONAL TEMPORAL VARIATION

Figure 6 is a typical example of the observed temporal variation of T_{ap} for the six-week duration of the 1977 investigation. The top part of Figure 6 shows the snow wetness m_v (of the top 5 cm layer) as a function of time, and the bottom part of Figure 6 shows the gradual increase in snow water equivalent from about 5 cm on 2/11/77 to about 13 cm at the end of the winter season.

T_{ap} is observed to respond consistently to snow wetness variations throughout the observation period shown in Figure 6. In addition to the observed values of T_{ap}, Figure 6 shows the values predicted by Equation (4). Assuming that the mass scattering and extinction coefficients $\kappa_s^\prime$ and $\kappa_e^\prime$ vary linearly with snow wetness m_v, expressions for $\kappa_s^\prime$ and $\kappa_e^\prime$ were determined [9] by fitting the measured data to Equation (4). The penetration depth δ is defined as:

$$\delta = 1/\kappa_e = 1/\kappa_e^\prime \rho \tag{13}$$

where κ_e is the extinction coefficient, $\kappa_e^\prime$ is the mass extinction coefficient and ρ is the snow density. The expression for $\kappa_e^\prime$ determined through fitting the measured data to Equation (4), was used to compute δ as a function of m_v for a snow density of 0.3 g/cm^3. The result is shown in Figure 7 (37 GHz) together with curves for 8.6 GHz and 17 GHz computed in a similar fashion on

the basis of fitting radar backscatter data to the backscattering model given in [1].

CONCLUDING REMARKS

Although additional research is needed to evaluate the sensitivity of the apparent temperature T_{ap} to variations in crystal size, snowpack layering and other snowpack characteristics, the observations made to date indicate that snow wetness and water equivalent are perhaps the two most important factors influencing T_{ap}. Surface roughness has been found to exercise a minor influence on T_{ap} for dry snow, but can cause significant changes in the magnitude of T_{ap} for wet snow. For satellite observations of large cell sizes, however, the effects due to snow surface roughness would be "smoothed" out.

An easy way to separate the influence of snow wetness from that due to water equivalent is by making night and midday observations. The night observations of the cold snowpack would provide water equivalent estimates and the daytime observations would provide information on the snow state (frozen or wet). According to Figures 5 and 7 (and Equation (4)), angles close to nadir are preferable for measuring W than higher angles of incidence since the effective saturation value of W, at which the snowpack starts to appear electromagnetically semi-infinite in extent, decreases with angle. Since the snow extinction coefficient increases with frequency, the measurable range of W decreases with frequency. On the basis of these observations, it may be concluded that from among the three radiometers used in the present investigation, the 10.69 GHz radiometer would provide estimates of W over a greater range of values than the higher-frequency radiometers would, and that the angular range between 0° (nadir) and about 30° is preferred. The choice of this angular range has the additional advantage in that the emissivity is almost angle-independent over this range, and therefore variations in local slope of the snowpack would exercise minor influences on the estimated values of water equivalent.

If, in addition to monitoring snow water equivalent, it is desired to monitor snow wetness variations, a frequency higher than 10.69 GHz would be needed because of the weak sensitivity to m_v at 10.69 GHz (Figure 4). Hence, a feasible configuration for monitoring W and m_v would be either (a) a combination of two radiometers operating at 10.69 GHz and 37 GHz, or (b) a compromise consisting of a single radiometer operating at an appropriate frequency in between, such as 19.35 GHz. A third option is to use a combination of a 10.69 GHz radiometer and a radar with an operating frequency in the 8-18 GHz band. In addition to providing snow wetness information, the radar has a distinct advantage over a 19.35 GHz or 37 GHz radiometer in that it is significantly less sensitive to cloud cover.

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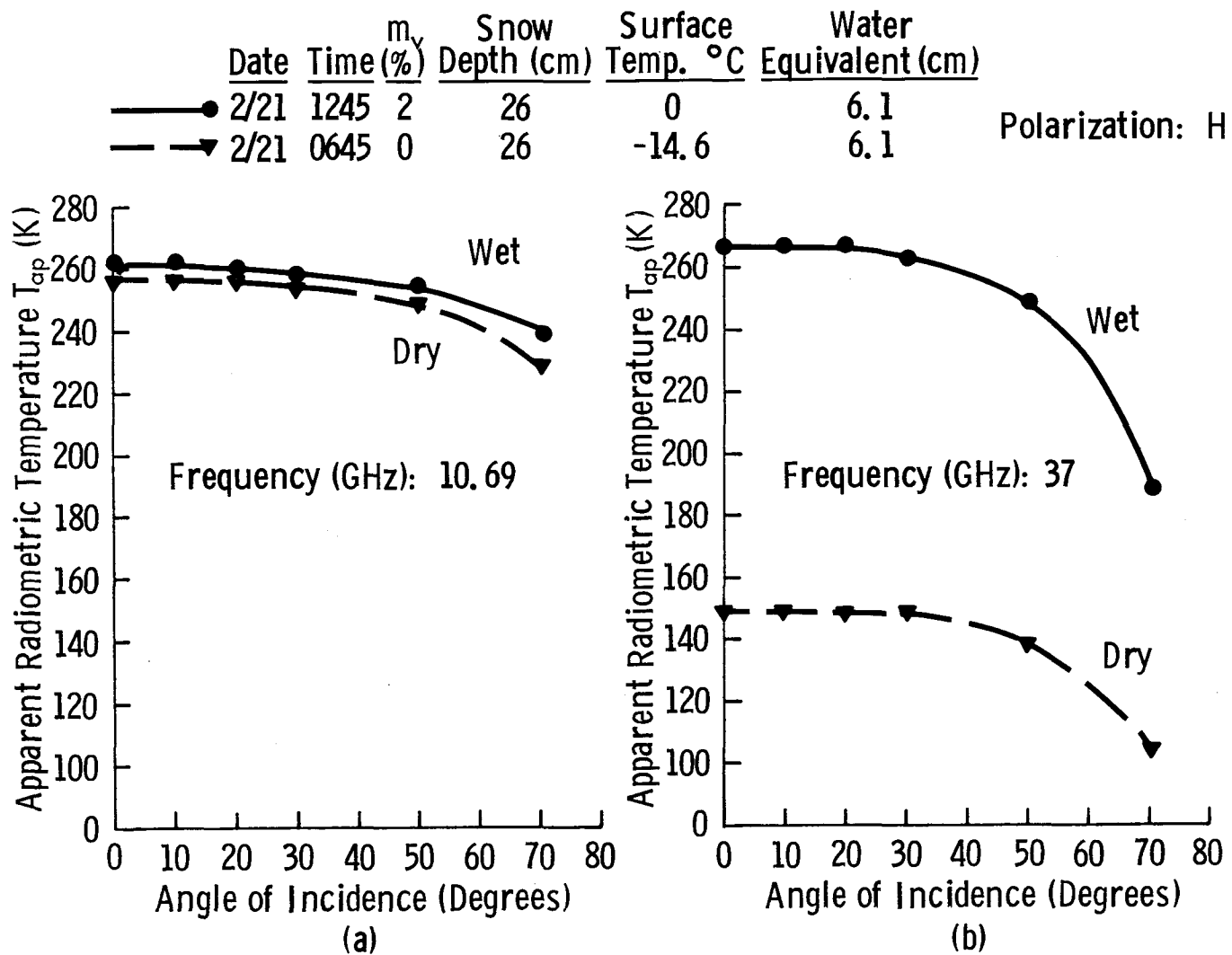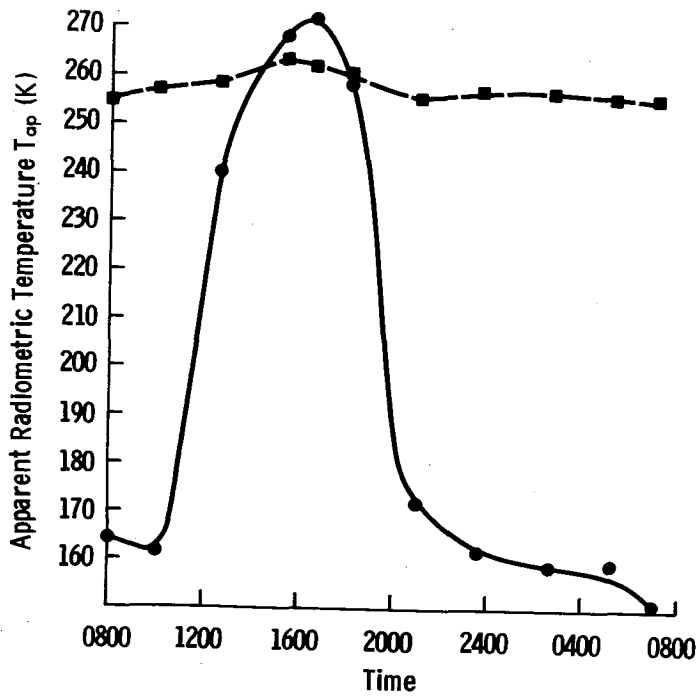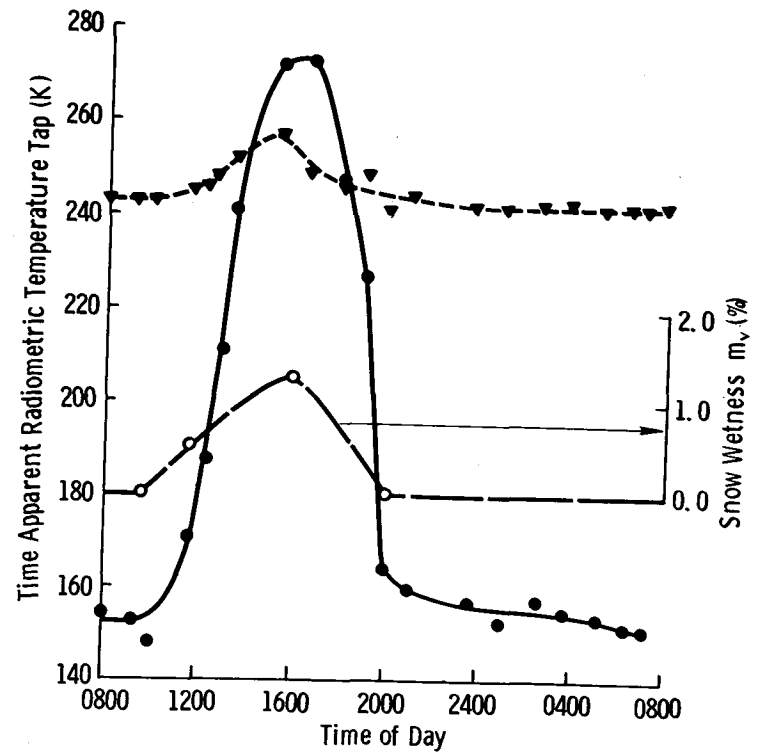Journal of Nordic Hydrology, February, 1979.

Date	Time	m_v (%)	Snow Depth (cm)	Surface Temp. °C	Water Equivalent (cm)
2/21	1245	2	26	0	6.1
2/21	0645	0	26	-14.6	6.1

Polarization: H

Figure 1.   Angular response of T_{ap} at (a) 10.69 GHz and (b) 37 GHz to wet and dry snow on 2/21/77.

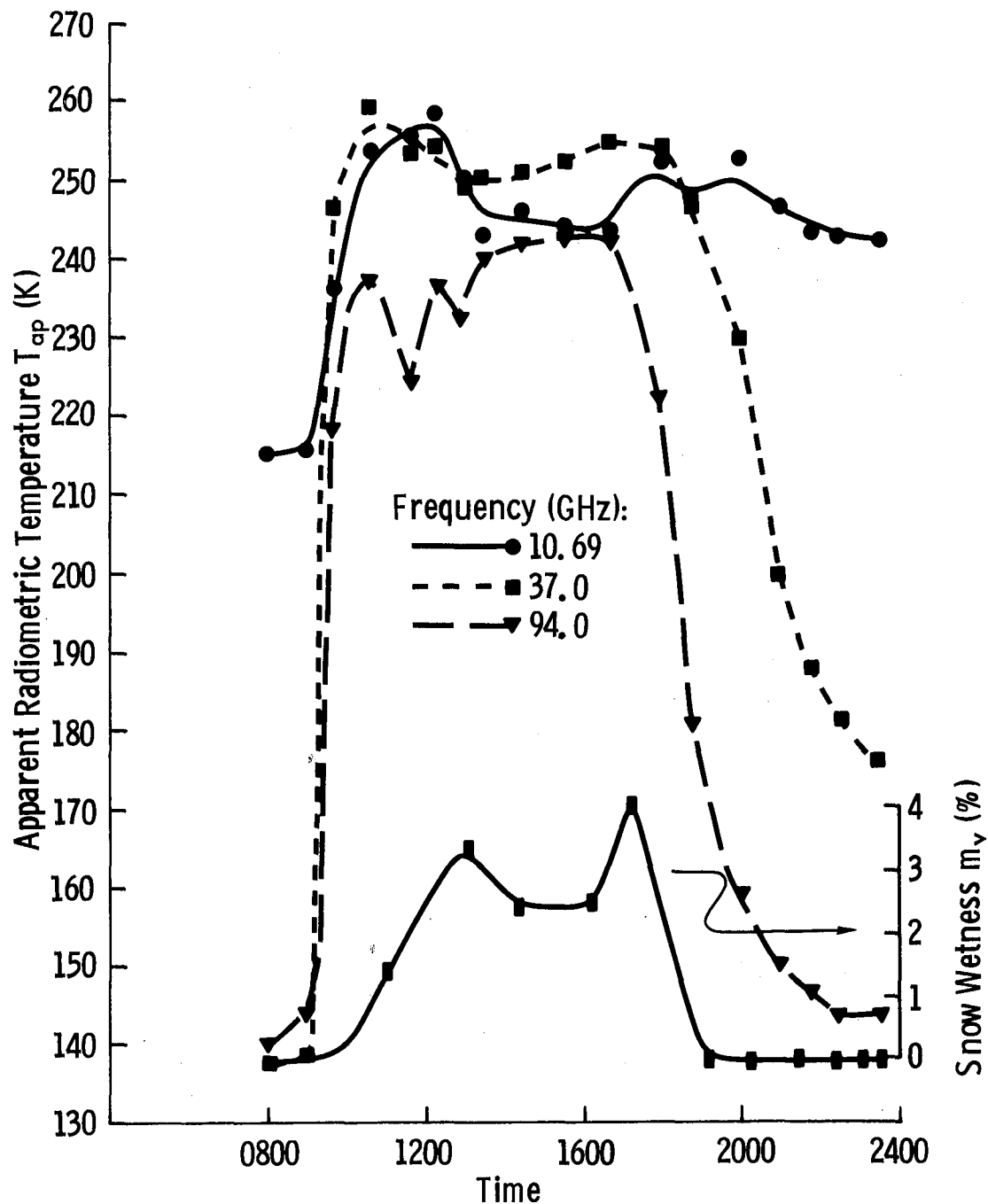Figure 2. Diurnal Variation T_{ap} at 10.7 and 37 GHz at (a) 0°(Nadir) and (b) 50° Angle of Incidence.

Figure 3. Diurnal Variation of T_{ap} at 10.69, 37 and 94 GHz at 50° Angle of Incidence.
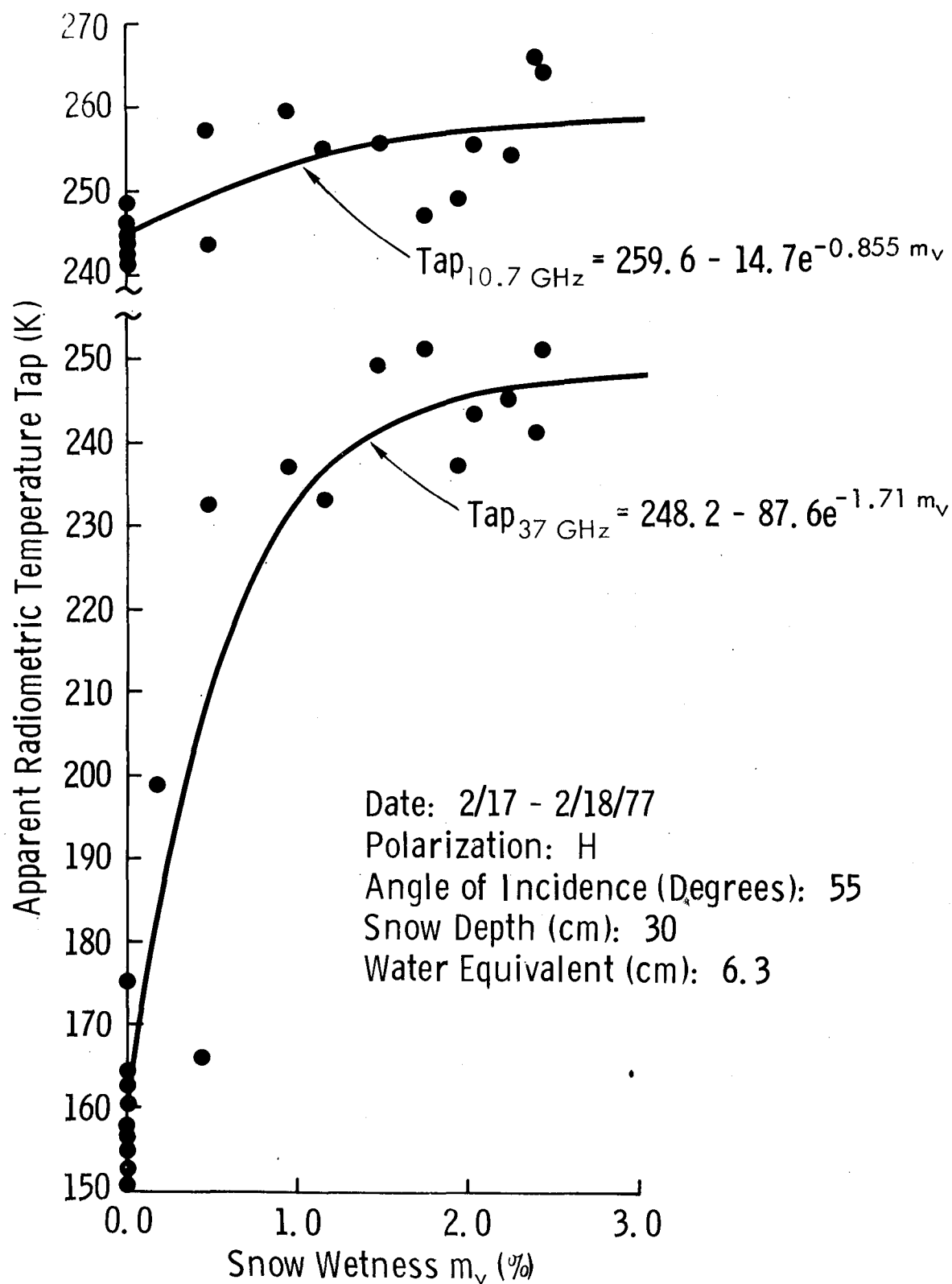
Figure 4. T_{ap} Response to m_v at 50° Angle of Incidence.

The figure shows Apparent Radiometric Temperature T_{ap} (K) versus Snow Wetness m_v (%) with the following fitted curves:

$$Tap_{10.7 \text{ GHz}} = 259.6 - 14.7e^{-0.855\, m_v}$$

$$Tap_{37 \text{ GHz}} = 248.2 - 87.6e^{-1.71\, m_v}$$

Date: 2/17 - 2/18/77
Polarization: H
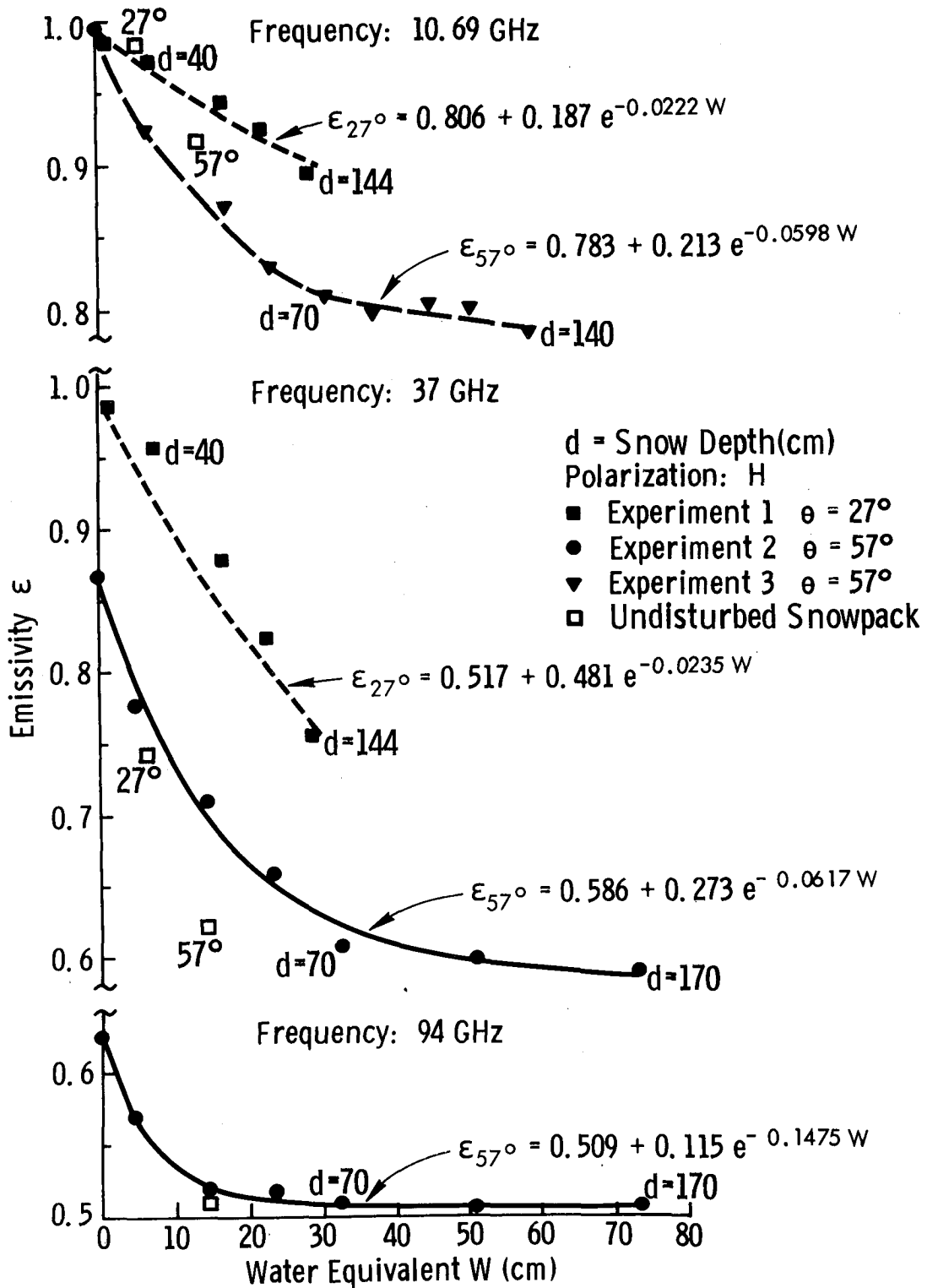Angle of Incidence (Degrees): 55
Snow Depth (cm): 30
Water Equivalent (cm): 6.3

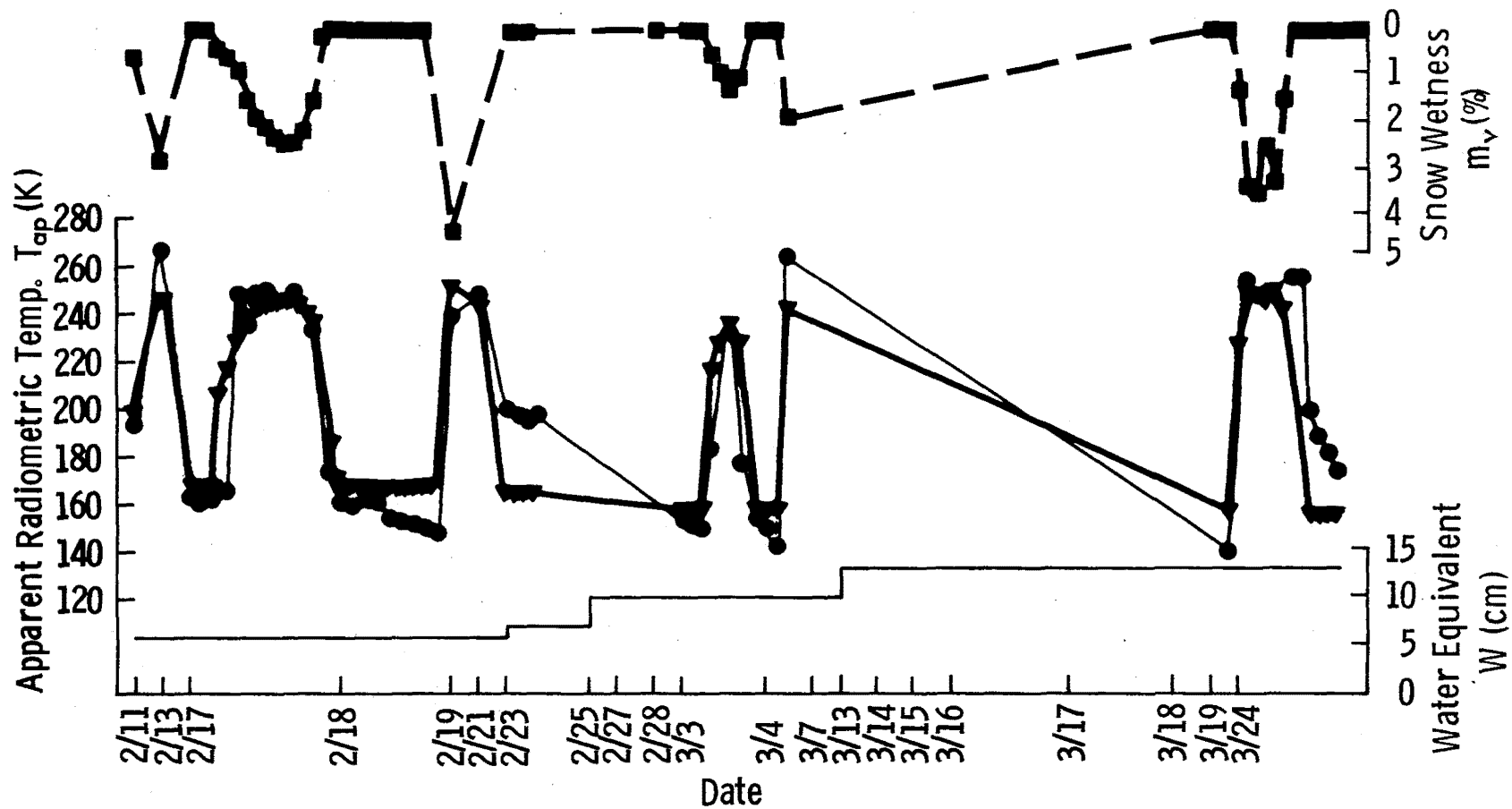Figure 5. Measured Radiometric Emissivity Response to Dry Snow Water Equivalent at 10.7 GHz, 37 GHz and 94 GHz. Average Snow Density was 0.42g/cm³.

Figure 6. Emissivity Model (x273.2) and Observed Apparent Temperature Comparison at 37 GHz.
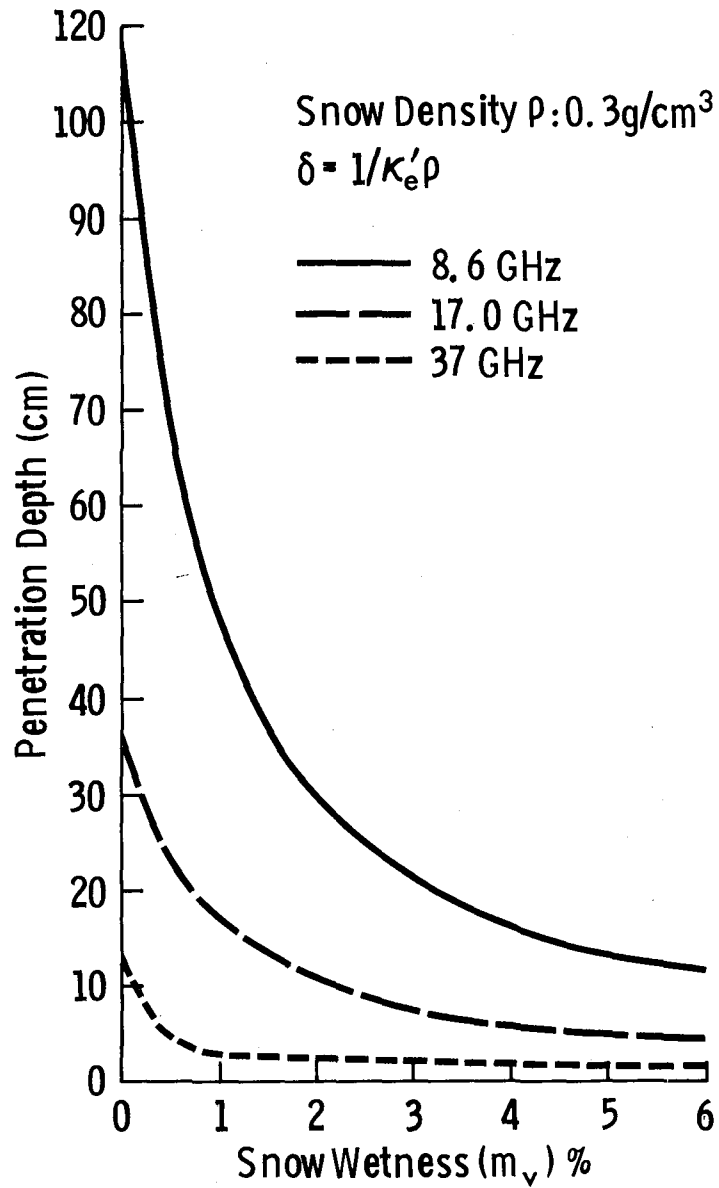
Snow Penetration Depths From Models

Figure 7. Penetration Depth Calculation from the Active and Passive Microwave Models.

MICROWAVE SIGNATURES OF THE NATURAL SNOW COVER

AT WEISSFLUHJOCH

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ABSTRACT

The natural snow cover on a high altitude (2500 m) alpine test site has been monitored with a multi-frequency (1.8, 4.9, 10.5, 21, 36, 94 GHz) radio-meter for more than three years. Some measurements were also made with a 10.5 GHz scatterometer. The microwave observations are supported by a large set of ground truth data. From year to year a wide variation in the development of the snow pack above and below average was observed. Typical microwave data are presented for the different snow conditions in view of the applicability as signatures for remote sensing:

1. During winter condition, characterized by the absence of any melting meta-morphism, the brightness temperatures of the mm wavelength range are related to the water equivalent of the total snow cover, but the data contradict measurements made on artificially packed snow on USA test sites: The natural growth of the snow cover leads first to a rapid decrease of the brightness temperature, and at a water equivalent of approximately 20 cm, a reversal to a linear increase is observed. This increase, first noted by Hofer and Mätzler (1980), can now be confirmed by the data of two additional winters showing the same relationship in a large range of snow depths.

2. In spring when the snow undergoes considerable melting and refreezing the spectra of the brightness temperatures suffer dramatic changes making it easy to distinguish humid and refrozen snow. However it is difficult to find quantitative relations between microwave and ground truth data because of the high variability and heterogeneity of the upper snow layer and of the penetra-tion depth.

3. During the depletion phase when the penetration of microwaves is negligible we expect to find unambiguous signatures of the snow state. This situation occurs in the summer snow when the whole pack is wet. It has been found that the decrease of the brightness temperature due to rain on snow is correlated with the precipitation intensity which also strongly influences the back-scattering coefficient.

INTRODUCTION

A long-term program of monitoring snow parameters by microwave instruments at the standard test site at Weissfluhjoch (Davos) of the Swiss Federal Institute for Snow and Avalanche Research (SLF) started in 1977. The extensive ground-truth measurements (published in annual reports of SLF: "Schnee und Lawinen in den Schweizer Alpen") at this test site and the advanced experience in snow physics are an excellent basis for our work.

The objective of the microwave program is the study of microwave signatures of the ground-truth data for interpretation of remotely sensed data. The equipment (see Figure 8) consists of an elongated box which can be rotated around the longitudinal, horizontal axis. The box contains 6 units with rotatable polarization angle for the containment of radiometers and scatterometers. Together with the first results the radiometers and the measuring procedure have been described by Schanda et al. (1978) and Hofer and Schanda (1978). Other studies dealt with a multivariate data analysis of the first measurements (Hofer and Good, 1979), with the seasonal change of the microwave brightness temperatures and with the penetration of microwaves in snow (Hofer and Mätzler, 1980). In the present work a new instrument, the noise scatterometer, is introduced including first results.

On the other hand the radiometer data have been recalibrated with the aid of the sky temperatures, leading to noticeable correction at 21 and 94 GHz predominantly at low brightness temperatures. Therefore the older data may have changed compared to earlier publications.

In this work the data of the snowcover during winter and summer conditions are investigated primarily, whereas the strongly variable spring condition was investigated earlier (e.g. Schanda and Hofer, 1977).

THE SCATTEROMETER

The simplest way of implementing a scatterometer in a radiometer system is to use a quasi non-coherent transmitter together with the radiometer as receiver of the backscattered signal. If the bandwidth is large enough such a system is very effective because every measurement is the ensemble average of a large number of independent samples. Care has to be paid to a sufficient decoupling of the transmitting and receiving antennas.

Our scatterometer uses a broad-band noise-diode, amplified to an equivalent temperature of $9 \cdot 10^9 K$ in the frequency interval 10 - 11 GHz. This transmitting signal can be attenuated in 7 steps of 10 dB to adapt the power to a convenient receiving signal on the order of 100 K. To cancel out the self emission of the target, the difference between the radiances of the target when the transmitter is switched on and when it is switched off is measured with the phase sensitive detector of the radiometer.

The backscattering coefficient, γ, is defined as the ratio of the radar cross section of the target to its projected area as seen from the position of

the scatterometer. Indices, v and h, denote the polarizations, vertical and horizontal, of the transmitting and receiving antennas, respectively.

At a distance of 20 m the minimum measurable backscattering coefficient is 10^{-4} if we require a minimum temperature difference of 1 K at the receiving antenna, whereas the noise limitation is 0.1 K at 1s integration time. The uncertainty, $\Delta\gamma$, in the measurement of the backscattering coefficient is divided in two parts:

- a systematic uncertainty due to the uncertainty of the antenna gains which we estimate to be about ±10%,
- and a statistical uncertainty which is determined by the number, N, of independent samples

$$\Delta\gamma/\gamma = 1/\sqrt{N} \tag{1}$$

In a stationary configuration of a noise scatterometer the reflections from two different points can be regarded as independent, i.e. their superposition is incoherent, if the dispersion $\Delta\varphi$ of the phase angles over the bandwidth Δf, is at least unity. If the reflections from all points are distributed continuously over a range, Δr, of distances between target and scatterometer, N is given by

$$N = \Delta\varphi\,(\Delta r) = 2\;\Delta r\;\cdot\frac{2\pi\;\Delta f}{c} \tag{2}$$

where c is the speed of light. In case of single reflections on a horizontal surface measured from a height, h, at a nadir angle, θ, the variation Δr is given by

$$\Delta r = h\;\Delta\theta\;\frac{\sin\,\theta}{\cos^2\,\theta} \tag{3}$$

where $\Delta\theta$ is the beamwidth of the scatterometer. With the following scatterometer specifications:

$\Delta f = 0.9$ GHz, $\Delta\theta = 7°$, h = 14.5 m we find

$$N = 67\;\frac{\sin\,\theta}{\cos^2\theta} \tag{4}$$

at $\theta = 45°$ the statistical, relative uncertainty $\Delta\gamma/\gamma = 10\%$.

The effect of volume scattering is to increase Δr by the penetration depth, and multiple reflections increase Δr further, so that (4) can be regarded as a lower limit of N.

SOME CHARACTERISTICS OF AN INHOMOGENEOUS SNOWCOVER

The ground truth on the alpine test site of Weissfluhjoch generally corresponds to a highly inhomogeneous medium. In the worst case the snowcover is built up of water in its gaseous, liquid and solid phase together with air and various impurities.

The inhomogeneities are of different levels:
Stratification due to a succession of snowfalls as reflected in an example of stratigraphic- and ramhardness- profiles of Figure 1. As long as the temperature is below the freezing level (profiles from December 1977 til April 1978), no free water is present in the snow cover, however.
This may not be true for the surface layer (s), exposed to radiation, to an increase in air temperature (see days with an air temperature above freezing point in Figure 1), or built up by precipitation of wet snow. Cooled down, these icy layers of a higher density than the ambient snow become narrow lamella between rather homogeneous strata.
Weak surface layers (e.g. surface hoar) on the other hand are another source of discontinuities.
Both formations are not to be detected by the ramsonde and will appear only in a very carefully made stratigraphic profile.

Besides the inhomogeneities, size and shape of grains widely vary for different snow layers. This is illustrated in photos of Figures 2 and 3, and in Table 1, where means and standard deviations of a few selected parameters from thin section analyses are given. The samples were taken out of a pit from 13-Feb-1979 at the radiometric test site. Vertical thin sections (20 microns thick) represent the state of the snow cover at 6, 25, 35, 55.5 and 66 cm above a metallic foil covering the bare ground. The density is found to vary by a factor of 2.5, whereas the mean intercept length L ('mean grain diameter') changes by a factor of 4.

When free water is present in the top layer (or in the whole snowpack), discontinuities as described above will disappear or their effect on the brightness temperature will strongly be attenuated.

In spring condition, the snowcover approaching 0 C, melt-metamorphism, a succession of melting and freezing, will compel the small grains to disappear and the larger ones to grow yielding spherical particles (firn) (see profiles of May and June 1978 of Figure 1). Considerable runoff begins with summer condition when the whole snowcover remains wet. The typical depletion of the snowcover is 5 cm snow depth per day.

ON THE WINTER CONDITION

We call a snowpack to be in winter condition when melting metamorphism is absent. Measurements under this condition were made in Terms 2, 3 and 4 of the program and first results of Term 2 were reported by Hofer and Mätzler. The development of the snowcover in Term 2 (1977/78) is shown in Figure 1. Although the snow depth was below average in early winter it reached average values in March. The snowcover of Term 3 was even less, and in Term 4 the snow depth was clearly above average, so that data from very different situations can be analyzed.

Although the distance between the standard test site and the microwave test field is only 70 m, the snow depth is always lower at the microwave test field. On the average the water equivalent of the microwave test field is 0.7 times the water equivalent of the less wind exposed standard test field.

Figure 4 shows the brightness temperature of the snowcover of January 30–31, 1980 at 10.4 and 36 GHz as a function of nadir angle and sky temperature at two zenith angles. Figure 5 shows the backscattering coefficients at 10.4 GHz of the same period. The snowcover was already larger, but still comparable to the situation in March, 1978 shown in Figure 1. In Figure 4 there is a clear difference between vertical and horizontal polarization. However, the Brewster maximum (vertical polarization) is not well developed, the curves are flat. This behavior is indicative of Lambert scattering which is here superimposed on a Fresnel-like reflection mechanism. The former scattering is diffuse and can indeed be identified in Figure 5 as a diffuse component. The backscattering coefficient of this component fits the Lambert $\cos\theta$ law for γ.

$$\gamma_{vv} \simeq \gamma_{hh} = \gamma_o \cos\theta$$

where $\gamma_o = 0.1$. The albedo, A_d, of this diffuse component is independent on θ

$$A_d = \gamma_o/4$$

With the average snow temperature, $T_{snow} = 268$ K, the sky temperature, $T_{sky} = 5$ K, the depression of the brightness temperature due to this diffuse component becomes $(T_{snow} - T_{sky}) \cdot A_d = 6.5$ K which is half of the depression of the 10.4 GHz brightness temperature from the black-body situation at the Brewster maximum.

During winter condition there are no rapid changes of the microwave response. Especially there are hardly any daily variations and also additional fresh snow hardly shows an immediate effect. A slow variation, correlated with the growth of the snowcover, was reported by Hofer and Mätzler as shown by their Figure 2 (December – March). This behavior was reexamined with the data of the later terms, and the results for vertically polarized emission is shown as a scatter plot versus water equivalent in Figure 6. The data are averaged over nadir angles from 25° to 55°, and since the difference between the 4.9 and 10.4 GHz data is insignificant, their average values are shown. The figure contains the data of all the measuring periods at winter condition. A pronounced change of T with water equivalent is found at 36 GHz with strongly decreasing values of T at low snow depths and with a linear increase between water equivalents of 20 and 50 cm. From very few measurements made at low atmospheric absorption at 94 GHz there is an indication that the described behavior is more pronounced at higher frequencies. At horizontal polarization the same behavior is found with all brightness temperatures being about 20° lower (45° nadir angle) than at vertical polarization.

A qualitative explanation of the phenomenon is possible with the ground truth data of Figures 1, 2 and 3, showing a layer of large crystals (depth-hoar) which was built up at a water equivalent of 15 cm at the bottom of the snowcover. This depth-hoar layer remains constant throughout the winter, and due to the accumulation of the fine-grain snow the depth-hoar is covered more and more. Because effective scattering of microwaves is limited to the

large crystals at the bottom the lowest brightness temperatures are to be expected when the depth-hoar just has been created. The fine-grain snow primarily acts as an absorbing layer thus increasing the brightness temperature with increasing thickness.

Figure 6 does not agree with measurements made on dry snow by Meier and Edgerton (1971) and Ulaby and Stiles (1980). Both experiments were made on artificially packed snow. Furthermore the measurements of Meier and Edgerton and also Experiments 2 and 3 of Ulaby and Stiles were made with refrozen snow which has stronger backscattering than winter snow (Hofer and Mätzler, 1980). From all the measurements made at winter condition two composite spectra (1.8 to 94 GHz) can be constructed for water equivalents of 20 and 50 cm. The result for vertical polarization is shown in Figure 7. Because the penetration depth is much larger than the snow depth at the low frequency end, the change of the spectrum at 1.8 GHz is influenced by the underground. The actual effect of the snowcover on the brightness temperature is reflected at frequencies above 10 GHz.

MEASUREMENTS ON SUMMER SNOW (FIRN)

When the ripe snowcover is melting the spherical snow grains have diameters of 1 to 2 mm, and the snow density is about 0.45 g/cm^3 at the surface. The wetness, F, of the snow hardly exceeds 10%, because the water starts to percolate through the snow at a wetness exceeding a certain limit, F_i, the irreducible wetness (Colbeck, 1971). Therefore the microwave signatures do not change strongly from year to year, and they do not depend on the snow depth because the penetration depth of microwaves is very small.

In 1979, a typical firn situation was monitored continuously from May 31 to June 12. During this period the air temperature was always above 0 C, so that the snowcover did not freeze, except for a thin surface layer during a few clear nights. Frequent rainfalls changed the almost smooth surface on June 1 (see Figure 8) to a strongly eroded surface on June 12 (Figure 9). Figure 10 shows the spectra of the brightness temperatures of the two different snow conditions. The surface roughness primarily decreases the difference between horizontally and vertically polarized emission. Figure 11 shows the backscattering coefficients of the two snow surfaces. As it is to be expected the rough surface has higher returns than the smooth surface but still lower ones than the dry snowcover. The error bar at 45° nadir angle shows the variability of γ_{vv} within one day, excluding precipitation and freezing of the surface. An interesting effect is observed when rain falls on snow. Measurements made at 45° nadir angle, vertical polarization of the brightness temperatures at 1.8 GHz and more at 4.9 GHz show a decrease of T with increasing rain intensity. At the same time the backscattering coefficient, γ_{vv}, at 10.5 GHz increases drastically, whereas at 36 GHz, the quasi-blackbody situation is hardly changed. Figure 12 shows time profiles of microwave signals and of the rain intensity of June 7, and Figure 13 is a scatter plot of the 4.9 GHz brightness temperature versus rain intensity. A similar scatter plot of γ_{vv} can not be made with the present data because of the limited dynamic range during the automatic registration.

The infiltration of rain water into the snowcover can be studied by comparing the data of Figure 13 with Figure 14 which shows the dependence of the brightness temperature at 4.9 GHz on the wetness F in percent by volume at a similar snow surface. The sensitivity dT/dF is 1K/% at vertical polarization (45°), 3K/% at nadir and 7K/% at horizontal polarization. Although the sensitivity is much larger at horizontal than at vertical polarization, the statistical variations are smaller at vertical polarization enabling the retrieval of the wetness from the vertical polarization as well as from horizontal polarization. Further the extrapolation of the curves in Figure 14 to F = 1 leads most closely to the corresponding brightness temperature of water in case of vertical polarization, so that linearity might be preserved up to high values of F. That the wetness of the snow surface must be very high during heavy rain is demonstrated by the significant depression of T in Figure 13. With the assumption of linearity between F and the vertically polarized T we find a quadratic relationship between wetness and rain intensity. According to the equation of continuity the rain intensity must be equal to the volume flux, u_w, of water at the snow surface, and the result we find from Figure 13 and 14 is

$$u_w = 60 \ (F-F_i)^2 \qquad (mm/h),$$

where F_i 0.04. This equation is a consequence of Darcy's law for gravitational flow. However the exponent should be 3 for laminar flow on free surfaces. The value of 2 which can be explained by saturated flow due to the presence of layers with thin pores was already assumed by Colbeck (1971) in his theory of one-dimensional water flow through snow.

CONCLUSIONS

Measurements obtained from a long term program of monitoring an Alpine snowcover with microwaves under well defined conditions have been presented. The advantage of these data as microwave signatures is the reduction of variables by using the natural snowcover at the same test site. Thanks to this simplification the program starts to provide certain signatures of the snowpack, namely the water equivalent of the snow at winter condition, the wetness of the melting snow and the rain intensity. Furthermore it has been shown that the microwave data, together with ground-truth data can be used to study snow hydrological processes such as the percolation of water. In order to test the applicability of the results to other snow regions the measurements have to be compared with data obtained under comparable situations from other test fields. Furthermore in case of dry snow the ground-based measurements must be extended to airborne mapping of the whole test area in order to assess the influence of the underground on the microwave data.

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Table 1   Properties of the snow cover of 1979 February 13, and parameters of the vertical thin-section analysis.

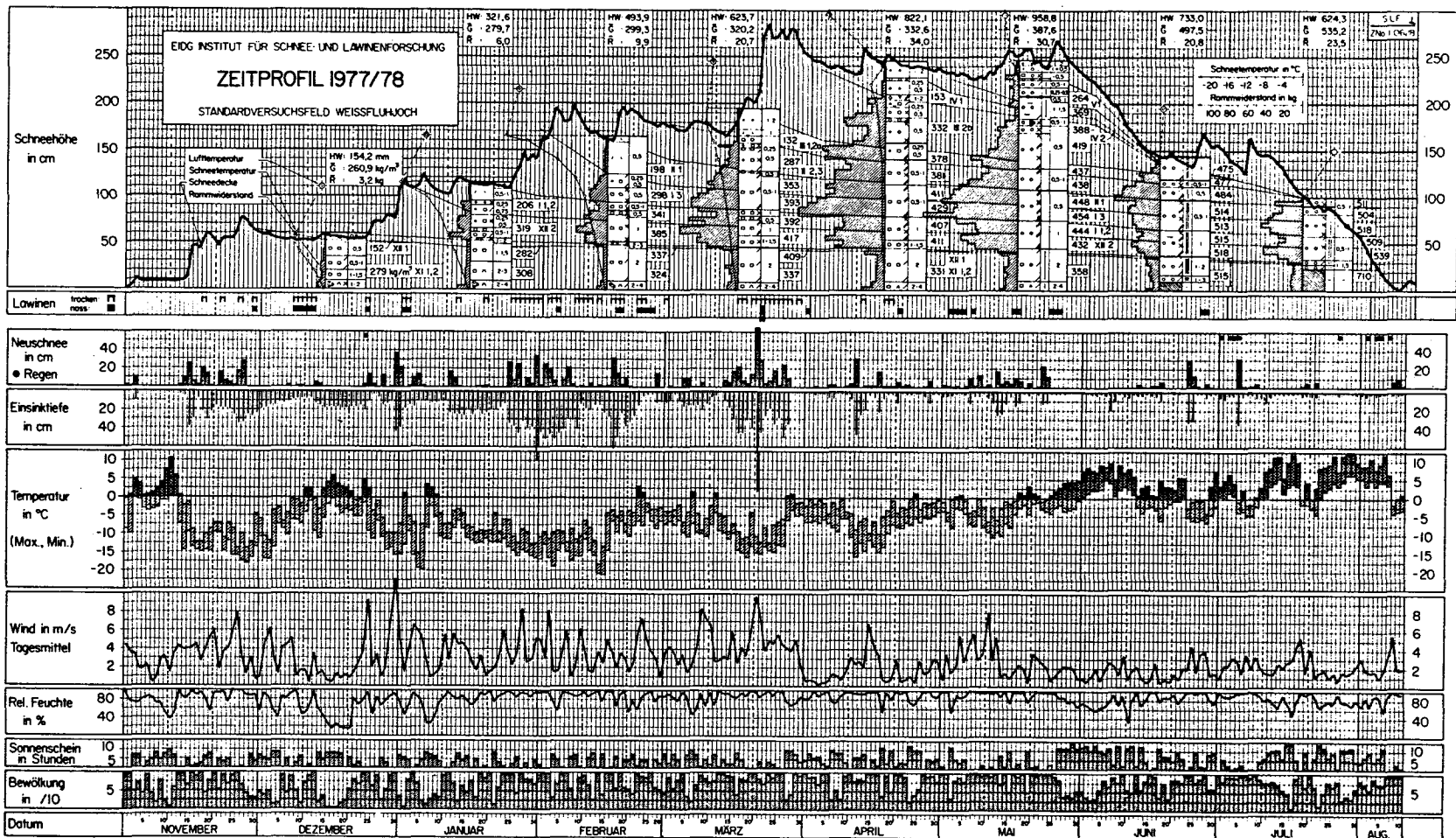| Upper level of snow sample (m)	Snow density (g/cm³)	Snow temperature (°C)	Point density	Intercept length (mm)
0.66	0.12	−9.5	0.13 ± 0.04	0.22 ± 0.13
0.555*	0.21	−5.1	0.23 ± 0.04	0.26 ± 0.10
0.35	0.25	−3.7	0.28 ± 0.05	0.32 ± 0.10
0.25	0.30	−3.5	0.32 ± 0.12	0.90 ± 1.0
0.06*	0.26	−3.3	0.29 ± 0.07	0.74 ± 0.45

*see Figures 2 and 3

Figure 1 Characteristic time profile of the snowdepth and meteorological parameters at the standard test site at Weissfluhjoch with monthly profiles of the temperature, the ram hardness and crystal structure. Given are also the water equivalent, HW, the average density, $\bar{G}$ and the average ram hardness $\bar{R}$ (from "Schnee und Lawinen in den Schweizer Alpen, Winter 1977/78, Eidgenössisches Institut für Schnee- und Lawinenforschung, Weissfluhjoch/Davos, No. 42, 1979).
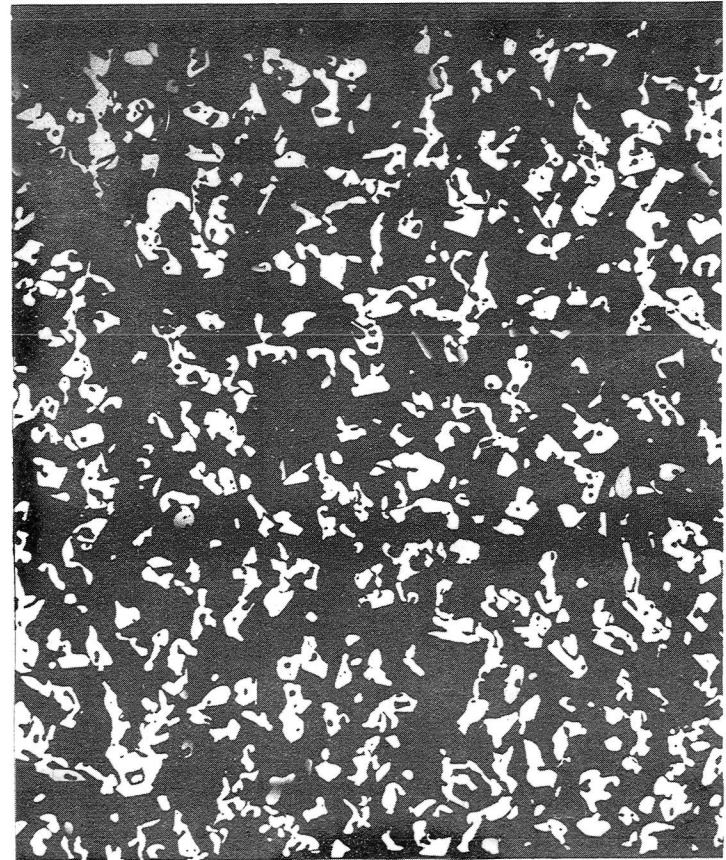
10 mm
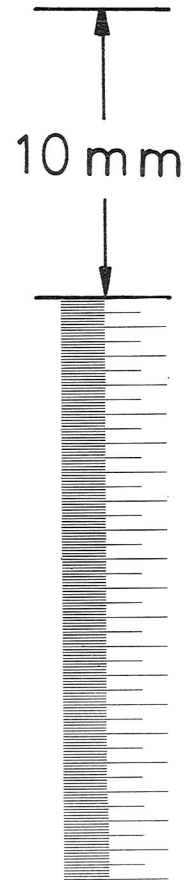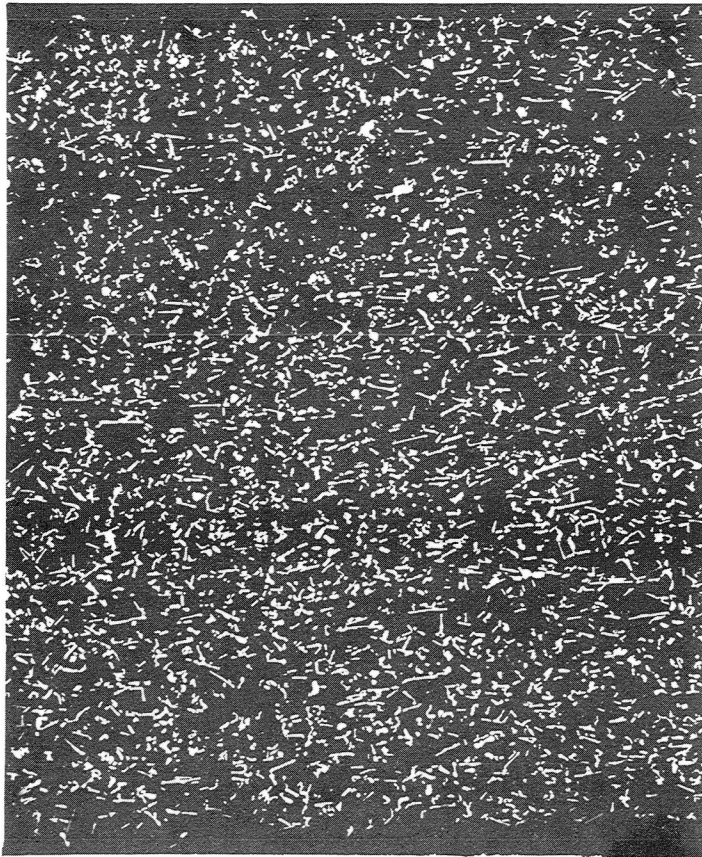
Figure 2   Vertical thin section of
         snow sample of 1979 Feb.
13, top of photo is 55.5 cm above ground.

Figure 3   Vertical thin section of snow
         sample, top of photo is 6 cm
above ground.
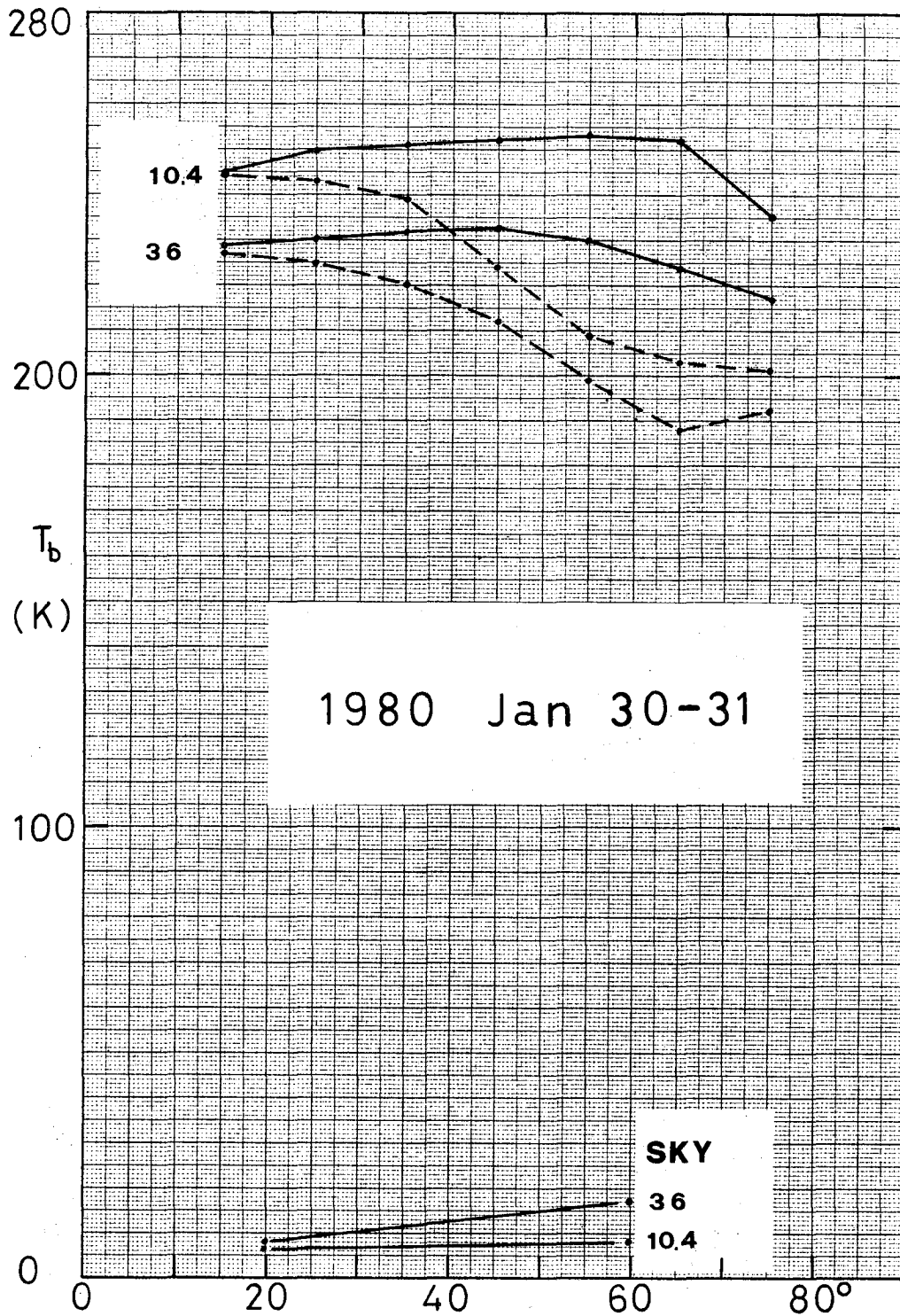
Figure 4   Brightness temperature of a snowcover and of the
            sky at winter condition versus nadir angle and
zenith angle, respect. at 10.5 and 36 GHz. Solid lines: ver-
tical polarization, dashed lines: horizontal polarization.
Snow cover: water equivalent = 47 cm, snow depth = 150 cm,
average temperature = -5°C.
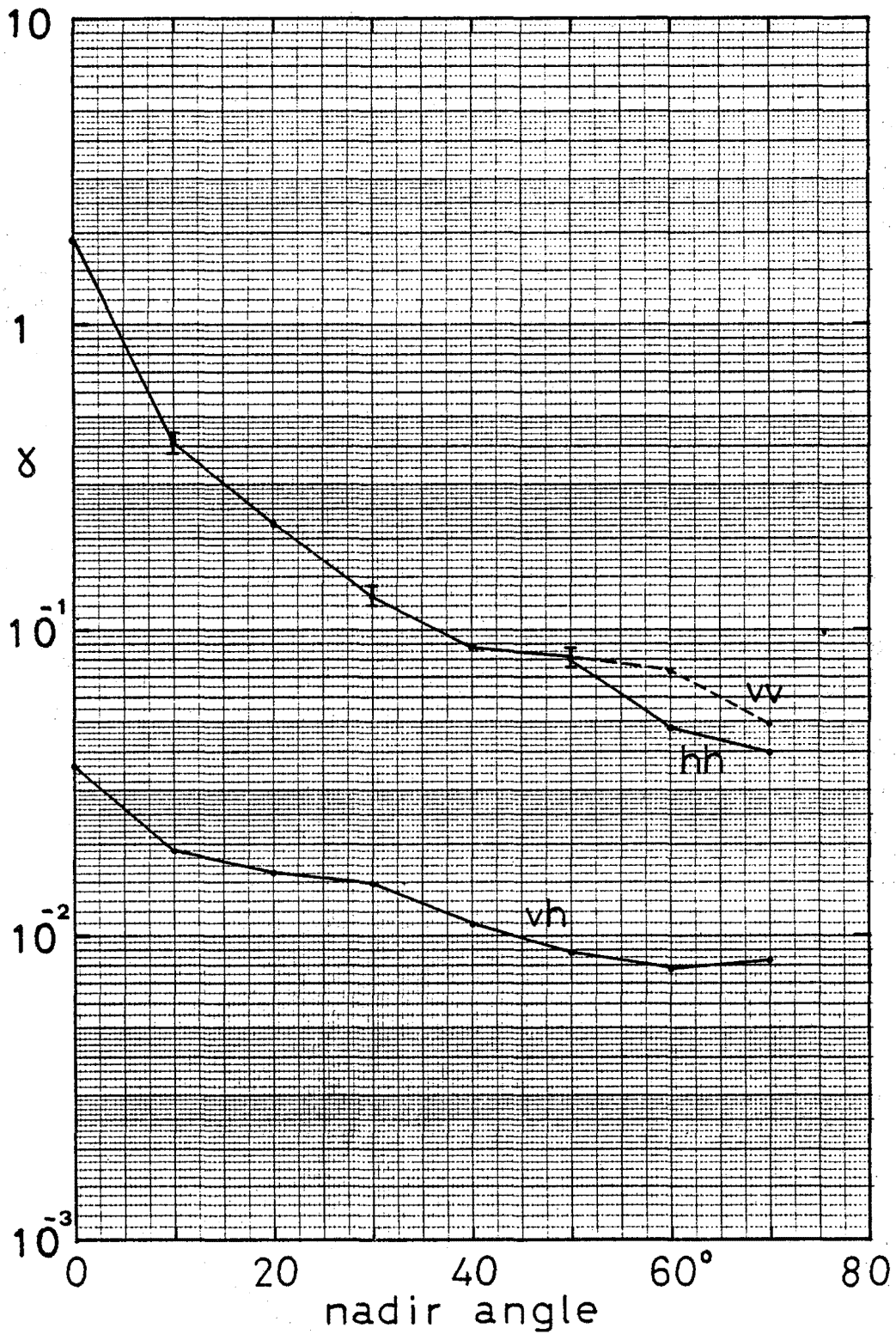
Figure 5   Backscattering coefficients at 10.4 GHz versus
            nadir angle of the snowcover of Figure 4.
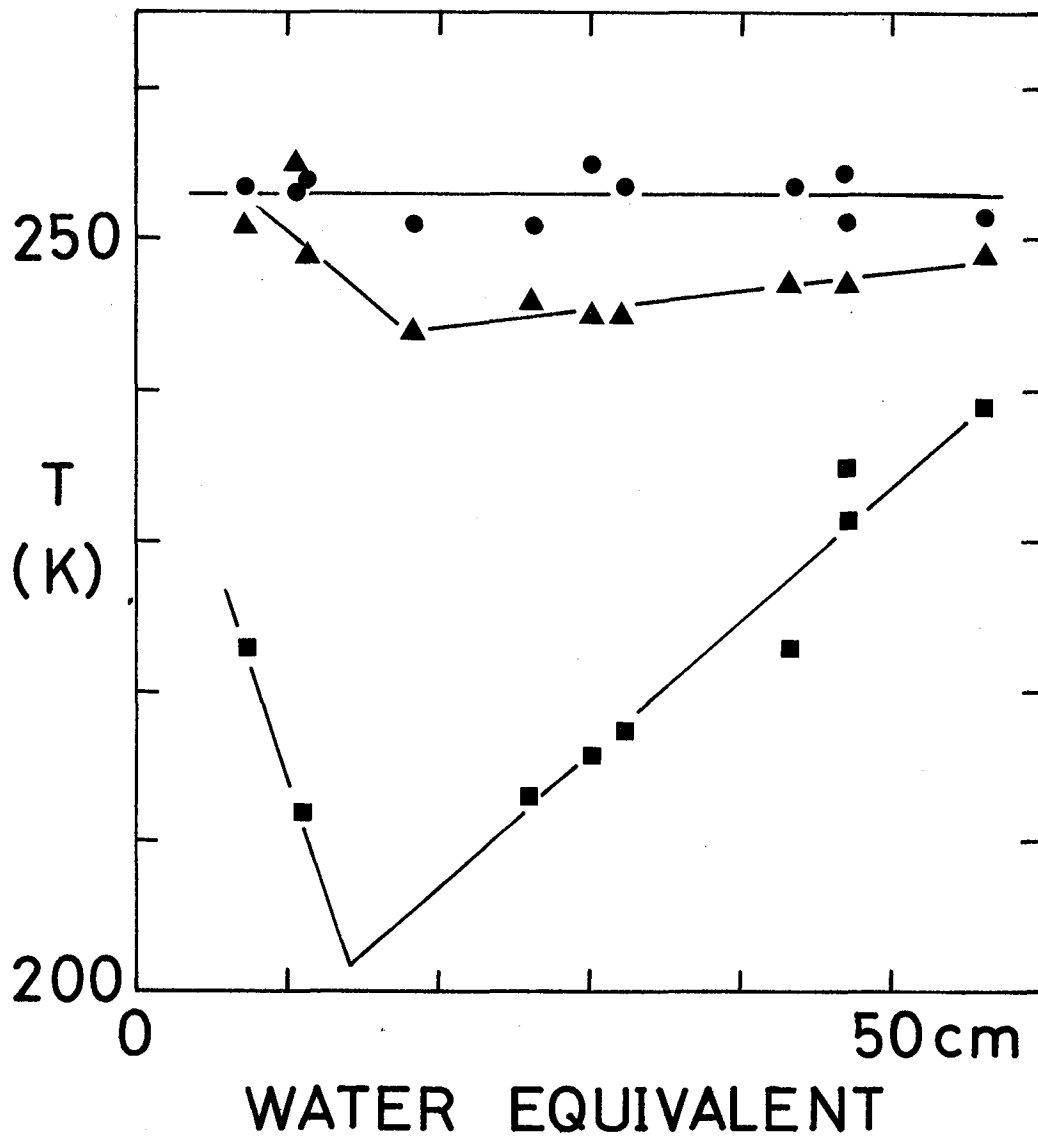
Figure 6   Scatter plot of brightness temperatures at ver-
            tical polarization, averaged over nadir angles
from 25° to 55° versus water equivalent of the snowcover at
winter condition. Circles are mean values of 4.9 and 10.4
GHz, triangles 21 GHz and squares 36 GHz.
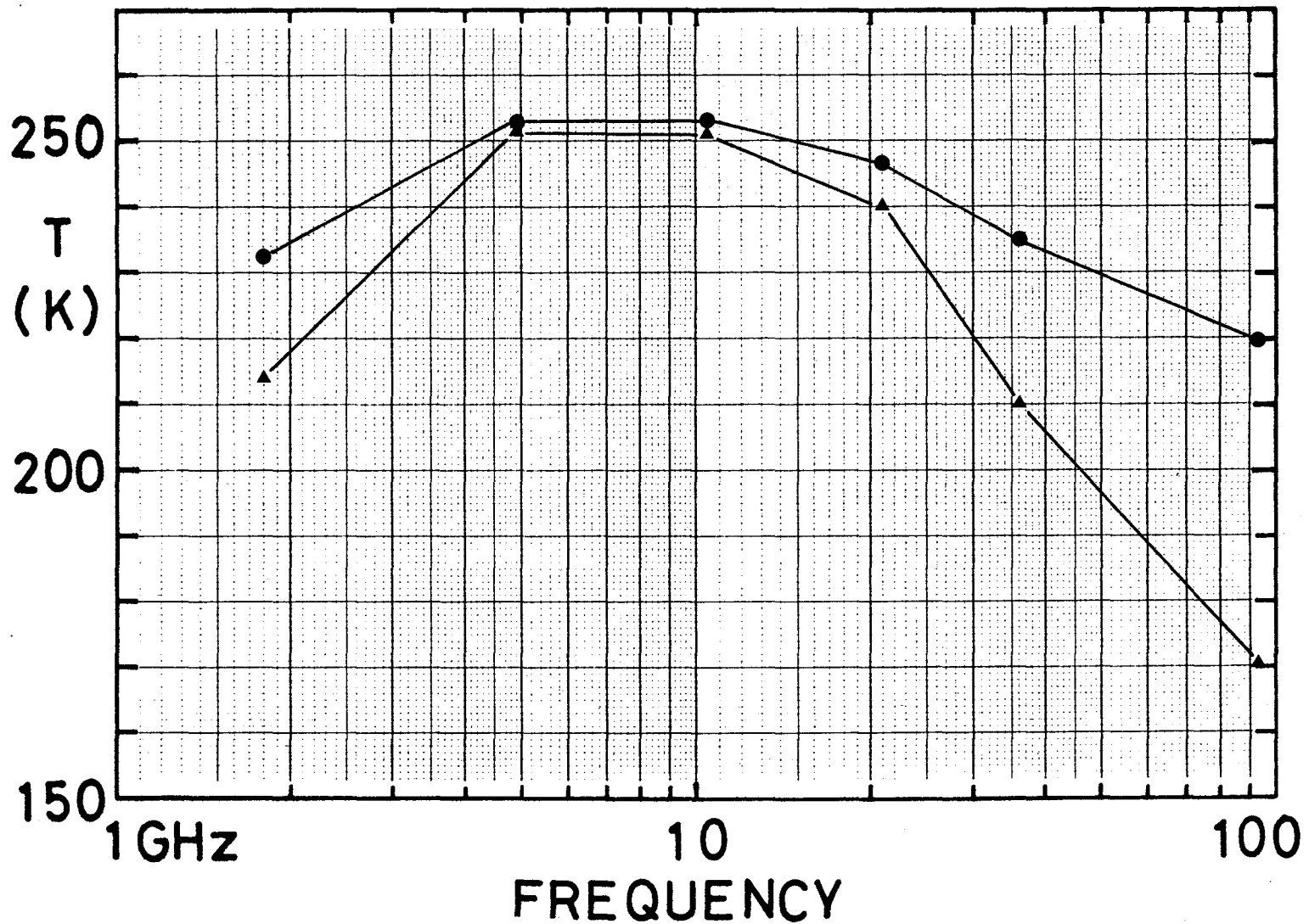
<u>Figure 7</u>   Composite spectra, vertical polarization of an
             Alpine snowcover at winter condition for water
equivalents of 20 cm (triangles) and 50 cm (circles).

Figure 8   Photo of the test site with radiometer box on the tower, the Ambach snow wetness instrument on the refrozen snow surface in early morning of June 1, 1979.
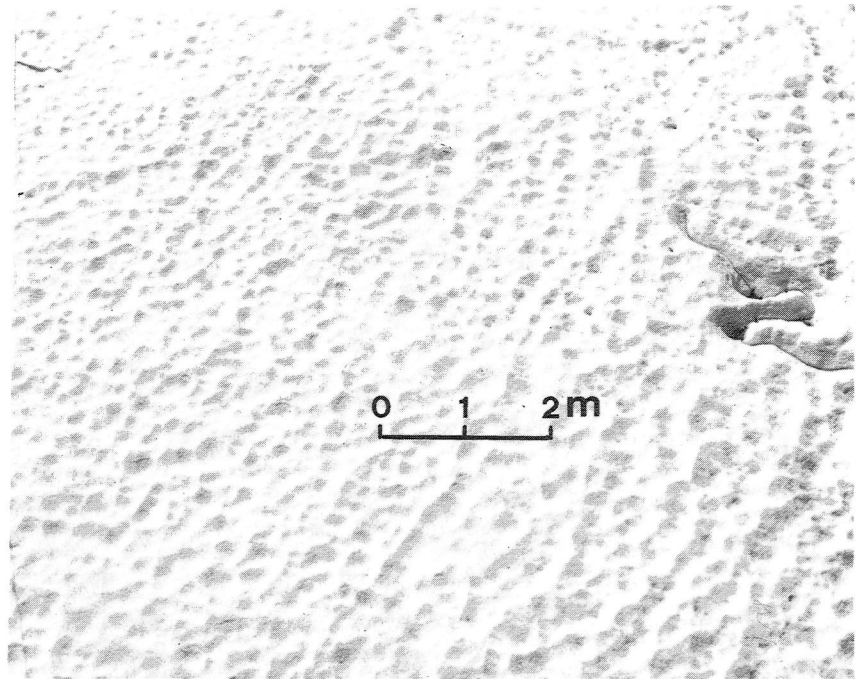
Figure 9   The strongly eroded snow surface on June 12, 1979.
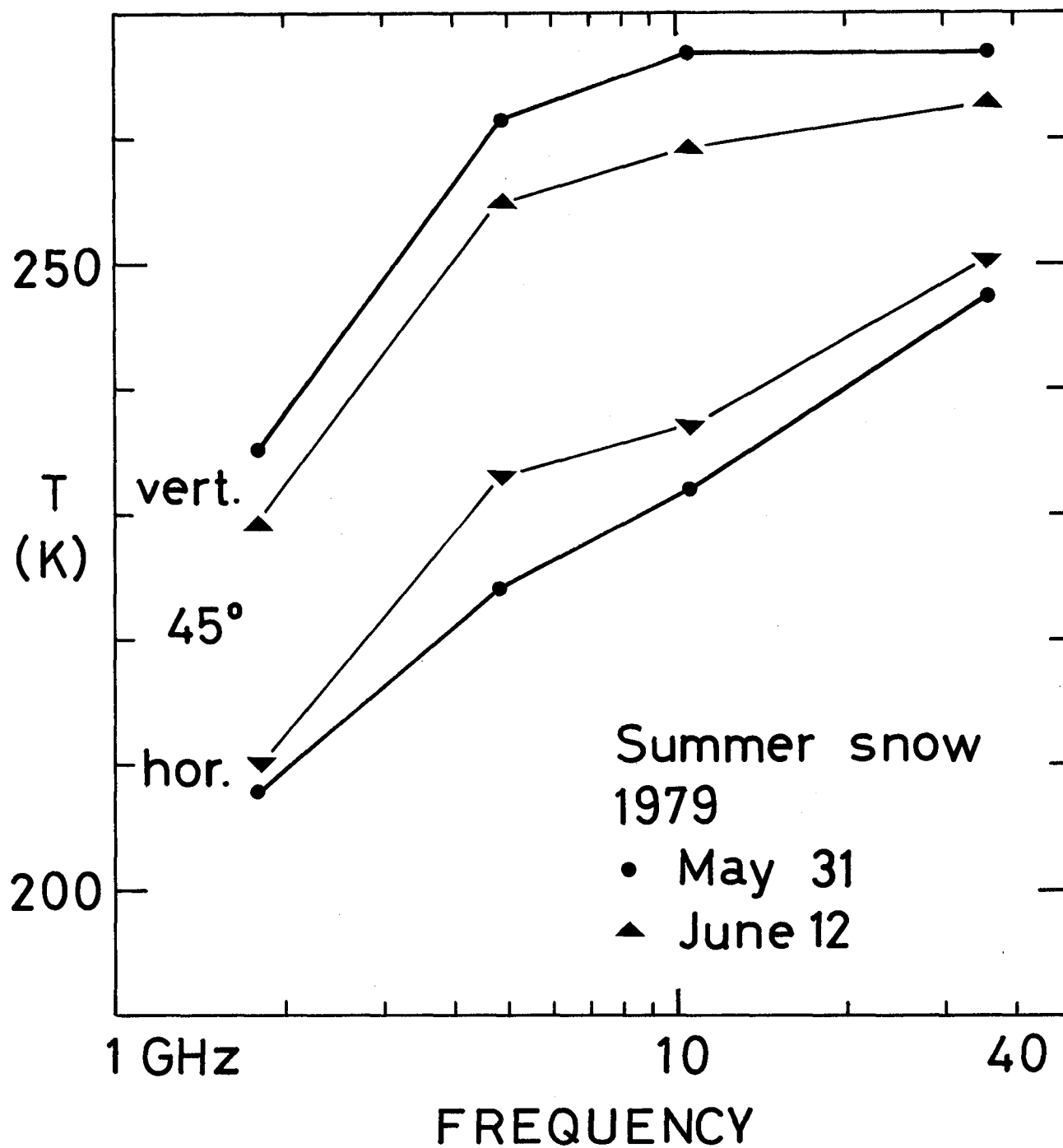
Figure 10   Spectra of brightness tem-
            peratures of a snow cover
at summer condition before and after a
period of rain, horizontal and vertical
polarization, 45° nadir angle.
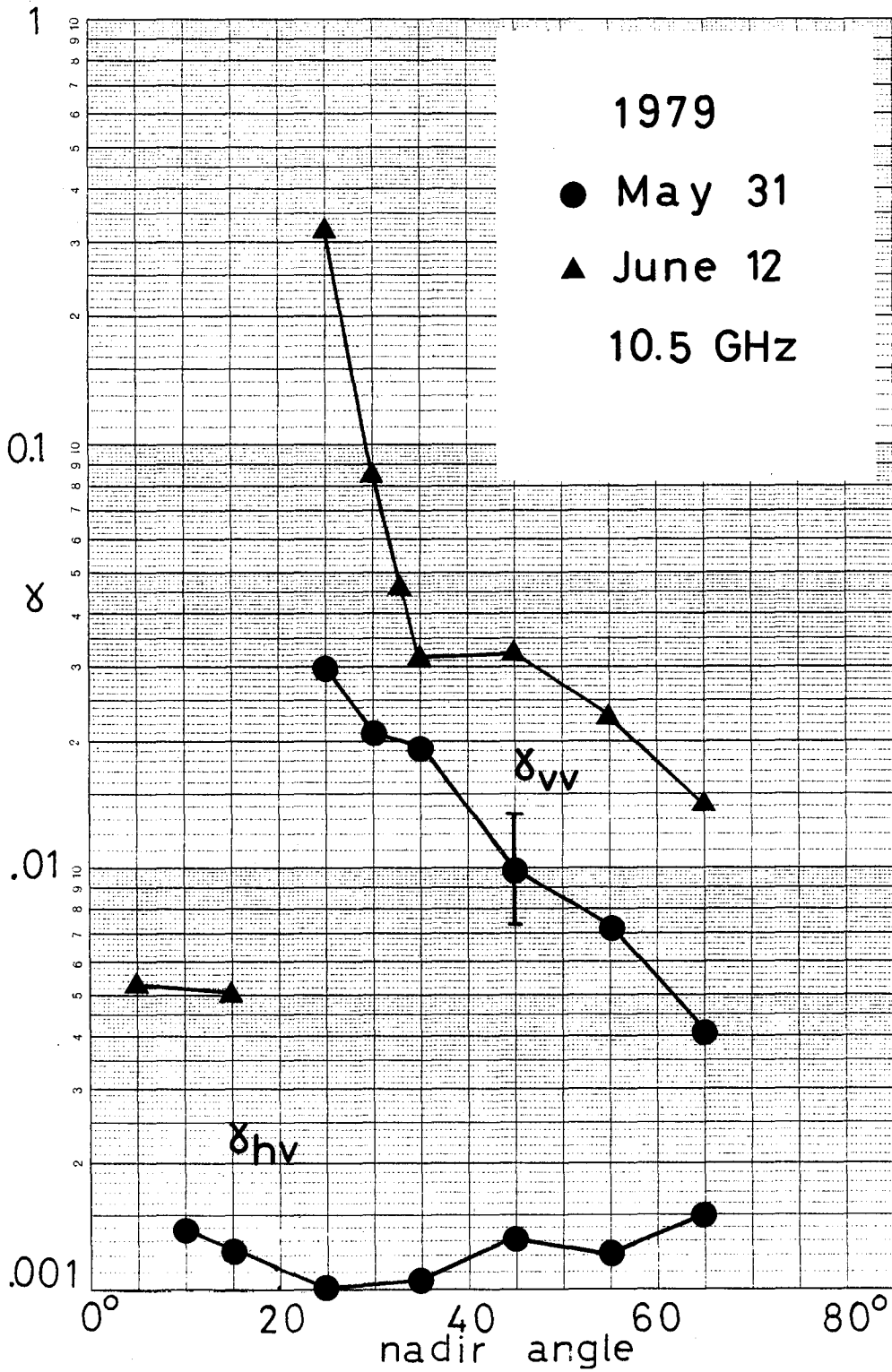
Figure 11   Backscattering coefficients
γ_{vv} and γ_{hv} of a snow cover
at summer condition before and after a
period of rain. The variability of
on May 31 and June 1 is shown by the er-
ror bar at 45° nadir angle.
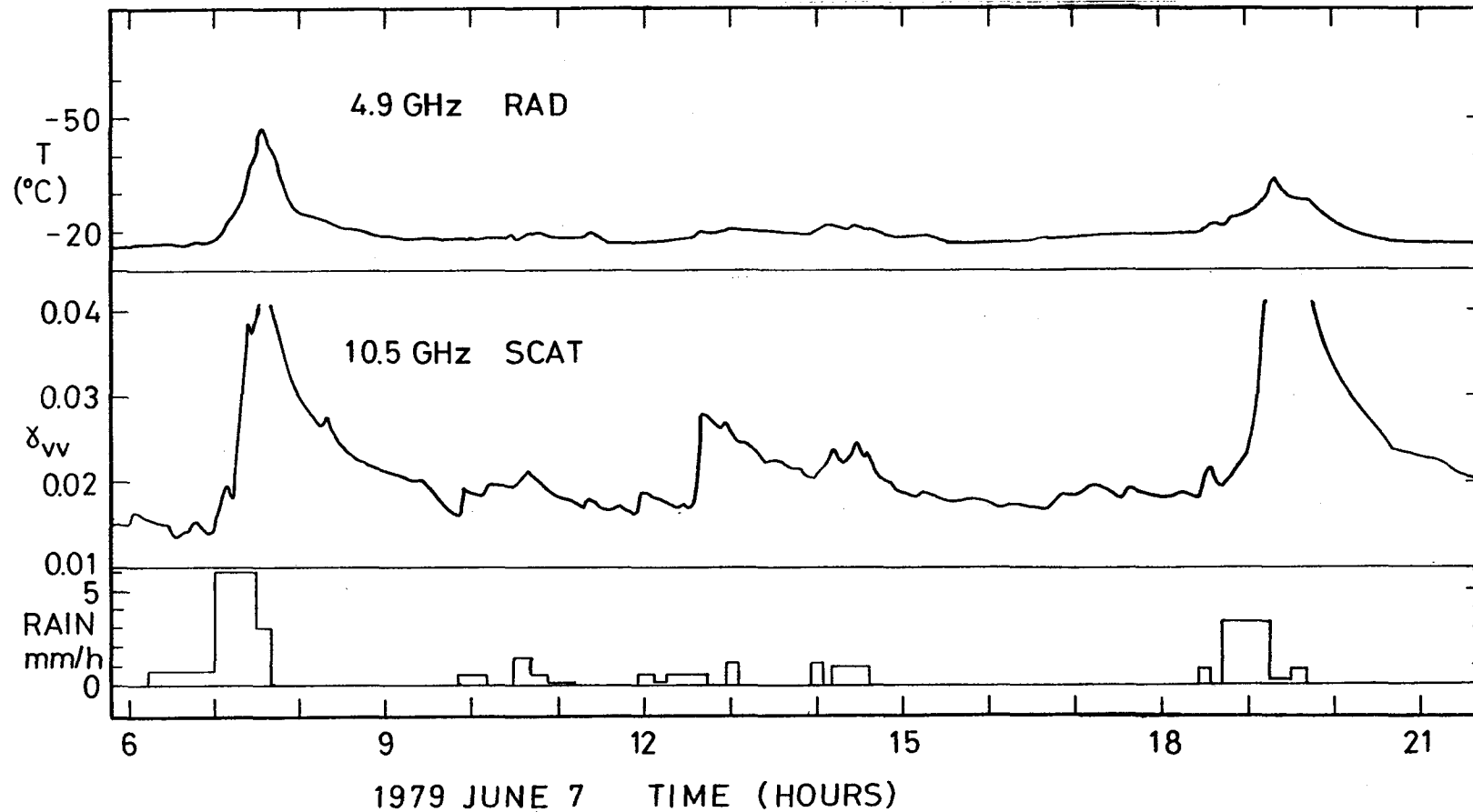
Figure 12   Time profiles on a rainy day of the 4.9 GHz brightness tempera-
ture in degrees Celsius and of the 10.5 GHz backscattering coef-
ficient γ_{vv}, both at 45° nadir angle, vertical polarization, looking on wet
snow. The rain intensity in mm/h is averaged over 10-minute intervals.
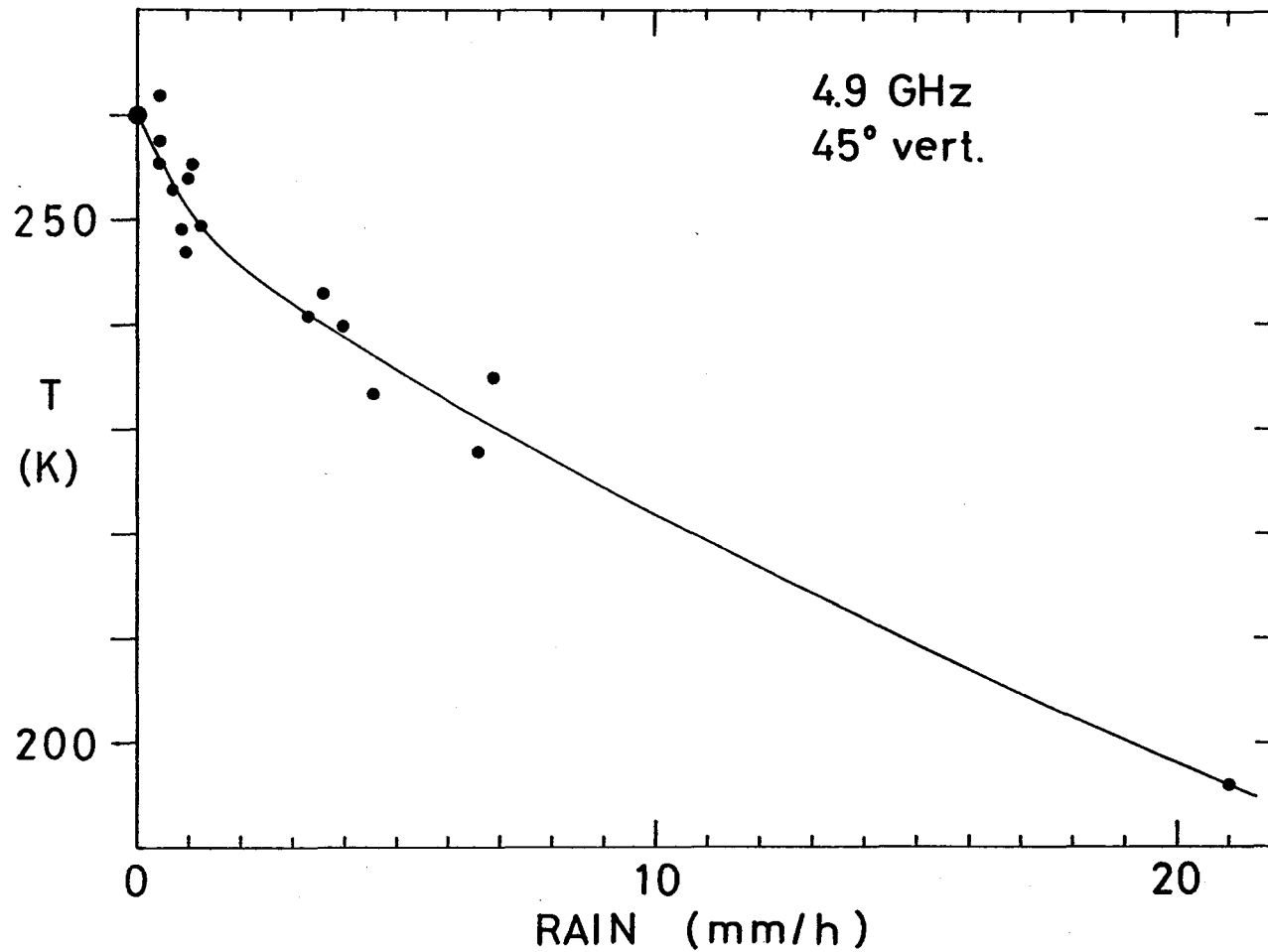
Figure 13   Brightness temperatures at 4.9 GHz, vertical
              polarization, 45° nadir angle of snow at sum-
mer condition versus rain intensity, averaged over 10 minute
intervals. The data were taken from rainfalls with total du-
ration of at least 10 minutes of the period 1979 June 1 to
10.
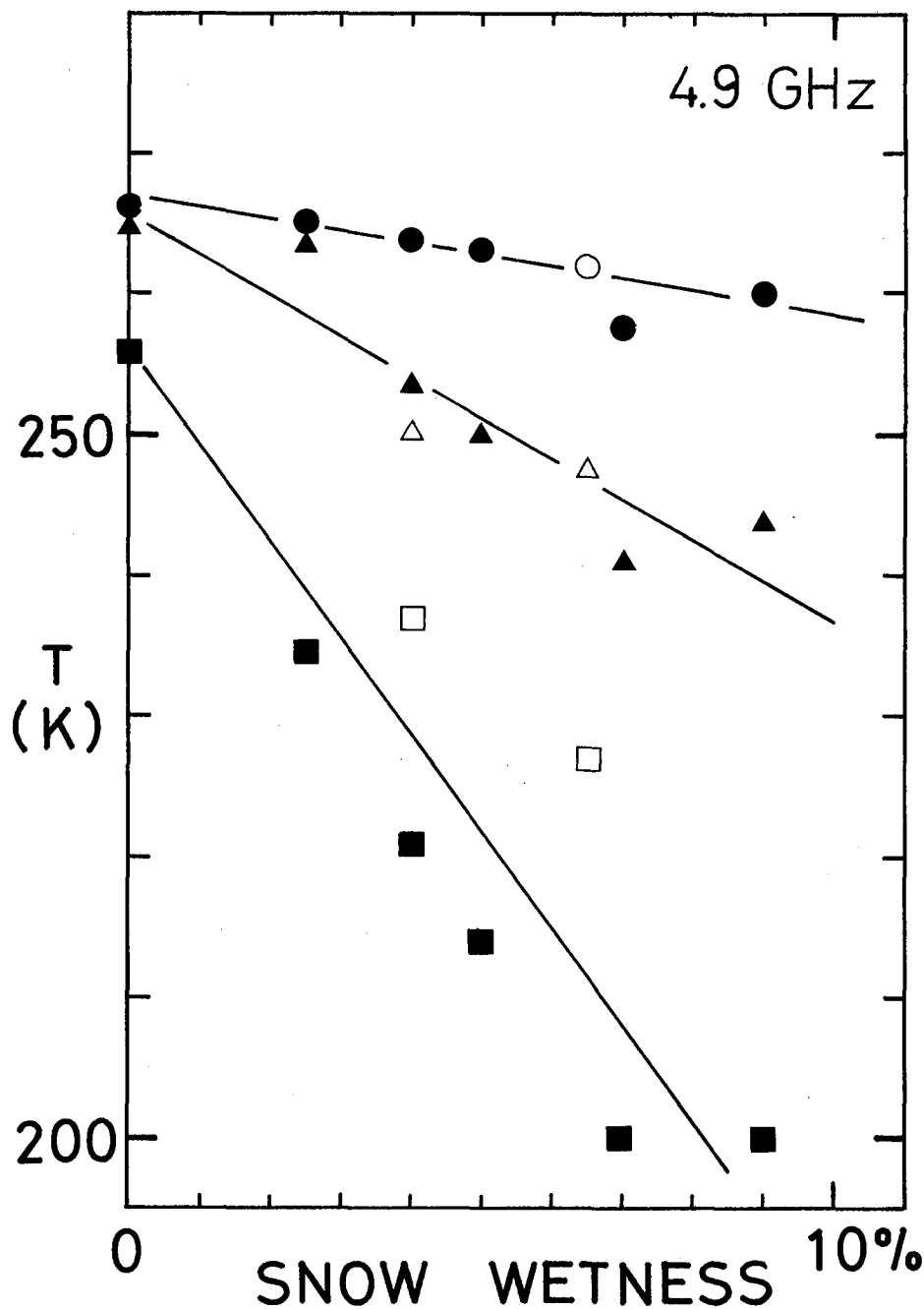
Figure 14   Brightness temperature at 4.9 GHz versus snow
            wetness in percent by volume on melting snow.
Circles are measurements of vertical polarization at 45°
nadir angle, triangles measurement at nadir, and squares
horizontal polarization at 45° nadir angle. Full symbols
are data from 1978 May 31 and open symbols from 1979 May 31.
Snow density at the surface is 0.45 g/cm³ and the snow cry-
stals are spherical with diameters of 1 to 3 mm.

THEORETICAL AND EXPERIMENTAL STUDIES OF MICROWAVE RADIATION

FROM A NATURAL SNOW FIELD

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ABSTRACT

The brightness temperature of a natural snow field in the northern Europe has been studied theoretically and experimentally at 5, 12 and 37 GHz for satellite remote sensing applications. A snow model consisting of ice spheres covered by a water shell has been used in calculation taking into account scattering and absorption.

The brightness temperature of a natural snow field as a function of view angle has been measured from a tower in 1978 and 1979. The measured brightness temperature curves can be fitted with calculated ones by assuming reasonable values for the wetness and the particle size of snow. Experimental results also show that relatively small changes in the snow conditions cause large changes in the brightness temperature.

To obtain a more controlled situation experiments have been continued in 1980 using a measuring site covered with aluminum sheets and determining the wetness and the particle size in addition to the density and physical temperature.

INTRODUCTION

The Nimbus 7 satellite is mapping the Earth's surface using several microwave radiometers in the frequency range 6 to 37 GHz. The European Association of Remote Sensing Laboratories has formed a working group to study the possible application of satellite observations in determining the water content and other characteristics of the snow on the Earth's surface. In this report the brightness temperature of snow field has been studied theoretically and experimentally for obtaining ground truth data in typical Northern European conditions. Similar measurements have been done in Switzerland for the mountain snow [1].

CALCULATION OF THE BRIGHTNESS TEMPERATURE

For calculating the brightness temperature of a snow field a snow model is needed. The snow is assumed to consist of spherical ice particles in the

air. The wetness W (weight percentage of liquid water contained in the snow) is taken into account by surrounding the spheres with a thin water shell. This model can be expected to hold at least for a few days after the deposition [2].

The complex dielectric constant of snow, ε_s, is determined from the dielectric properties of ice and water using the multiphase mixture theory presented by Tinga and Voss [3]:

$$\varepsilon_s = \frac{3 \left[\left(\frac{R_w}{R_a}\right)^3 (\varepsilon_w - 1)(2\varepsilon_w + \varepsilon_i) - \left(\frac{R_i}{R_a}\right)^3 (\varepsilon_w - \varepsilon_i)(2\varepsilon_w + 1) \right]}{(2+\varepsilon_w)(2\varepsilon_w + \varepsilon_i) - 2\left(\frac{R_i}{R_w}\right)^3 (\varepsilon_w - 1)(\varepsilon_w - \varepsilon_i) - \left(\frac{R_w}{R_a}\right)^3 (\varepsilon_w - 1)(2\varepsilon_w + \varepsilon_i) + \left(\frac{R_i}{R_a}\right)^3 (\varepsilon_w - \varepsilon_i)(2\varepsilon_w + 1)} + 1$$

(1)

where i, w, a denotes ice, water and air, respectively. The radius of the water covered ice particle is

$$R_w = R_i \left(1 + \frac{0.92W}{1-W}\right)^{1/3}$$

(2)

and the radius of the air space

$$R_a = R_i \left(\frac{0.92}{\rho_s}\right)^{1/3} \left(1 + \frac{W}{1-W}\right)^{1/3}$$

(3)

where ρ_s is the density of snow. The dielectric constant of water is calculated from the Debye equation:

$$\varepsilon_w = \varepsilon_{w\infty} + \frac{\varepsilon_{w_s} - \varepsilon_{w\infty}}{1 - j2\pi f \tau_w}$$

(4)

where $\varepsilon_{w\infty} = 4.9$, the dielectric constant of water at high frequencies, and $\varepsilon_{w\infty}$ is the static dielectric constant of water:

$$\varepsilon_{w_s} = 87.74 - 0.4008T + 9.398 \cdot 10^{-4}T^2 + 1.410 \cdot 10^{-6}T^3$$

(5)

T is the temperature of water. The real part of the dielectric constant of ice is assumed to be 3.15 in the whole frequency range to be considered. The imaginary part of the dielectric constant of ice is shown in Fig. 1 [4].

Using equation (1) for a two phase mixture the equivalent scattering particle dielectric constant is calculated from the dielectric constant of snow.

The brightness temperature of an infinitely thick snow layer is calculated using the radiative transfer equation

$$\frac{\partial I_p(z,\mu)}{\partial s} = -I_p(z,\mu) + E_p(z) + \frac{\omega\alpha}{2} \sum_{q=H,V} \int_{-1}^{+1} P'_{pq}(\mu,\mu') I_q(z,\mu') d\mu' \tag{6}$$

where $I_p(z,\mu)$ is the radiation intensity in the direction $\phi' = \cos^{-1}\mu$ in the snow, $E_p(z)$ is the self emission of the snow, α is the sum of the absorption coefficient α_a and the scattering coefficient α_s, ω is the scattering albedo α_s/α, and $P'_{pq}(\mu,\mu')$ is the normalized scattering phase function which denotes how much of the radiation in the direction μ' is scattered in the direction μ. The solution of this equation follows that presented by A.W. England [5]. This solution assumes that the scattering is isotropic and that the energy is partioned according to the polarization. The theory could be improved by assuming Rayleigh-scattering and taking into account the coupling between the two polarizations as is done by A.W. England [6]. In practice a different result would be obtained only at large view angles where the difference between the two polarizations will be greater than now calculated.

The theory is expanded for a snow field composed of different homogeneous layers by neglecting reflections between different layers and using the equation:

$$T_B = T_{B_2} e^{-\tau_1} + T_{B_1}(1 - e^{-\tau_1}) \tag{7}$$

where τ_1 is the optical depth of the upper layer. The reflection coefficient between the snow layer and the air is taken into account as well as the reflected sky temperature at 37 GHz. The vertical brightness temperature of the sky at 37 GHz is assumed to be 25 K (clear sky).

Fig. 2 and Fig. 3 show some theoretical results. In Fig. 2 the brightness temperature of an infinitely thick snow layer is shown as a function of frequency. The wetness of snow is assumed to be 3 percent, the density 300 kg/m^3 and the temperature 0°C. It can be seen that the scattering reduces the brightness temperature at higher frequencies. If the snow is wet, snow particles must be relatively large to cause appreciable scattering at frequencies lower than 30 GHz.

Fig. 3 shows the calculated brightness temperature for a snow field (with density 400 kg/m^3 and physical temperature 0°C) as a function of its thickness at the frequencies 5, 12 and 37 GHz. At 5 GHz the ground will show through a dry snow field with a thickness of as much as several tens of meters. At 37 GHz the scattering reduces the brightness temperature of dry snow but for particle diameters larger than 0.5 mm the brightness temperature is not dependent on the thickness of the snow field as long as it exceeds a few decimeters.

EXPERIMENTAL RESULTS

For brightness temperature measurements 5 (4.8) and 37 (36.8) GHz radiometers were used in a tower. Later a 12 (11.6) GHz radiometer was added. The radiometers were looking towards a snow field and their antenna beam (beamwidth 10°) could be turned from straight down (0° angle) to vertical (180°). By turning the radiometers 90° around their axis both vertical and horizontal polarizations could be used. The ground under the snow field was flat within a few centimeters and its brightness temperature was measured before the snow fell. The density of the snow in different layers was measured as well as its physical temperature.

Many results of the measurements have been presented in a report [7]. Fig. 4 shows typical experimental results (solid curves) in the case where the same snow field was measured in the morning after a cold night and in the afternoon when the Sun had been shining several hours. The brightness temperature goes in different directions at 37 GHz and at 5 GHz. At 37 GHz the brightness temperature of the snow with a wet upper surface is much higher than that of the frozen snow due to the high attenuation in the wet layer. At 5 GHz the brightness temperature decreases a little when the snow warms up. This is probably due to the higher reflection coefficient between the wet snow and the air.

The dotted curves in Fig. 4 show the calculated brightness temperatures. The density and the temperature of the snow field were measured and are shown at the bottom of the figure. The wetness and the particle size have been selected for the best fit. It can be seen that a relatively good fit has been achieved for all cases.

In the spring of 1979 the brightness temperature of the snow field was monitored continuously. It was often found that relatively small changes in the snow conditions caused large changes in the brightness temperature. Fig. 5 shows the variation of the brightness temperature when at first a very light rain occurred and then about two hours later some 20 mm of light new snow fell.

The brightness temperature measurements of the snow field have been continued in the spring of 1980 at frequencies 5, 12 and 37 GHz. To obtain a more controlled situation for comparisons of calculted and measured data the ground at the measuring site has been covered with an aluminium surface giving a cold background without snow. Because the brightness temperature is strongly dependent on the wetness and the particle size, these have also been measured in addition to the density and physical temperature. Fig. 6 shows the calculated brightness temperature of a snow layer on the aluminum sheet as a function of the thickness. The preliminary experimental results agree reasonably well with the calculated ones.

CONCLUSIONS

It has been shown that theoretically calculated and experimentally measured brightness temperatures of a snow field can be fitted when the density and the temperature of the snow are known and the particle size and the wetness are selected.

For remote sensing applications in Northern Europe radiometer frequencies from 5 to 35 GHz seem to be appropriate. At frequencies over 35 GHz only the surface of the snow field can be seen. Frequencies lower than 5 GHz are needed for very thick snow. In the spring it can be expected that relatively small changes in the snow conditions (e.g. a thin wet layer on the surface or a very thin layer of new snow) cause considerable variations in the brightness temperature.

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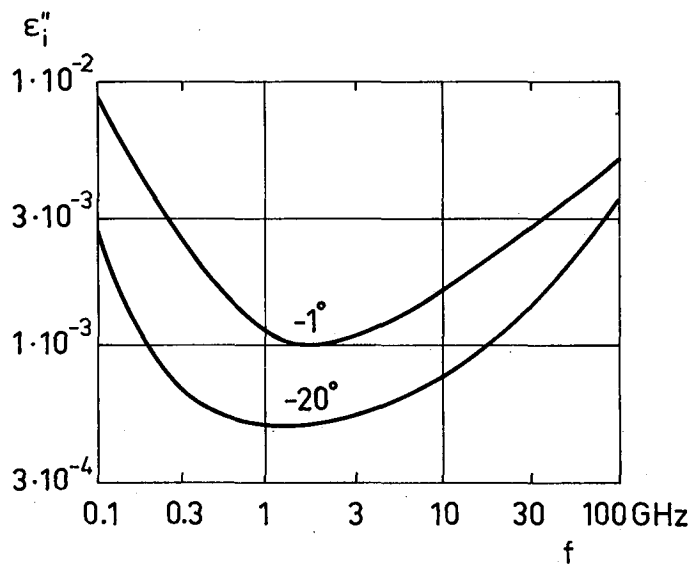
Figure 1. Imaginary part of the dielectric constant of fresh water ice as a function of frequency [4].
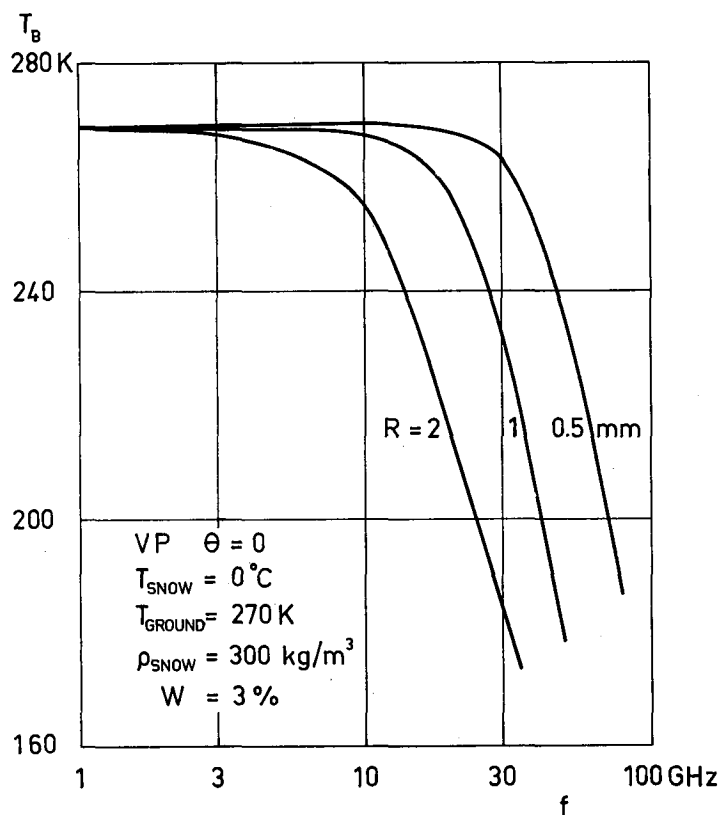
Figure 2. Theoretical brightness temperature of an infinitely thick snow layer as a function of frequency. R is the radius of the snow particle.
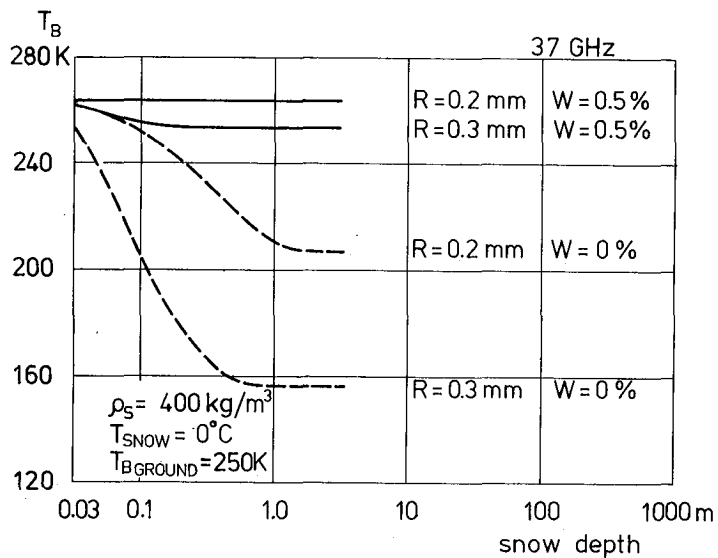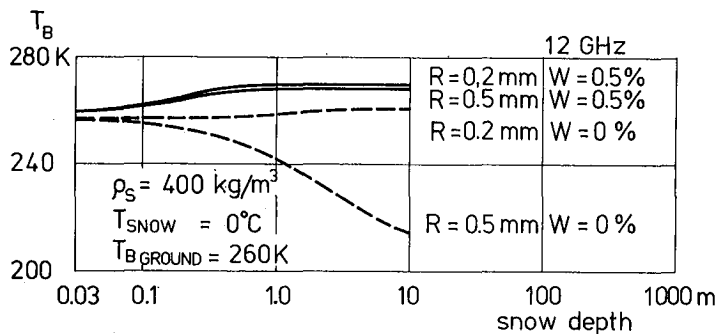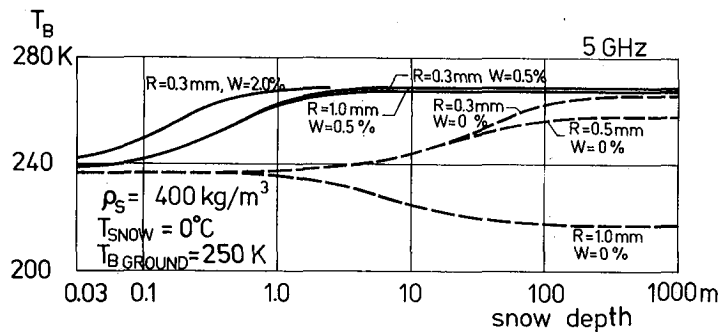
Figure 3. Theoretical brightness temperatures of a snow field at 5, 12 and 37 GHz. The density of snow is 400 kg/m³ and its temperature 0°C. The wetness W and the radius R of the snow particle are parameters.
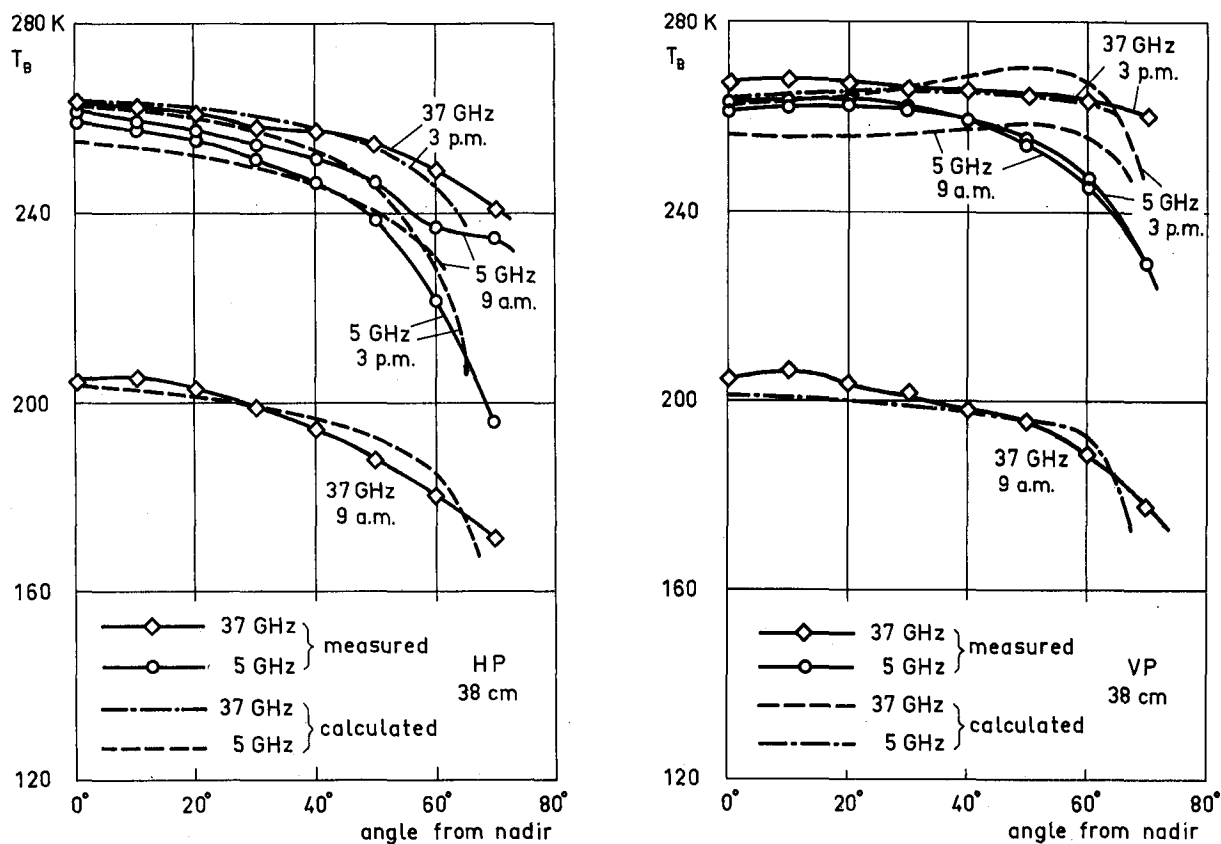
Thickness	Density	Wetness	Temperature	Particle radius	Optical depth	Diel.Re.	Diel.Im.	Albedo
(m)	(kg/m³)	(%)	(C)	(mm)	4.8/36.8	4.8/36.8	4.8/36.8	4.8/36.8
0.2	350	0.28	-0.1	0.5	0.17/6.30	1.59/1.57	0.011/0,0064	0.010/0.87
0.18	410	1.1	-0.3	0.6	0.66/13.1	1.78/1.70	0.048/0.029	0.0060/0.76

Thickness	Density	Wetness	Temperature	Particle radius	Optical depth	Diel.Re.	Diel.Im.	Albedo
(m)	(kg/m³)	(%)	(C)	(mm)	4.8/36.8	4.8/36.8	4.8/36.8	4.8/36.8
0.01	350	10	0	0.5	0.10/1.50	2.07/1.67	0.15/0.19	0.0027/0.26
0.19	350	6	0	0.5	1.74/19.9	1.92/1.62	0.13/0.12	0.0023/0.31
0.18	410	2	-0.3	0.6	1.07/15.8	1.86/1.71	0.080/0.052	0.0044/0.65

Figure 4. Measured and calculated brightness temperatures of a 38 cm thick snow field in the morning and in the afternoon. The density and the temperature of snow were known. The wetness and the particle size have been selected for best fit. The air temperature in the morning was -1°C and in the afternoon +4°C. HP = horizontal polarization, VP = vertical polarization.

T_B
280 K

snow depth 63 cm

240

—— 5 GHz VP
--- 12 GHz VP
······ 37 GHz HP

200

$\theta = 50°$

light rain   snow fall (20 mm)

T_{AIR}
0

-5°C

14   16   18   20   22   24   2
1979-03-27   LOCAL TIME
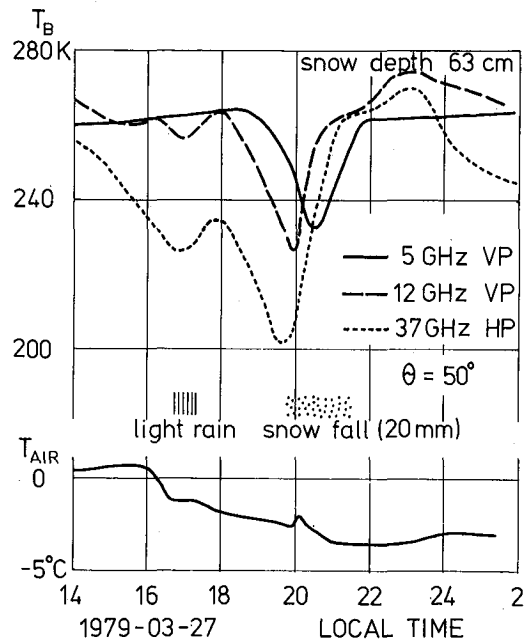
Figure 5. Measured brightness temperature of a snow field and the air temperature as a function of time at frequencies 5, 12 and 37 GHz. The snow was relatively homogeneous and fine-grained. The look angle was 50°.

1979-12-01
M. TIURI

T_B
275 K ω

12 GHz
—— R = 0.4 mm
--- R = 0.2 mm
$\rho_s = 300$ kg/m^3
$T_s = -2°C$
$T_{GR} = 0$ K

0.04

200

0.08
0.01

0.14
0.02

10
5
100
2
1
0.5
0.2
0.1

0 percent

0.72
0.24

0

0
0.01   0.025   0.05   0.1   0.25   0.5   1 m
Thickness

1979-12-01
M. TIURI

T_B
275 K

37 GHz
—— R = 0.4 mm
--- R = 0.2 mm
$\rho_s = 300$ kg/m^3
$T_s = -2°C$
$T_{GR} = 0$ K

0.07
0.23
0.39

0.70
0.55

200

0.91
0.85

2
0.5
0.1
100
2
0
0.5
0.1
0 percent

0.98

0
0.01   0.025   0.05   0.1   0.25   0.5   1 m
Thickness

Figure 6. Theoretical brightness temperature of a snow layer on the aluminum sheet at 12 and 37 GHz as a function of thickness. The wetness varies from 0 to 10 percent. ω is the scattering albedo.

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MONITORING SNOWPACK PROPERTIES BY PASSIVE MICROWAVE SENSORS

ON BOARD OF AIRCRAFT AND SATELLITES

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ABSTRACT

Snowpack properties such as water equivalent and snow wetness may be
inferred from variations in measured microwave brightness temperatures. This
is because the emerged microwave radiation interacts directly with snow
crystals within the snowpack. Using vertically and horizontally polarized
brightness temperatures obtained from the Multifrequency Microwave Radiometer
(MFMR) on board a NASA research aircraft and the Electrical Scanning Micro-
wave Radiometer (ESMR) and Scanning Multichannel Microwave Radiometer (SMMR)
on board the Nimbus-5, -6, and -7 satellites, linear relationships between
snow depth or water equivalent and microwave brightness temperature were
developed. The presence of melt water in the snowpack generally increases
the brightness temperatures, which can be used to predict snowpack priming
and timing of runoff.

INTRODUCTION

The use of remotely-acquired microwave data, in conjunction with essential
ground measurements will most likely lead to improved information extraction
regarding snowpack properties beyond that available by conventional techniques.
Landsat visible and near-infrared satellite data have recently come into near
operational use for performing snowcovered area measurements (Rango, 1975;
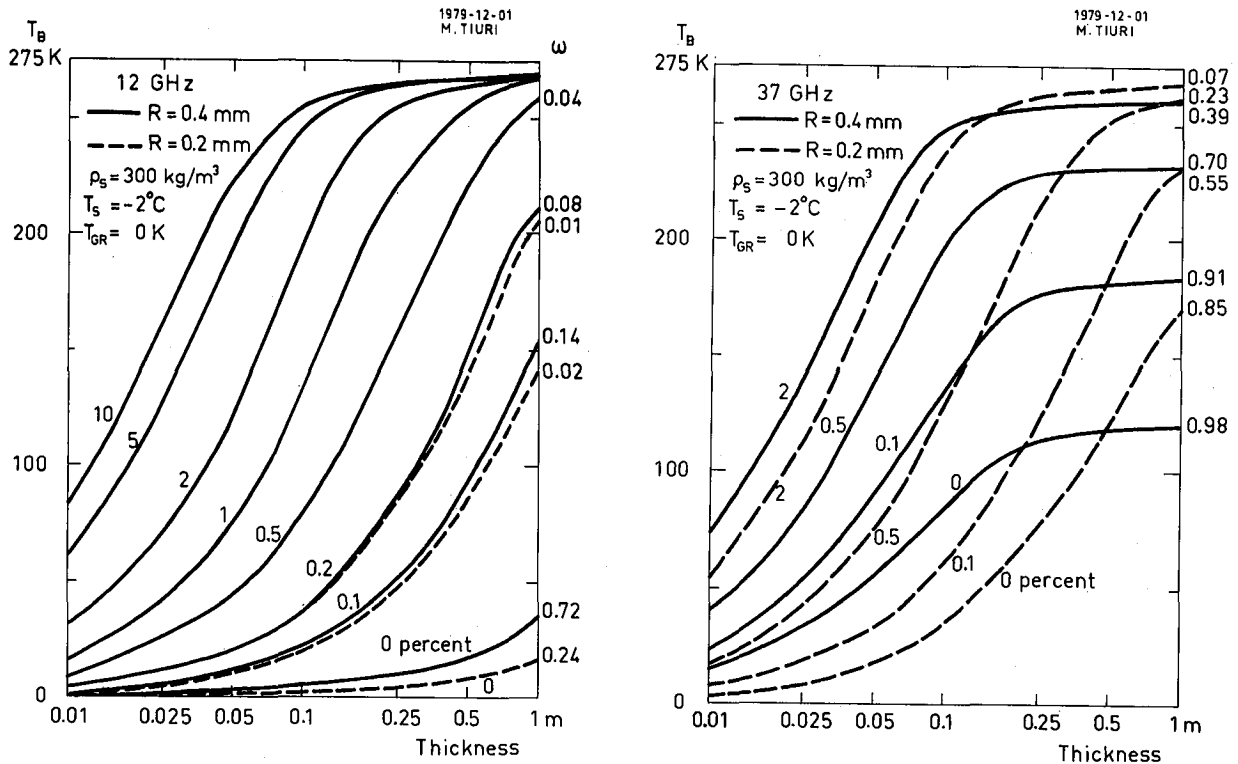1978). However, Landsat data acquisition is hampered by cloudcover, sometimes
at critical times when a snowpack is ripe. Furthermore, information on water
equivalent, liquid water content, and other snowpack properties germane to
accurate runoff predictions is not currently obtainable using Landsat data
alone, because only surface and very near-surface reflectances have been
detected.

Microwaves are mostly unaffected by clouds and can penetrate through
various snow depths, depending on the wavelength. Hence, microwave sensors
are potentially capable of determining the internal snowpack properties such
as snow depth and snow water equivalent (Hall et al, 1978; Rango et al., 1979).
However, operational use of remotely-collected microwave data for snowpack
analysis is not imminent because of complexities involved in extracting in-
formation from measured data. Snowpack and soil properties are highly
variable, and their effects on microwave emission are still being explored.
Nevertheless much work is being done to develop both active microwave

(Hoekstra and Spanogle, 1972; Ellerbruch et al., 1977; Stiles and Ulaby, 1980; and Ulaby and Stiles, 1980) and passive microwave techniques (Edgerton et al., 1971; Schmugge et al., 1974; Schmugge, 1973; Linlor et al., 1974; Chang et al., 1976; Chang and Shiue, 1979; Kong et al., 1979; Hofer and Mätzler, 1980) for analysis of snowpack properties. Passive microwave data obtained during recent flights by NASA aircraft and measurements made by ESMR on board the Nimbus-5 and -6 and SMMR on board the Nimbus-7 will be analyzed and compared with snowpack depth information.

PASSIVE MICROWAVE EXPERIMENTS

During the winter of 1976, 1977, and 1979 the NASA aircraft equipped with MFMR, Passive Microwave Imaging System (PMIS), and other support instruments was flown over test sites near Steamboat Springs and Walden, Colorado. Ground truth data including snow depth and temperature, free water content, density, structure, and soil mositure were taken along the flight lines.

Passive microwave data from space have been available since December 1972 when Nimbus-5 was launched with the ESMR onboard sensing at the 1.55 cm wavelength. Further data became available in June 1975 with the launch of Nimbus-6 with an ESMR instrument capable of receiving dual-polarized microwave radiation from the earth at 0.81 cm wavelength. Due to the coarse spatial resolution of these two instruments, a large homogeneous area in Canada was selected for this study. Nimbus-7 SMMR was launched in October 1978, measuring microwave radiations at five frequencies and dual polarization (0.8, 1.4, 1.7, 2.8, and 4.6 cm). Two test sites (in Russia and in Canada) were selected for this study.

INTERPRETATION OF MICROWAVE EMISSION FROM SNOW

Snow particles act as scattering centers for microwave radiation. Computational results indicate that scattering from individual snow particles within a snowpack is the dominant source of upwelling emission in the case of dry snow. This type of radiation upwelling through snow is governed by Mie scattering theories for which a good description can be found in Chang et al. (1976). Microwave radiation emanating from snow originates from a depth of ~10-100 times the wavelength used. However, when the snowpack thickness is less than the microwave penetration, the underlying surface will contribute to the T_B (Chang and Gloersen, 1975).

Using the multifrequency analysis approach, one can make inferences regarding not only the thickness of the snowpack, but the moisture conditions and the condition of the underlying soil (wet versus dry). The shorter wavelengths such as the 0.8 cm, sense near-surface temperature and emissivity, and surface roughness. At the intermediate wavelengths, 1.4 and 1.7 cm, the radiation is less affected by the surface, and more information is obtained on the characteristics of the mid-pack. Longer wavelengths such as 21 cm, represent greater penetration through a snowpack and receive a strong contribution of emission from the underlying ground. All of the above generalizations apply to the snow depths encountered at Steamboat Springs and Walden during the study period.

In addition to snow depths, snow grain and crystal sizes, ice lenses and layers within the snowpacks were measured in the snow pits. Grains, crystals, lenses, and layers act as scatterers to the microwave radiation if their size is comparable to the wavelength. Short wavelength radiation tends to be scattered by snow crystals and grains (~ 1 mm) which are comparable to the wavelength, as well as by the larger ones. Longer wavelengths are not affected by the very fine crystals and grains, but will be affected by lenses and layers, the result of snow metamorphism.

The presence of liquid water in the snowpack and the condition of the ground below the pack were also measured. Liquid water in snow (5 percent by weight) causes a sharp increase in the T_B (Chang and Gloersen, 1975). This is because the effects of scattering of individual snow particles are reduced when liquid water coats the crystals, and emission increases.

The condition of the ground beneath the snow will determine the intensity of the radiation incident from below. Dry or frozen ground has a high emissivity (~ 0.90-0.95) with a T_B of $\sim 260°$K, whereas unfrozen wet ground has a much lower emissivity (~ 0.7) with brightness temperatures as low as 150°K. Knowledge of the condition of the ground underlying the snow is important for the interpretation of observed brightness temperatures and can generally be determined from the 21 cm observations.

OBSERVATIONAL RESULTS

SNOW DEPTH

Table 1 shows the various snow depths and average wetness conditions of the snow encountered at the two sites in 1976 and 1977. When the snowpack is dry (< 1 percent liquid water present), the T_B should decrease with increased snow depth, as shown in figure 1. Figure 1 illustrates the responses of the 0.8 and 1.4 cm channels of the MFMR to the various snow depths shown in table 1.

The greater T_B decrease evident in the plot of the 0.8 cm channel (solid line in figure 1) is due to the fact that more particles are present which can scatter the 0.8 cm radiation than the 1.4 cm radiation (dashed line) because of the size range of particles within a snowpack. A deep snowpack obviously has more crystals and/or grains than does a shallow pack. Crystals and grains large enough to scatter the 1.4 cm and longer wavelength emission are inherently fewer.

Recently, studies have been performed to determine relationships between the snow depth and brightness temperature measured by ESMR on board the Nimbus-5 and -6 satellites and by SMMR on board the Nimbus-7 satellite. The areas studied were homogeneous areas located on the Canadian high plains in southern Alberta and Saskatchewan and a winter wheat area in Central Russia. Figure 2 illustrates the snow depth versus brightness temperature data for Nimbus-5 and the resulting significant (at the .002 level) regression line

TABLE 1. AVERAGE SNOW AND GROUND CONDITIONS AT THE STUDY AREAS

	depth cm	snow condition	ground condition
March 1976			
Steamboat Springs	75.4	dry	wet
Walden	10.5	moist	frozen
January 1977			
Steamboat Springs	36.1	dry	frozen
Walden	trace	N/A	frozen
March 1977			
Steamboat Springs	41.1	dry to moist	frozen
Walden	2.5	moist	wet

and statistics. The Nimbus-5 data are from the nighttime pass on March 14, 1976, and the snow depth data are from March 15, 1976. Air temperatures prior to March 15 were well below 0°C with little chance of significant melting, and as a result, dry snow conditions were assumed. Figure 3 presents a comparable plot for snow depth and Nimbus-6 vertically-polarized brightness temperature from the daytime pass on March 15, 1976. It does appear that in simple regression analysis that the Nimbus-6 data produce better relationships than the Nimbus-5 data, probably because the emission from the relatively thin snow-cover at 1.55 cm contains a more significant contribution from the variable underlying soil layer than at 0.81 cm.

Figure 4 presents a plot for snow depth and the brightness temperature of 0.81 cm horizontal polarization from Nimbus-7 SMMR over Central Russia. The Nimbus-7 data are from the nighttime passes between February 23 and 25, 1979 and the snow depth data are from February 24, 1979. The slope of the linear relationship varies slightly with different geographic locations due to different snow and physiographic conditions. The sensitivity of the measurements are approximately 2K per cm of snow for 0.81 cm wavelength and 1.2K per cm for 1.55 cm wavelength.

SNOWPACK AND SOIL MOISTURE CONDITIONS

The response of the MFMR data to snow moisture has also been analyzed. Snow wetness is very important to runoff forecasting, as is the condition (wet or dry) of the underlying ground. Variations in snow moisture have been measured using a freezing calorimeter technique during the 1976 and 1977

aircraft experiments, and it has been found that free water in a snowpack will raise the 0.8 cm T_B. Note the peak in the response of the 0.8 channel to the 41.1 cm depth snow in figure 1 (open circle ~250°K). This is the March 1977 snowpack at Steamboat Springs. The peak is apparently caused by surface moisture to which the 0.8 cm radiation is very sensitive. The longer wavelengths did not respond as markedly (i.e. show the sharp T_B increase) because they emanate from deeper, drier layers within the snowpack. If all wavelengths were to show the peak, theoretically the snowpack would be ripe.

Figure 5 compares the responses of all four wavelengths over frozen ground in January, 1977 (solid line) to that over shallow (2.5 cm), moist snow and wet ground at Walden in March 1977 (dashed line). The shortest wavelengths, 0.8 and 1.4 cm, have slightly higher average brightness temperatures for moist snow (March) than for the frozen ground (January). The 1.7 cm channel shows approximately the same T_B for frozen ground and moist snow, while the difference in the 21 cm T_B between moist snow and frozen ground is 47°K. The 21.0 cm radiation is apparently unaffected by the moist snow because the snow is so shallow. The low T_B of the 21 cm wavelength in March results from the wet ground beneath the snow.

Figure 6 shows the relationship of snow depth and the measured 0.81 cm brightness temperature when the snowpack is melting. The slope of the linear curve reversed as compared with figures 3 and 4. This is consistent with the results from aircraft and truck measurements.

SUMMARY AND CONCLUSIONS

It has been demonstrated that there are differences in the microwave brightness temperatures for the snowpacks studied at Walden and Steamboat Springs, Colorado during the 1976 and 1977 experiments. An average T_B decrease for the shorter wavelengths (0.8, 1.4, and 1.7 cm) of 35°K has been shown to correspond with a 39.3 cm greater snow depth for the March 1976 as compared to the January 1977 Steamboat Springs snowpack. A T_B decrease of 50°K for the 21 cm wavelength is attributed to wet soil conditions in March 1976. Furthermore, a sharp rise, ~49°K, in the 0.8 cm T_B corresponds to moist snow on the surface of the Walden snowpack in March 1977 demonstrating the sensitivity of microwave radiation to moist snow. Also, a greater T_B decrease for a given snowpack is evident for the short, 0.8 cm, as compared to the longer, 1.4 cm, wavelength. This is because shorter wavelength radiation is more scattered than longer wavelength radiation, resulting in a lower emissivity and a lower T_B for the short wavelengths. A dry snowpack has particle sizes typically <0.1 cm. As the wavelength of the radiation approaches the particle size, the scattering will increase. This greater scattering lowers the T_B and the emissivity of a snowpack.

Snow depth, liquid water within the pack, and underlying conditions were addressed in this paper. Varying conditions of these parameters were encountered in the study areas and subsequent correlations made with the microwave data were consistent between different measurements. The brightness temperatures obtained by ESMR and SMMR for relatively homogeneous test areas

showed a significant regression relationship with snow depth has been developed. The brightness temperature decreases with increasing snow depth when the snow is dry. When the snowpack is melting, the inverse relationship has been obtained. Further study will be performed to obtain similar specific relationships for other snow study areas with greater snow depth ranges and different snow conditions.

The challenge in the analysis of the microwave response to snowpack properties lies in the fact that snowpack conditions are complex, and their interaction with microwave radiation is not completely understood. Because snowpack character can change so rapidly, and, is in fact constantly changing, a complicating factor is added to data analysis. It is believed that with additional measurements in the coming years a more quantitative relationship between brightness temperatures and snow depth will be possible for snowpacks of known wetness condition. This understanding of microwave emission from snow will be used to define an improved system for snowpack monitoring from a remote platform.

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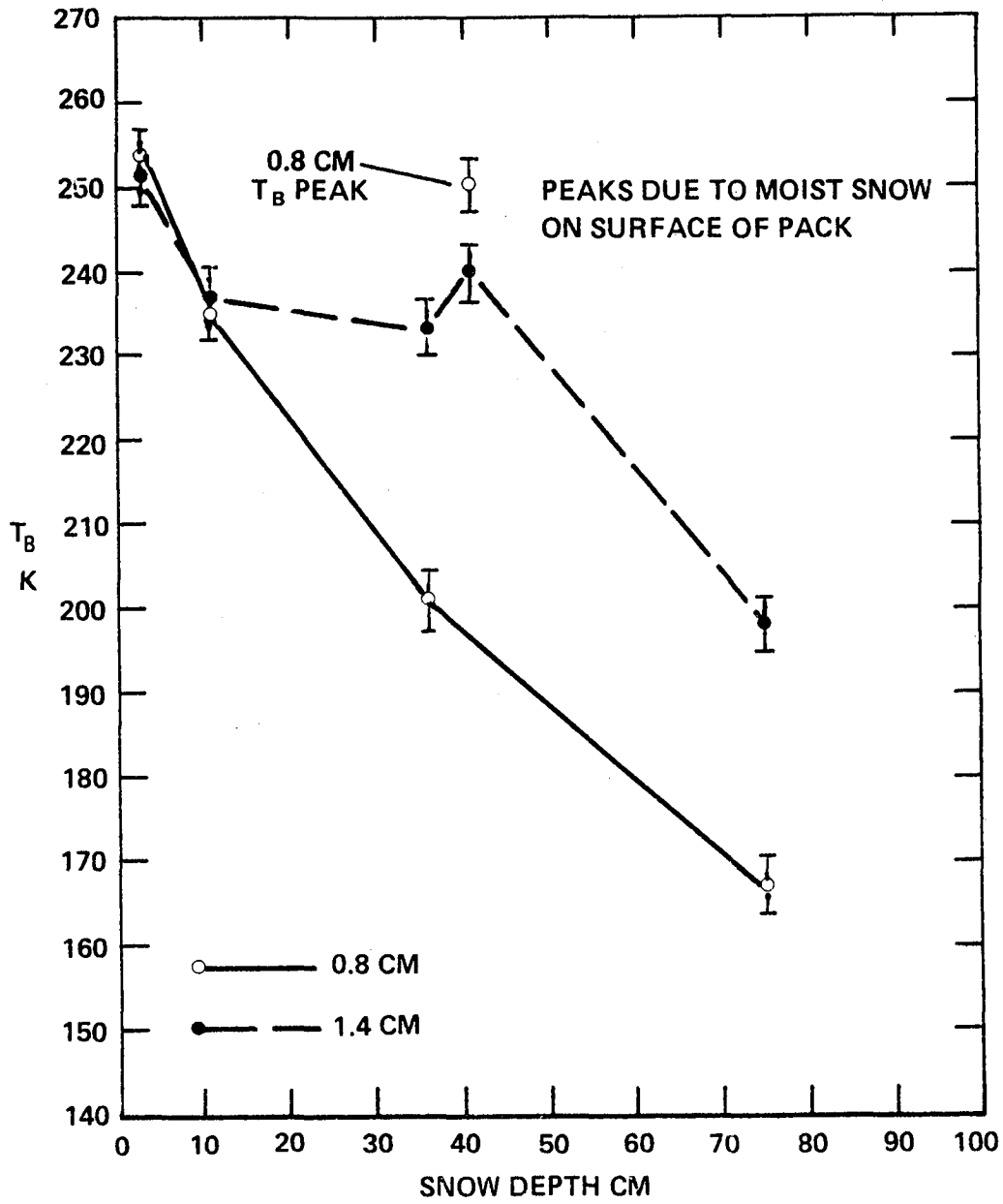
Figure 1. Horizontally Polarized Microwave T_B Responses
to Snow Depths
(from Hall et al., 1978)
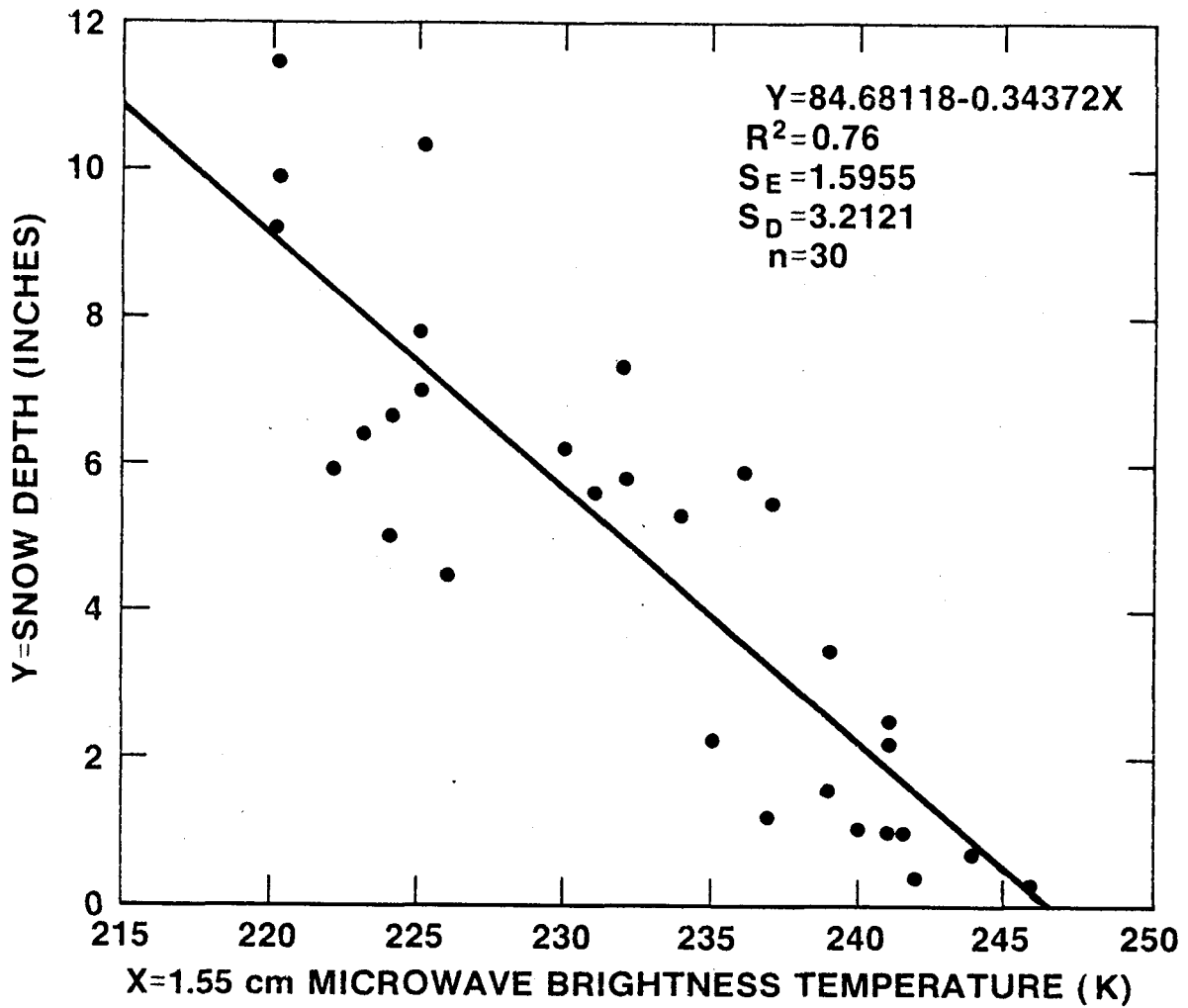
Figure 2. Nimbus-5 Microwave Brightness Temperature
versus Snow Depth on the Canadian High Plains
(from Rango et al., 1979)
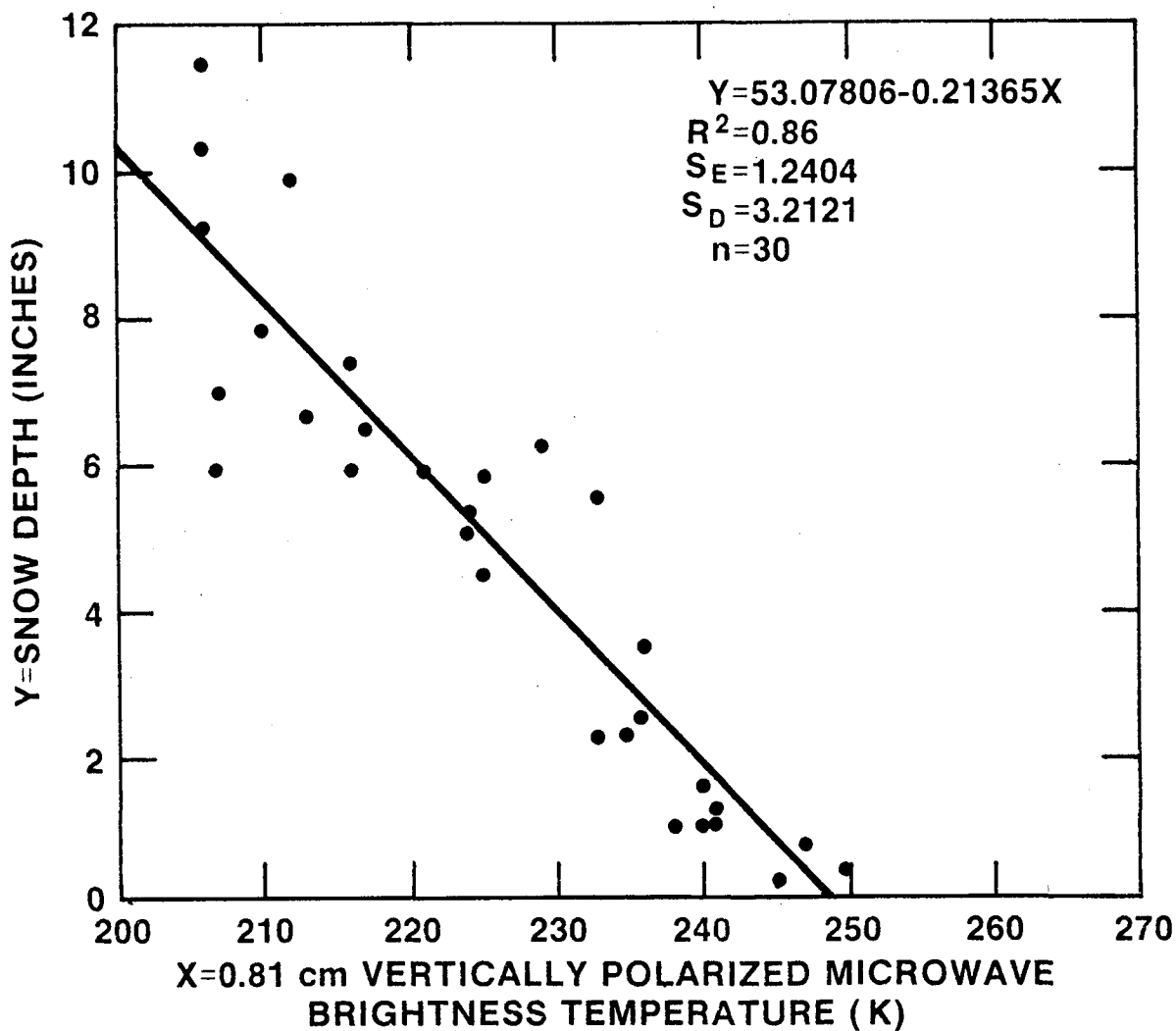
Figure 3. Nimbus-6 Vertically Polarized Microwave Brightness Temperature versus Snow Depth on the Canadian High Plains (from Rango et al., 1979)
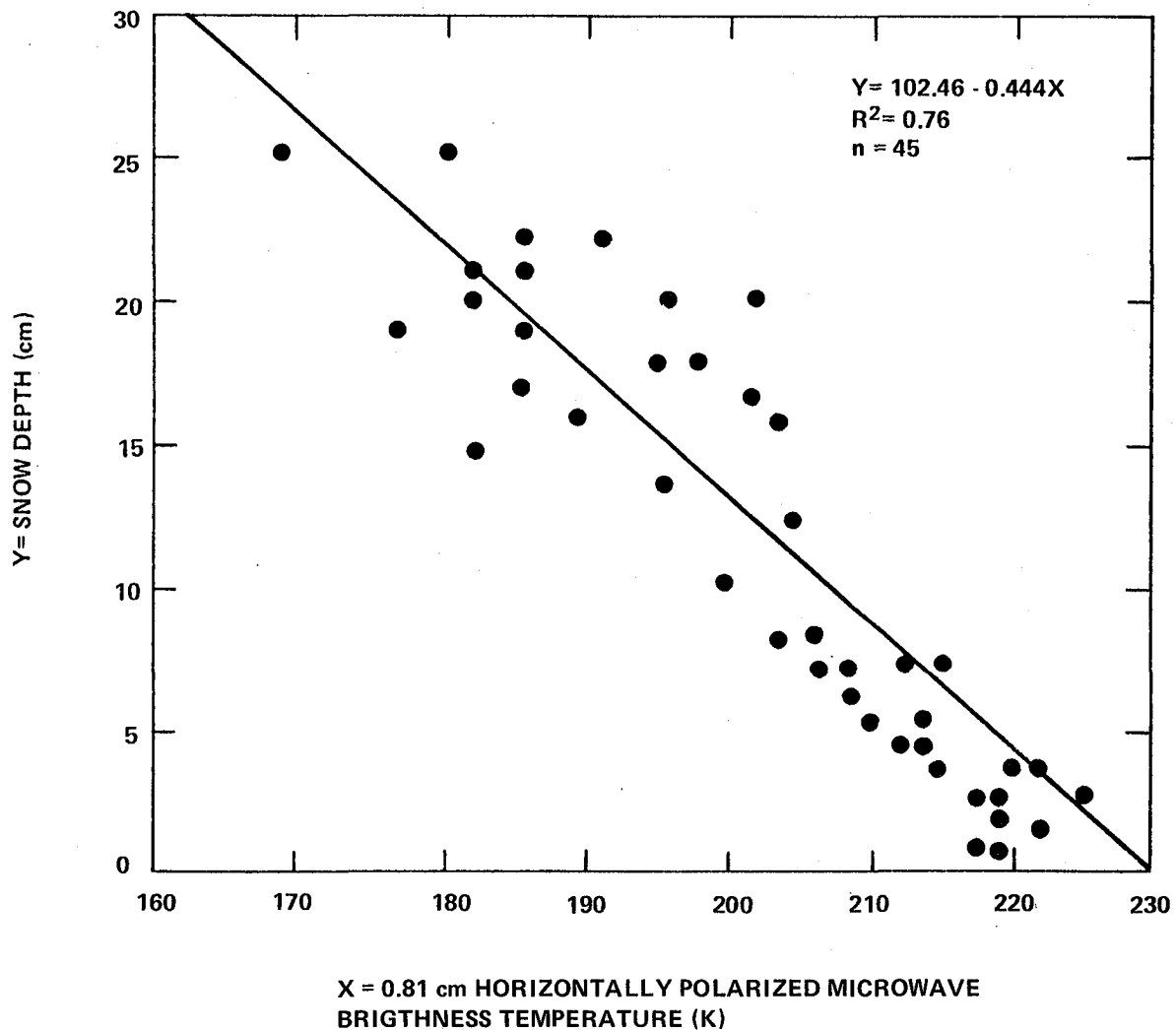
Figure 4. Nimbus-7 0.81 cm Horizontally-Polarized Microwave Brightness
Temperature versus Snow Depth over Central Russia
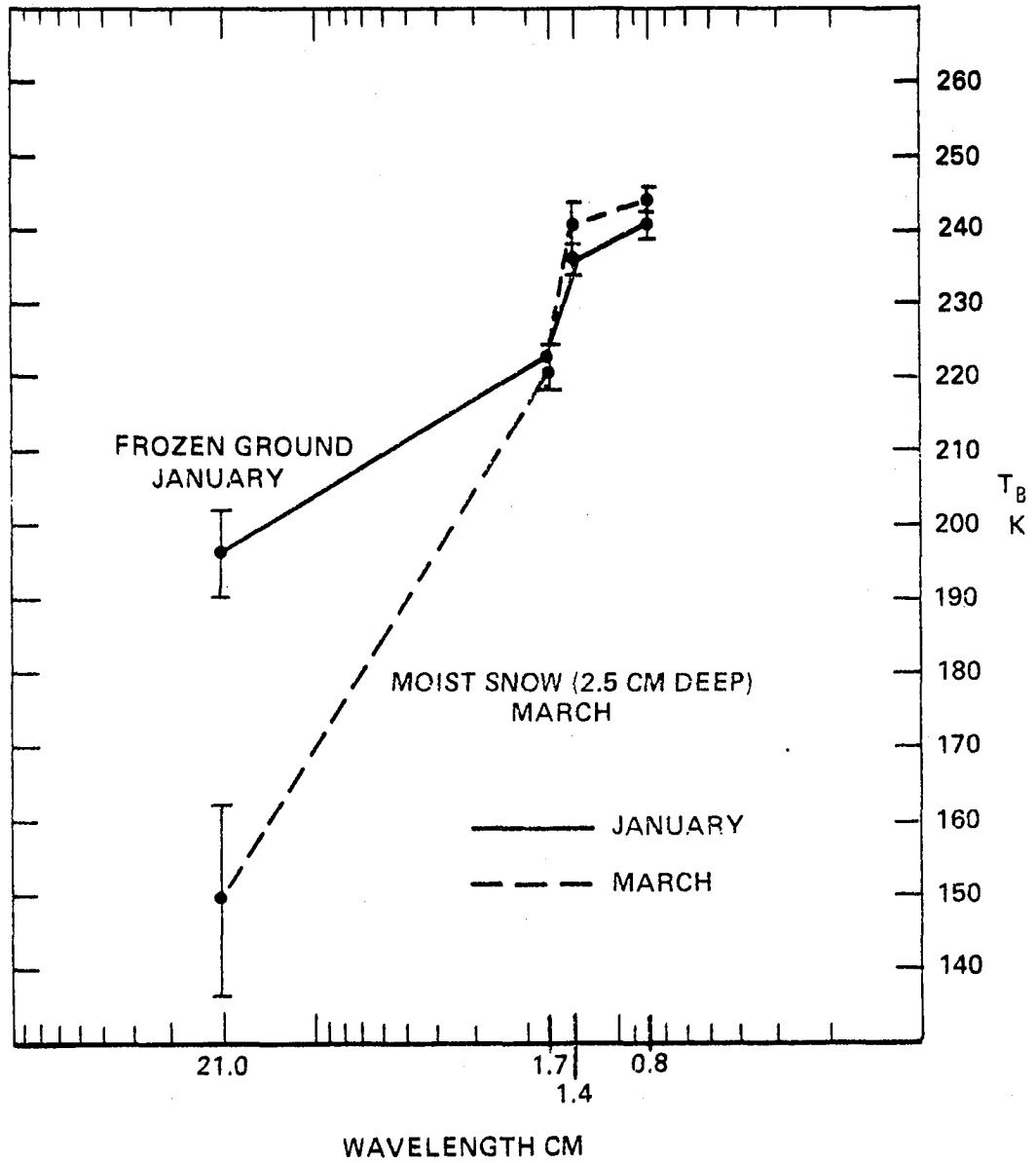
Figure 5.   Variation of Microwave T_B with Radiometer Wavelength,
Walden, Colorado, 1977
(from Hall et al., 1978)
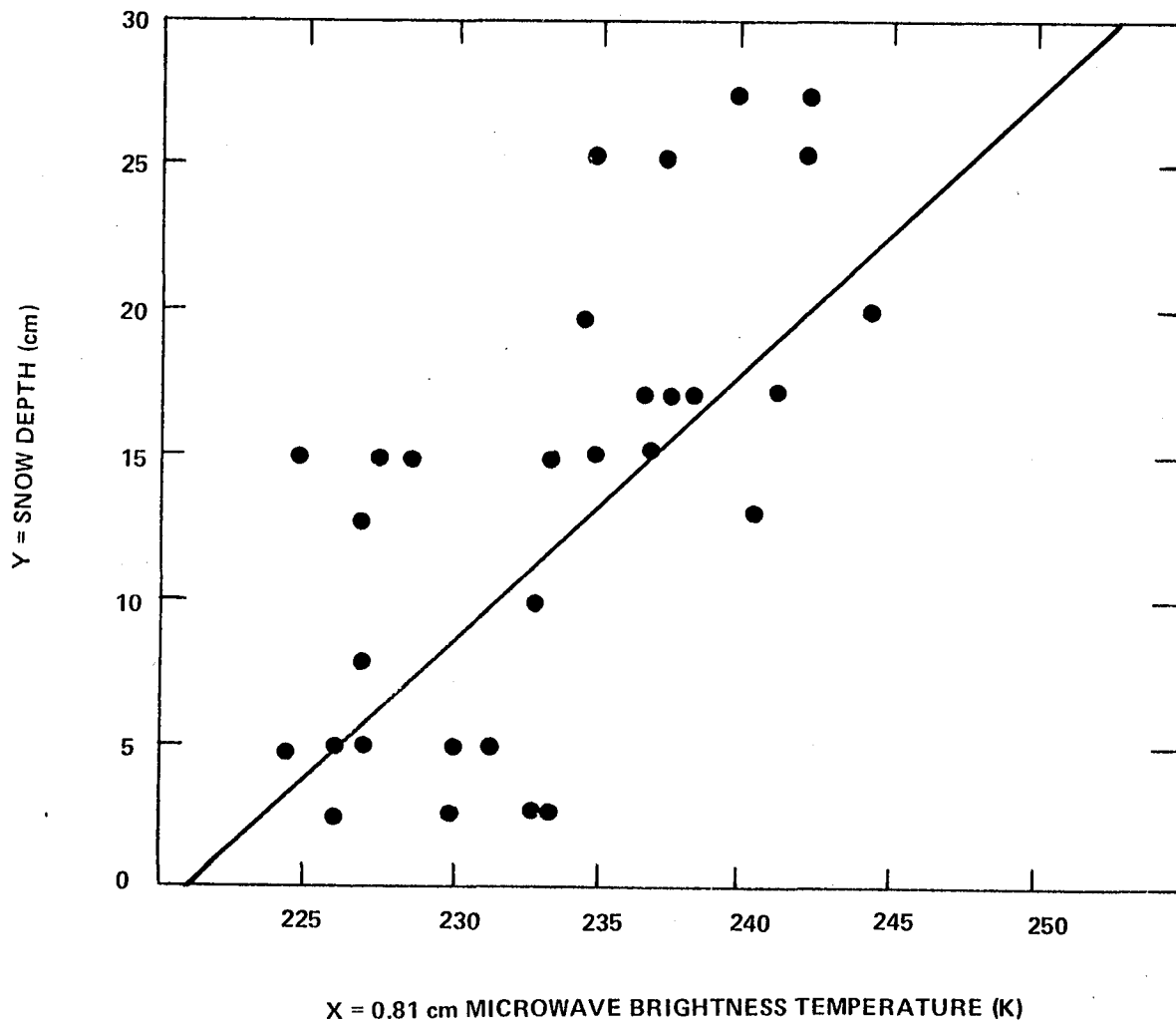
Figure 6. Nimbus-6 Horizontally-Polarized Microwave Brightness Temperature versus Snow Depth on the Canadian High Plains During Melt.

DISCUSSION SESSION ON A MICROWAVE REMOTE SENSING OF

SNOWPACK PROPERTIES RESEARCH PLAN

ATTENDEES

D. J. Angelakos

A. Denoth

J. C. Barnes

R. E. Bates

H. S. Boyne

A. J. Brown

M. A. Calabrese

A. Chang

S. Colbeck

R. Crane

J. Dozier

H. Gubler

R. Hofer

K. Ishihara

E. B. Jones

W. Linlor

D. F. McGinnis Jr.

C. Mätzler

N. Peterson

R. Peterson

A. Rango, Chairman

G. Reinhardt

B. Shafer

W. H. Stiles

F. T. Ulaby

W. O. Willis

INTRODUCTION

The Workshop on the Microwave Remote Sensing of Snowpack Properties featured the exchange of much valuable information between investigators accentuated by lively discussion sessions. Significant discussion took place during specially reserved periods after the presentation of a block of papers and during the coffee breaks. In addition one entire morning was devoted to discussion of issues brought up during the Workshop. Finally on the afternoon of the last day, May 22, 1980, a smaller group of investigators discussed the possible formulation of a snowpack properties research plan to be used to guide NASA research in this rapidly developing field. The following transcription documents the results of these discussions.

Albert Rango, Chairman

DISCUSSION

RANGO: This group is very similar to the soil moisture group. There was a working group convened there that came in at about the same phase that we are in here in snowpack properties. At that time we met approximately every six months for about a year and a half or so to generate a plan, and the plan is basically finished. There are only a few changes to make in it. It is used to guide NASA research in soil moisture for agriculture, water and climate. I would think that we could do something similar. At this particular time, I think first of all, we should decide what we want to do, second, how to go about it, and third, what should be in such a plan. So I think we are open to suggestions, and I think anybody who feels moved to come up with something as a starting point to start off the discussions should

do so. Maybe Mike Calabrese would have some information, that would help us a little bit, as to how they would use it at NASA Headquarters.

CALABRESE: Well I think, as I indicated before, that if we are to get significant resources committed to this area, we are going to have to: A) make the case regarding the importance of snowpack properties applications and the need for remote sensing research and, B) develop a research plan. The plan, should address the applications such as hydrology and agriculture and their requirements to plead our case for support to move ahead with a program for some finite timeline such as 5 years. The plan should also contain a set of research objectives to test and demonstrate, perhaps on a basin basis, that these measurements can be obtained, and that the techniques work.

In addressing the remote sensing of snowpack properties, I think we are in a more formative stage than soil moisture. I don't think we have polarized communities or interests here, not like we were in soil moisture with active vs passive microwave techniques and agriculture vs water resources applications, prior to developing our integrated soil moisture plan. But you are faced with the same kind of situation regarding our research approach. It should have the right balance of theoretical modeling and experiments. I would also recommend a strong initial emphasis on sensitivity analysis. The kind of questions we will encounter are as follows: "Ok, so you want to embark on this new program to use remote sensing for monitoring snowpack properties. Show us the model for this parameter and all the parameters that go in the model. Show us the sensitivity analysis of all these parameters, pick out the ones which are most sensitive. Now which of these can be addressed by remote sensing?" Now that is a pretty difficult order. I am not aware that every area of research that we get into is really analyzed in that kind of fashion. Certainly it is a very logical way of looking at the problem and provides NASA with insight regarding the impact of NASA research dollars. But that's the sort of thing I am confronted with at headquarters and I would like to see at the front end of any research plan.

Let me wind up by saying that the front end has to plead a strong case for the applications. What is needed? Why it is needed? And why it is important?

COLBECK: In that regard maybe I can say I chaired a working group for polar research for the National Research Council, and one of the things we did was to put together a lot of information about the practical implications of snow research, and we gathered everything from transportation to recreation, to agriculture. And when it came to getting some of the important facts about the implications of snowmelt runoff for a given area such as this, I found it fairly easy to get people to generate numbers. I remember Jack Washichek saying that they were going to Congress to get the money to justify the SNOTEL system. It was very easy to get the users to write many letters to Congress. These are the people who use the water, not the researchers.

CALABRESE: That's who we want to hear from. The people who are really going to use the information. We need to show NASA, OMB and the Congress that we're really putting out research dollars into something that is needed, will be used and that has a real payoff.

SHAFER: Are you saying then that what you would like to see is for the people that are here and have contact with other people to talk to some of the users and tell them what the potential might be and to have them write NASA letters of support? Or at least tell you how that might support their operation.

CALABRESE: Well I think you have done a pretty good job so far as I indicated previously. We have a whole series of letters in Washington supporting the program. I think you have made the case initially. I don't think we need to continue the campaign. I think it needs to be documented in the front end of this report along with the requirements.

COLBECK: The front end of this report says in Colorado so many million acre feet of snowmelt runoff are used, and the agriculture implications of that is so many dollars.

BROWN: It is so difficult to establish the dollar value of any improved snowmelt runoff forecast. I would go so far as to say we can't do it, unfortunately.

COLBECK: Can't you put a minimum figure on the value?

SHAFER: Wait a minute, I think it has already been done. That ASVT report that was done by ECOSystems. Number one, they looked at the entire Western United States where snowcover was important. They looked at the agricultural impact, hydropower etc. We know first of all how much water we are talking about in the Western United States. Also, there is a dollar figure for the given amounts of water and for specific water supply forecast points. There are some dollar figures attached to that water as improvements in forecast accuracy are made. So I think what we are talking about is if remote sensing is developed to the point where you could measure water equivalent or snow covered area, that report with a little bit of change or updating could be the vehicle that you are talking about.

CALABRESE: That's an input. I think that's a good start.

BOYNE: Would you say it's like that Wayne, on the implications of the prairie snowpack?

WILLIS: We have some of that type of information and, on the report that Sam alluded to, we might elaborate just a little bit. In the group that was involved, we put together a lot of numbers. Now those numbers are available and could be updated if necessary. But I want to shift gears just a little bit. The comments you made this morning Mike, I think were right on target. You mentioned one item that I would like to qualify, and that is that competition is healthy, which is true.

However, for the subject being discussed, I would hope that one of the bottom lines is that we view it more in terms of coordinated competition. This coordination would include coupling with the NASA soil moisture program. Snow is a source of water, water is needed for plant growth, and we are all interested in eating. These factors should be coupled, and at different levels. We are interested in separating the component parts so they can be understood and tied together. If we pursue, and I would highly recommend that if it is in order, that we have this group proceed and really do our best to package together a program which cuts across lines because I have a strong feeling that we are all in the same boat together or we had better be.

The utilization of snow is important. Hal mentioned the plains environment, and this is a good example. Bernie has talked about the SNOTEL system, and they've done a tremendous job with what they have had to work with, no question. Most of the water that comes down the creek from the mountain snowpack is used for irrigation or for municipalities. In the plains environment, you don't have that type collection system. But the data are irrefutable that snow is a resource that is not being used to best benefit. Snow can be managed. You manage it by managing the wind. So, through management techniques you can increase the soil water supply. Snow management reduces somewhat, the risks that are attendant to dryland agriculture. In the Missouri River basin, about 90 percent of the flow is derived from the mountains. That other 10 percent, if it is runoff and if it is all at one time can have a tremendous local impact. For example, Tom Carroll mentioned flooding on the Souris. With good snow management, such flooding can be reduced. The snowmelt water can be retained in place where it could be used for beneficial crop production. Also, the soil is protected as a resource base, and this is mandatory. A lot of these things do not sell, because who's hungry. I think we have a prime opportunity and I hope I detected a real ray of hope in what you said this morning, Mike, that we have a chance to work together on an important program.

CALABRESE: I think so.

WILLIS: This program can then cut across all sorts of lines. Many of these problems that we face are too complicated for any one individual or any one group to handle, and we'll be far better off if we do everything we can to do a better job of working together.

CALABRESE: I think that coordination with agriculture is important. Some of this is taking place in AgRISTARS in the conservation project. The great plains case needs to be written up.

WILLIS: I just used that as a comparison. About 10 percent of the land is irrigated. What do you do with the 90 percent that is dry land? Snow management goes into all areas, socioeconomics and clear across the board. We have the entire rangeland part that needs to be dealt with, as well as the cultivated land. I didn't mean to say that the Great Plains is one or the other; it is a part of the total. One of the important parts that I hope this group can address, is the tie between the snow group, including the physics of snow, the modeling, the whole bit, and the soil water group, and then the interfaces.

CALABRESE: I think that's a good point. There are a number of people who cross the related areas in snow and soil water. It is important to break them out as we have, and to focus them, otherwise we are looking at too big of a problem and we just don't seem to make headway. The key is to break it down to the right level of aggregation and the right community and then tie those together at the interfaces. I think we have the ingredients for doing that here. Al Rango is closely associated with the AgRISTARS soil moisture work along with Tom Schmugge.

RANGO: Let's get back to the first thing you are trying to get into Bernie; isn't there a fair amount of documentation on the importance of snow just recently generated by the SCS that could possibly be used to some degree?

SHAFER: I might give you a little background on that. We just went through this scenario recently and are still in the process. It turns out that somebody back in Washington decided to take a look at the snow survey for whatever reason. The impetus of that reason is that they did a cost benefit analysis of the program several years ago and found out that if the benefits of the snow measurement program applied only to agriculture were 20 to 1 and when you applied snow program information that was gathered to all applications it was like 40 to 1. There are very few government programs that approached those kinds of benefits to cost ratios. We thought that was a fantastic reason for us to continue with the program and thought it should be extended. But the rationale you can never understand is how the minds are going to work back there. They turned that around and said if it is that good then the Federal Government shouldn't be in it. So anyway we went through a public participation project just recently where people were asked to voice their opinion on how the snow survey should work and whether private industry, state government, county government, or whatever other entity, could take over and get the Federal Government out of it. There was a tremendous ground swell in just the opposite direction. Newspapers, TV and user groups began contacting their congressmen. We have literally thousands of letters from people who have attended public meetings. In Colorado alone, we have a binder 4 inches thick of letters from irrigation companies, federal agencies, farmers, individuals, municipalities, county governments, state governments, the whole smear. If you ever want to feel good about a program you should read those kinds of letters. There is a tremendous body of people out there who are interested in snow and can give you exact dollar figure on the impact snow has on their particular business enterprise or whatever they are involved with. And I think that kind of documentation would be useful to the kind of thing we are talking about and is available through the Department of Agriculture.

CALABRESE: I think the key is now, to relate snow benefits, snowpack properties, snow wetness and the contribution of remote sensing.

SHAFER: What it shows you more than anything else is who uses the information.

CALABRESE: That's good.

SHAFER: You can define user groups by whatever kinds of categories you want. You can divide up by private industry, agriculture, state government, whatever.

CALABRESE: That would be good. Cost benefit studies are nice to have but we are looking for strong user-need statements and documented applications.

COLBECK: You're saying beyond establishing that it is important, then you have to establish the reasons to do it better. Is that right?

CALABRESE: Yes.

BOYNE: In the Western U.S., water resources are finite and the user groups are increasing. We are going to have a big energy production operation over the next 50 years which is going to compete with the agricultural for water resources.

CALABRESE: Is that written up?

BOYNE: The Oil Shale and Environmental advisory panel has information on their water needs.

JONES: In fact if you read the Rocky Mountain News this morning, you would have seen an article questioning whether or not the West would have enough water to supply these projected new requirements in addition to the current demands. The point is that there is only a limited supply of water. There has been some interest in weather modification to increase the winter snowpacks in the mountains. Studies of such possibilities are currently being undertaken by various groups including the Water & Power Resources Service, USDI (formerly known as the Bureau of Reclamation). With this interest in cold orographic weather modification as a method of augmenting water supplies, it would seem that remote sensing of snowpack parameters would certainly be important on an operational basis as well as during the research phase.

Federal studies of weather modification are also being undertaken in California. Perhaps someone would want to speak to this. Weather modification may be one answer to the need for additional water. So here again, water is the key to economic survival of many of the western states, and as Bernie pointed out earlier, the supply is limited and the uses unlimited.

BOYNE: The population has been increasing for 40 to 50 years and municipal water needs are critical. The whole front range from Ft. Collins to Pueblo is going to grow and the projections of population and water needs are documented.

SHAFER: The key when you talk to people and political type people is everybody likes to use the word resource and now use the terms resource inventory. Well to some people resource inventory is if you do a land inventory go out and do it once, then forget about it. If you have a minerals inventory you have to do it every year and the inventory changes from the next year to the next year. This continual inventory and monitoring process is never really finite and just keeps going over and over again. To some people

the connotation of a resource inventory is a one shot deal. It's like the agricultural thing that you were talking about before. That has to occur each year and the importance of it goes up each year, as we're seeing here in the West.

RANGO: It sounds like there is enough material and information around that we shouldn't have a problem justifying the construction of such a research plan because it is a very important area. The other thing we might want to put in there is something on the measurement needs that the various users have, and I wonder if we know those things as well as we know how important it is.

WILLIS: There was some discussion this morning, Mike, on the Eastern versus the Western users. I don't have any problem at all justifying the selling of the Western, no question about it. The Eastern takes on a little different complexion. At the same time it seems to me that either is very important if you look at, for example, the heavy snow year. People were doing a lot of hollering, as I recall.

COLBECK: I think the fact that the independent, privately owned power companies do their own snow surveys indicates there is some level of interest in surveying. They're not spending their money for nothing, they're spending it because they need the forecast for optimum use of their hydroelectric facilities. It's not only the Corps of Engineers, Geological Survey and Soil Conservation Service doing these surveys.

WILLIS: Do we know enough about what they are doing?

COLBECK: I don't think anyone knows enough about what the many different groups in the East are doing. It's a tremendous mess.

WILLIS: The question, coming back to the point that Al raised, is that if we knew what they were doing and their needs, we could identify to them some possible improvements.

CALABRESE: It sounds like a Western problem but Sam is indicating that it's a National problem. I think all this could be articulated.

SHAFER: There is a report that has been done recently. Jack Washicheck spent about a year going back and talking to many of the user groups in the Eastern or Northeastern United States and the Northern Great Plains. He documented a lot of the users and the potential for improved snow surveying techniques. At that time the study was for the expansion of the SNOTEL system, but the same rationale that went into the study would be valid for satellite monitoring. It happened that it was directed at SNOTEL because we were concerned about that telemetering system. But the real need is for a broad surveying technique, and whatever mechanism you could use to do it. All I'm saying is this report is available and could be used to help document some of the needs in the Eastern US and upper mid-west.

BARNES: Serious consideration is being given in areas such as New England to redeveloping hydroelectric power dams that were abandoned years ago. If more hydroelectric power is developed in the coming years, it certainly seems as if the need will grow for more information on snowpack properties; at least a part of the information needed will have to come from satellite remote sensing techniques.

WILLIS: If it's not out of order, we could get supportive type backup from the Canadians since they have similar interests and similar conditions.

BOYNE: There should be no problem, Canada is part of the watershed.

COLBECK: I think we are focusing too much on trying to justify snow hydrology without trying to justify remote sensing. Yet it is clear that snow hydrology, snowmelt and runoff are important. It's much less clear what remote sensing can do to improve the situation.

WILLIS: I think that's easy. Taking a point at a time, I can see why we can't sell microwave remote sensing as our starting point. A prime point is that one of the big operational problems is to do surveillance of the snowpack. Anybody that has worked with snow in the field is in the business of taking point source information and extending it over an area-wide, basin-wide basis. Then, you are asking for trouble. We know that our resources are limited and there is no way we can go out and measure every point on every field sufficiently in a temporal sense to really get the answers we need, if you are only taking an operational view.

COLBECK: That's a sampling problem.

WILLIS: But, if we can obtain this sampling information, aggregated at different levels via the remote sensing techniques with a tremendous saving and a tremendous increase in the accuracy of your forecasting or whatever you're doing, then I see no problem in building a case as to why we need some means to measure the snowpack both in space and time. We cannot do it manually over extensive areas.

DOZIER: It seems to me that this section of the report is something that has to be very carefully done but it doesn't necessarily have to be very exhaustive. Nothing is going to turn off a hard-nosed bureaucrat more than a poorly done cost-benefit analysis or a long one. However, there are some, I know of at least one, pretty exhaustive reports by some individual water agencies, namely the Kern Water Agency, on just what benefits accrue on improvements in forecasting or improvements in the local allocation of water. I think we could dig up some comparable studies from different environments around the country. Then that really would be much more convincing an argument than speculative estimates of the benefit carried out by people sitting around the table. Part of the problem is that if you really want a thorough analysis of the benefits, this group is not equipped to do it. There are some, at least I know of one, pretty good estimates for a single water district.

COLBECK: The percentage improvement in snowmelt runoff in the Bonneville district is like 3 percent improvement in their operational capability through better forecast, 3 percent is small, but 3 percent times megabucks is large.

RANGO: I think the front end of the report then looks like it would be made up of the three topics.

The first would be to establish the importance of snow, which I don't think anybody has a problem with doing. There's plenty of information for that.

The second part would take a little more work although the data are probably there, too; and, that's the establishment of the snow user needs. We need to know the user requirements for measurements so we have some way to evaluate what we can do.

And then a third is, as Dr. Colbeck mentioned, the rationale of using remote sensing for snow measurement. Those are the elements following in logical sequence.

That's what I consider introductory material, but maybe now we could talk a little bit about the actual material in a report that we would be generating that would be new, i.e., what we should be concentrating on in a research program, develop emphasis for various parts of that program, and what should we recommend be done. Perhaps someone would care to review some of the comments that went on in the discussion earlier today that pertains to it or to bring up an approach we can use.

WILLIS: Is it advisable to categorize this?

RANGO: Categorize what?

WILLIS: Against user groups. More things that are needed by various users.

RANGO: I'm not sure, or is it easier or better to categorize it in terms of snow properties?

CALABRESE: Start with the former and work toward the latter, the needs of the various users.

RANGO: In the introduction this is established, but the question is how do we establish the rationale for remote sensing? We know there are problems that have't been solved. We can't solve all of them at the moment. How do we nail these things down is the question?

BOYNE: As a start, there are two consideration, one which has been talked about over the last few days, and the other not much at all. The first consideration is storage and runoff in mountain watersheds.

The second consideration is snow cover on the plains and prairies where snow management is important. As Tom Carroll was showing the other day, one

doesn't know the depth of snow or the melt condition of the snowpack, one knows only that the snowpack covers a wide area. What Tom Carroll is doing with gamma ray background radiation is to attempt to interpret radiation attenuation with snowpack water equivalence. This requires ground truth to determine snow depth, water equivalence and soil moisture. The latter is important because one cannot rely on soil moisture measurements made in October being the same as that in March or April when runoff is in progress. This snow management problem for flood control and soil moisture recharge could be addressed using remote sensing techniques. A scenario can be then made for the development of promising techniques into an operational system.

JONES: It seems to me that you are starting off with ideas very similar to those I have had. It is mainly the idea that there are two basic parameters that you need to know anywhere in the West or Great Plains. They are the areal extent of the snow and the water equivalent of the snowpack. If you had those two items of information, and only those two, you could evolve a very powerful management technique from that information.

DOZIER: It seems that we are dealing with sort of a hierarchy of levels of user needs. On the one hand, we can sort of put Bernie Shafer and Jean Brown as the part of our user group in charge of producing runoff estimates from data about snowpack. One of the ways we can serve them is ask, "What do you want to know about the snow?" And we can then think of remote sensing as a way to supply that information. Certainly that is extremely important and ought to be given a very high priority in additional research, that is, how can we do a better job in measuring the snow water equivalent and snow areal extent? In addition to that, we can take the view point that Bernie and Jean are not the user community but they're part of the same community that we're in. But there is another user community, namely the people that are going to use the runoff forecasts. We can say, there are some other types of information we can get from remote sensing that might be combined with different types of runoff models to give that user a somewhat better result. And it seems that those two things are really separate. I think Bruce's point is good. We have a first duty to accomodate the people who are producing the runoff forecasts. But we ought to think a little beyond that.

N. PETERSON: May I comment on that. We in the Department's Snow Survey Branch, as coordinators of the California Cooperative Snow Surveys Program, represent the users in California. They look to us for leadership. Also the agricultural users and other water managers look to us for direction, and so we do have a large segment of the users represented here.

RANGO: That brings us to the point of how we would improve snow measurements with remote sensing. And I might again just indicate that in the soil moisture approach the working group felt it necessary to break down the various stages of the research into a modeling section that talks not only about modeling but also how one would design the proper experiments to produce data to interpret as well as to verify models and improve them. And then the various levels at which experiments would be done - ground based, aircraft and satellite systems. Now at this meeting we've talked about

perhaps even finer details; that is, not only modeling but controlled laboratory experiments and then perhaps some field experiments that would spin off of those laboratory experiments. And we haven't talked much yet about aircraft or maybe helicopter extensions of that work and then finally some sort of satellite program. What kind of satellite and sensor would be best? There seems to be a lot of basic research to be done and we probably should review the techniques by which we collect data and try to compare it among the various investigators. So maybe some ideas on that sequence of events could be discussed. Is it something we use for this approach or do we want to take an alternative that someone might suggest.

SHAFER: Al, I kind of feel you're skipping a step. First you've identified that there's this big need for this data out there to study the snow. But an intermediate step, it seems to me, is that there should be some well defined goals out there, what the research is aimed at. Is it aimed at providing improved areal estimates of water equivalent? Of snowcover? Or is the snowpack wetness one of the primary goals? And beyond identifying those areas that research seeks to address, maybe identifying the level at which you would like to arrive at the end of 5 years is important. Maybe only in a research mode but with the potential for that application in an operational mode. In other words, what you describe seems to me you are jumping from the general field of snow into a description of where we currently are in the research mode, leaving out the goals and levels.

BOYNE: Once we define what the user needs are, we have to make a general statement of what possibilities there are with remote sensing applications to improve the situation and then break it down into detail. We have to get the specifics of what advantages remote sensing offers and then how and what steps need to be taken in order to achieve results.

WILLIS: I'd like to see some things included too, that would identify categories of work for research needs that might not necessarily be handled just by remote sensing. These would be identified needs to couple with remote sensing that would be some other agency responsibility. In other words, this gives a tie and a mechanism for creating an interaction that we've seen a lot more of recently, that 10 years ago we did't see.

CALABRESE: I think we're open on that as long as it is associated. NASA may not be in position to support this kind of development, however, another agency might and we'll tie it together.

WILLIS: To me, it would show a more complete picture. It shows precisely where NASA would have a very strong hand and at the same time show why there should be some interagency involvement to keep it tied together. I think this would help sell the program.

RANGO: Any comments from the observers that I can't see?

R. PETERSON: I would like to ask one thing. It's been said like in cost benefits, Jeff mentioned the study Kern Water Agency; Bernie mentioned Jack Washicheck's study has a lot of good information. Like the Kern Water Study, is it possible Jeff you could send that to us?

DOZIER: I hope so.

R. PETERSON: I'm looking for information that maybe we would use or wouldn't use but if we had information like that it would be helpful. Am I not correct Al, that we would be receptive to all of it?

RANGO: We'd like to get that information for use by this working group.

COLBECK: I have some information out of that report that we've used.

R. PETERSON: Al, could we generate an action item that anybody that has information that might be pertinent to send it to you.

RANGO: Yes, send it. Perhaps we can get that report that Jeff Dozier mentioned, the one Sam Colbeck worked on, the information Bernie Shafer has referred to, and have it all assembled.

ULABY: I have a comment to make. I see the process as consisting of two pieces. On the one hand, we have the users (hydrologic modelers) who try to specify the snow information of interest at, perhaps, two levels; the desirable level and the acceptable level. This information may consist of items like areal extent, water equivalent, wetness, spatial resolution, repeat cycle, etc. On the other hand, you have the sensor people who are trying to relate what the sensor observes to the parameters of interest. The process should have a continuous feedback-loop between these two groups. For example, the users may want a precision on water equivalent of two percent, the second group comes back and says, "We can give you only five percent, how acceptable is it? How does it impact the final utilization of the data in your models?" Hopefully, such a feedback system will lead to the generation of models that provide useful information and at the same time are compatible with the nature of the remotely sensed data.

DOZIER: Along that line would it help to get slightly more specific in a sense if you take models, not snowpack models but runoff models where you can get a lot. Is it going to help to find out what a satellite can do that nothing else can do? If saying you need something to 5 percent and something else to 10 percent not saying we can give it to you to 10 per- cent. If you can't give it to the accuracy that's wanted, why not forget it and concentrate on something you can give better than some other method. This would strengthen the use of remote sensing technology. This question of wetness, if you can get as much from air temperature you're not going to change the models you're using until they can give a lot better results than what you're using at present. So why waste time looking at those? Why not go to something you know you can get much better?

RANGO: The answer to that is correct in the optimum situation but there are complicating factors. Like Jean Brown mentioned he can't get in to large areas of the snowpack zone in California to do the conventional techniques. That's the best way, and he needs alternative methods.

The second aspect that was mentioned today was that there are many areas of the world that just don't have even basic data, and perhaps remote sensing

can't do it as well, but at least it can do it without actually having to go and collect that data in some remote region.

WILLIS: The only difference I see in what you're saying, I think I understand and tend to agree, would be the case where you know that you need something and because of technology or some other reason you can't get there yet. But, that does not necessarily say we should throw that away as a goal. In other words, we may want to, we can't do it now, but we want to remember that we want to get there when some new instrument or whatever it is allows us to get there.

SHAFER: Let me just ask a question associated with the soil moisture working group. Did they focus in on realizable objectives. Is that primarily what we're after? I mean we are not really pie in the sky at this point. They focused in on trying to solve some down to earth problems and I think that's what this working group should do, but not to the exclusion of what you said. I think there needs to be some mention of that but the real reason we're involved in this is to try to solve or help solve some identifiable problems and some realistic information needs.

WILLIS: Let me support Bernie by an example. In one of our research groups, we have felt that the farmer is one of our users. Now this immediately dictates that we have to get down to a field level basis. Even if we had all of the pieces in place right now with respect to remote sensing and so on that you think you like to have, we still have a problem with resolution in helping individual farmers on individual fields. And, again, we cannot physically measure each one. But that's not meaning to say that in another few years on the next satellite or something that we would not have that capability. In the meantime, we need to get all these other pieces in place so when that does arrive, we're in gear. So I support what you said.

SHAFER: It comes down to a priority thing. You decide when you can use the technology you have now for the most answers; the most good to the most people; and then, realizing that as you develop new technology, you are going to be able to help more specific groups. You're not excluding groups but you're saying, that this is our first priority; this is what we want to work on first. And this is what our second priority is and so on down the line. I would't think that what we would want to do is to focus on the most difficult of the problems, down to the individual in this case.

WILLIS: But it is a need, so I am using it as one example. We know some technology may be far away, but we should be smart enough to put together some of the pieces, in the framework of our current knowledge, to keep things packaged and on the right path.

RANGO: This will still operate in the framework that Dr. Ulaby has suggested that we have the modelers and users trying to identify various measurement requirements, the parameters that have to measured. We then could try to prioritize those and it would give the experimenters the direction on which they should be working on first and what kind of sequence they should be going through.

CALABRESE: Those like application models?   Physical process models?
Runoff models?

RANGO: It could be a variety of things.   I'm talking about models like
Dr. Kong's that just explain what is happening.   That's one kind of model.
The other kind that you might work with is a snowmelt runoff prediction
model.   That requires a certain type input.   We may focus then on some
parameters like water equivalent (a high priority user need), and perhaps
for better understanding we need to measure some properties of the snowpack,
grain structure or something; input to the models that would be a physical
understanding need.   And we would focus on those as our first topics and
then perhaps prioritize after that.   Does that sound like a workable thing
or are there other approaches we could use?

CALABRESE: It seems like we are in the same position here that we are
in soil moisture.   We don't have any models where it can have a direct input
but we're working on it.   In agriculture for yield, in water resources for
runoff.   And I guess it is the same thing here; if we gave you snow water
equivalent today, do you have anything we could input to directly?

SHAFER: Water equivalent, yes.   That's what we are telling you, we're in.

BROWN: Soil Moisture or snow wetness we would have to learn to use.

DOZIER: And snowcovered area you know how to use.

BROWN: We're just beginning to learn how to use snowcovered area (SCA).

CALABRESE: You're even a little further ahead than where we are in
some other fields where we're working around the parameters.

SHAFER: In our application we can tell you what we need.   We can
even tell you to some extent how accurate we need it over an area-wide
basis.   What we can't tell you is how accurate the instruments would have
to be meet this requirement.   The two parameters that we need are snow
water equivalent and snowcovered area.   But as soon as you throw in wet-
ness of the snowpack, given over a period of time, that becomes much less
clear to us how we would use it.   So what somebody would have to do is edu-
cate us in how we can use it.

DOZIER: Given that this is the case, it seems that the first step
would be to focus on those two things that the runoff predictors say they
would like to know better.   It seems that the next thing to do is to
identify some of the problems which adequately correct the collecting of
data by present day means and then when additional remote sensing capabili-
ties might overcome these.   It seems in snow water equivalent there are
problems in collecting it frequently enough, and with a good enough sampling
rate to give you a really reliable estimate.

Perhaps there's some way remote sensing can give you a measurement of snow
equivalent at a site which is as good as you can get by a conventional
manner, but is cheaper so you can get it at lots more sites.   Secondly the

problem in snowcovered area is similar. We can't get the data frequently enough. You have a problem measuring it under cloud cover. And we have a problem in measuring it in forest. Those are four fairly serious problems to really be able to use the system.

BROWN: Knowledge of snowcovered area alone without water equivalent is not meaningful. The perfect example was that in 1977 we had more snowcovered area than in 1978, yet 1977 was the driest year in over 50 years of record, and 1978 was better than 150 percent of normal.

DOZIER: Well I guess the way I like to look at it is from a volumetric standpoint. From existing snow water equivalent measurements, we are sort of trying an interpolation, to integrate snow water over the basin. And what the snowcovered area really tells us, is where our interpolation algorithm has to go to zero. They provide a nice boundary on the interpolation, that's really what is going on.

SHAFER: You think of snowpack in a basin as a wedge starting from zero where there is no snowcover up to a maximum near the basin head. What we would like to know is where the distribution is uniform from the lower part of the basin to the top of the basin. We know from our historical snow course records that's not the case. In some years, percentage wise, there's much more snow at the higher elevations, in some cases, more than the normal amount at mid-elevations or the lower elevations.

DOZIER: Snowcovered area doesn't necessarily solve that problem for you, but it does say where that wedge hits the ground and that's useful information. And then couple that with water equivalent, and you do know a lot more.

MÄTLZER: There are several possibilities from the microwave standpoint to determine the water equivalent, for example by looking at the daily variation of the brightness temperature. If the variation between freezing and melting occurs over a long time period, then it means that the snow depth must be large. If the snow pack changes rather quickly from completely dry to completely wet, then it means that the snow depth is rather small. That is an indirect method which we have described in our paper. Eventually our long term ground based active and passive experiments will lead to a better understanding of the relationship between the microwave signatures and the classical snow parameters, at least in a statistical sense.

COLBECK: I think there are areas in this country where much of the snowpack is wet throughout the winter. In parts of the Olympics and Cascades, for example, there may be some frozen snow near the surface, but there may be 6 meters of wet snow, seasonal wet snow, below that. I think that there are a wide range of conditions. In the Colorado Rockies the snowpack is mostly subfreezing throughout most of the winter but exactly the opposite is true in much of the Cascades and Olympics.

HOFER: But there is "a priori" knowledge. Which situation is predominate or most probable for a given area or time? Because of this knowledge remote sensing must not answer every question for you. Operational remote

sensing will most effectively be used for interpolations in space and time from calibration points or ground truth points and for the timely detection of anomalies.

DOZIER: In California we have European snow in some years, in some we have Colorado snow, in some you have Cascade snow.

SHAFER: Maybe what you're saying is that we don't put all our eggs in one basket, neither remote sensing or the snow pressure pillows. We'll probably continue to have both, only we'll get more information. One of the things I think might work in this situation, is during much of the winter when we have a lot of problems in the snowpack with ice crust, the pressure pillows sometimes don't work very well, but that's also the time that, at least in the Rocky Mountain Region, the snow is dry, and that's also the time the microwave works best. But in the spring, when the snowpack has a lot of liquid water in the snowpack, microwave doesn't work very well but in those cases the pillow seems to work quite well. At least they give you the daily melt. I think what you're talking about is trying to meld the two together so you don't say that remote sensing is going to provide all the answers for all the questions and automatically improve forecasting. What you're going to say is we'll get a little bit more information and that little bit more of information should improve things for forecast purposes.

BOYNE: The SNOTEL system allows one to accumulate data over a period of time. These are point measurements, however, and they have to be related, to areal runoff through a model. SNOTEL sites have been installed at snow-course sites in order to compare SNOTEL and snowcourse data using the same runoff model parameters. I think if we focus on areal measurement of water equivalence we could develop a 5 year research program.

Snowpack wetness is going to be a determining factor in the applicability of remote sensing techniques. We do not understand the electromagnetic responses in terms of snowpack wetness well enough to define snowpack wetness uniquely. We need to investigate the interaction to advance the state of the art.

WILLIS: Well, if you can't measure, why not? By asking "Why not", there may be some way the technique can be developed.

As part of the real objective: I've got a lot of faith in people to come up with ways of finding solutions to those questions.

Another question: To what extent should this be identified as interaction with AgRISTARS? Is there a real need for that?

RANGO: No, at this stage we don't care where the funding comes from. All we want to know and define is the ideal research program for snowpack properties and then later on we'll worry about where we obtain funding. It may come from AgRISTARS, it may come from water research money, or a variety of different sources. But I don't think at this stage we worry about either who does the research of where the money is obtained. We want to know what has to be done.

BOYNE: I thought he was asking a different point or I interpreted what he asked incorrectly.

RANGO: Are you talking about the tie with soil moisture?

BOYNE: The tie with soil moisture and with the hope that after five years we have developed something for AgRISTARS. For example, if we can determine that a crop is stressed, what other resources can be brought to bear to correct the situation? That ties in with water management generally.

RANGO: My answer was particularly directed toward funding. You're right, we have to look at the end application. I go back to one statement about fundamental understanding of these processes. I mean there are parameters that we would need to know to improve that fundamental understanding through modeling or whatever. They may not necessarily be water equivalent. They may be some of these model parameters referred to by Dr. Kong. Parameters such as the correlation length and the like. Should there be an element of this research plan that deals with those kinds of things?

SHAFER: There has to be or you can't get to these other questions. This whole dilemma about whether we deal with wetness - maybe we just dwelled on that too long. I mean there are two goals, water equivalent and areal extent. Before you ever get there you have to address all these other research questions and that's really what the whole research program is about. If you answer those questions, the goals are already met, it seems to me. You should't really focus a lot of time on what those problems are, and say, these are some of the problems that have to be addressed to meet these two goals and there is a need to meet the goals.

COLBECK: Goals and objectives in a management sense; the goals are long range.

WILLIS: You can also argue that depth of snow is important for agriculture.

COLBECK: The first priority should be water equivalent and areal coverage. Then of secondary importance there is wetness, depth, density, grain size and layering.

DOZIER: I think we have to address to some extent those basins where the model isn't going to be accurate enough and/or those basins where there is no historical record. So there is something we need to address, even though I'm prepared to concede in this case that the large majority of basins are probably modeled adequately by present methods, but could use better data. I think, I'm sure Sam would agree, even maybe Bernie would agree, that there are some basins where those methods are not going to work. And there ought to be some attention paid to those.

JONES: Lets back up for just a second. Wayne was talking about snow depth as a goal. Where are we now? The ice depth concept is pretty well understood. What about snow depth? Can that be done with similiar techniques?

RANGO: I think you have the same problem with snow depth that you do with water equivalent. To me they're inseparable. An example of where you'd use one as opposed to the other, was in a satellite study where we found that the published water equivalent data was just sort of rounded off, whereas, the snow depth data was recorded more accurately. So that's why we used snow depth. I don't think that you can separate them in general although, for winter wheat kill and other similar applications, snow depth seems to be the important parameter because of its insulating effect. So we'd want to consider that in particular applications areas.

BOYNE: I think what Jean was trying to say is if we had snow depth we could get water equivalent indirectly, and if we had water equivalent, we could get snow depth indirectly. We can predict what the average density of the pack is going to be in the various geographical areas.

DOZIER: An added thought here. Is it possible for HCMM data to be used to directly measure density during periods before you get any melt to occur. There's a density term in the thermal inertia, there's a thermal conductivity term which is sort of an empirical function of density. Isn't that product the square root? Or is there something else in there?

BARNES: That's basically it. You can measure density from thermal inertia. We are currently conducting an investigation to determine whether HCMM thermal data can provide additional information on snowmelt. The question of using these data to determine snowpack density would certainly be worth looking at.

BOYNE: We're still back to the point where we are talking about areal extent and water equivalent.

BROWN: I agree completely with what Bernie has been saying. He has said it very well and there isn't much left to say. But I would like to add one other comment that goes beyond what we deal with today. In the future I visualize the need for some kind of parameter which would relate energy coming into the snowpack as a means of determing how fast or how well that pack is going to melt. The parameter could be albedo. Right now we are using temperature as kind of a crutch to get energy input. Albedo, or a similar parameter could be the third thing coming up in the future in addition to SCA and water equivalent that could perhaps very well lend itself to remote sensing, especially if we could determine that the parameter would relate to the amount of energy coming into the snowpack. With that taking place, then I can visualize us moving very rapidly into more sophisticated model techniques.

BOYNE: Surface wetness might be a way to go.

RANGO: I think we are sort of conceptualizing a framework here that's pretty good. But there was a topic brought up before about consistency of ground measurements to aid in comparison of the results of different experiments. Is this something we should deal with soon in the program?

BOYNE: Oh yes, I think so.

COLBECK: These should be objectives. The goals are still water equivalent and areal extent.

WILLIS: So Al's question is if we are going to have better understanding of the physical process, then this relates to your point, Hal, about having some standardizations.

BOYNE: We need standardization of ground truth techniques. Most ground truth characterization tends to be incomplete. One needs a handbook of standard techniques and standard nomenclature for reporting results. One may need to develop techniques to provide supplemental information. Techniques for measuring density, temperature profile, stratigraphy, grain size and wetness exist and standard procedures for documenting these observations have been agreed upon internationally (UNESCO, WMO, 1970). Additional ground truth measurements may be needed to characterize the electromagnetic interaction with the pack, such as correlation length, etc.

There has not been enough ground truth and/or systematic reporting of ground truth to intercompare results meaningfully.

ULABY: This can be accomplished by generating a manual that defines standard techniques and measurement procedures for ground-truth sampling and spatial sampling frequency. This type of approach was adopted by the soil moisture working group, which also recognized the need for the development of new, improved techniques for ground-truth measurements.

WILLIS: So generally there would be a minimum set of data and you have ancillary data. And then on your question of time, Hal, conceivably if this were to happen right away, then we are going into the summer period on this side of the equator. By knowing these needs before the next fall we might gear up some things so we begin to get some of these standard data sets, to have things in place before the next winter season.

JONES: I believe that attempting to standardize snowpack ground-truth procedures would be a very good objective. Conceivably we could prepare a draft copy of a "handbook" yet this fall and distribute it for comments. Then, during the winter we could actually take people into the field and train them in the handbook procedures. Based on this experience the handbook would then be revised and hopefully by this time next year a workable handbook would be available. This should help standardize snowpack ground-truth procedures.

At the same time we should also address the problem of when the ground truth is to be taken for specific time window requirements. In other words, if you have to prioritize the data taken in the field, which data are the most time dependent and which are the most valuable? Hopefully these questions could also be addressed in the handbook.

BROWN: During the research and development phase of developing satellite information we need as much ground truth as we can get, and standardization of that ground truth is unquestionable. But I want to emphasize to you that eventually when satellite information becomes operational I will want very little ground truth. I want to be able to get the satellite information

with such reliability that we will need very little ground truth, especially in the areas of the snow zone that are designated wilderness.

WILLIS: Is this an objective?

STILES: There are other problems in real life. We have to categorize the importance and timeliness for each piece of ground truth. That comes first, for you never have a million ground truthers and it's ridiculous to take density profiles 30 times a day. But wetness profiles, if you could get them, would be great 30 times a day.

JONES: This is what I was getting at. You will have to look at the timeliness of each piece of data as well as its value.

STILES: You have to also look at what is realistic as far as what you can measure, and the order of importance of these things.

RANGO: Well do we have any other aspects of this that we should be considering here.

COLBECK: Yes, I have a question that's been on my mind about ICEX. If I understood it, they were talking about a 10 cm altimeter is that correct? I think that's what Dr. Zwally said a year ago; a 10 cm altimeter. If there are 10 cm altimeters in snowcover studies, does that offer anything.

RANGO: First of all it may be 3 cm.

ULABY: It's a reference. It doesn't give you the absolute distance. It gives you the elevation relative to a reference.

DOZIER: Even if it is relative, we know what the elevation of the underlying terrain is.

COLBECK: If we had two of them, one which saw the underlying terrain and the other which saw the top of the snow.

CALABRESE: Can we register that?

HOFER: All altimeters work with respect to the mass distribution of the earth which is not completely known. For the SEASAT altimeter an rms-accuracy of less then 10 cm was found. The measurement is made with respect to the geoid which determines the satellite path. The measurement is averaged over the effective reflection footprint. For measuring sea state or significant wave height for example there is a wave model needed to describe the height distribution in a statistical sense. The height distribution is retrieved by a pulse form analysis. For snow the reflection mechanism and the penetration depth is dependent on the snow parameters. Additionally, snow is often found in areas with (compared to the footprint) rapidly varying and complicated height distributions.

BARNES: I think that it is important to make every attempt to integate a microwave research program, such as a scientific program to compare microwave

measurements and ground truth data, with other types of data that may be available from satellites launched for other purposes. There was some discussion this morning that before we need to worry about looking from space in the microwave, we need to understand better the physical properties of the snowpack. Certainly every effort must be made to make use of all types of spacecraft data that will be available in the coming years, and that may give us other information on snowpacks.

COLBECK: Suppose you could determine the thickness of the snow and the density of the top 10 cm rather than the whole pack. Suppose you guys could look only at the top 10 cm, wouldn't that be an easier problem than looking at 2 meters?

ULABY: In order for that to work you would have to be looking at glass because we cannot get spatial resolution and depth resolution simultaneously. You have to do one or the other.

BOYNE: Even if the satellite were geostationary one could measure the distance to the earth with a precision of only one part in 10^7, even with the best available clocks. To get a difference measurement between snowpack and topography requires subtracting a very small distance from two rather large numbers.

STILES: Your registration would be impossible. You're in the mountains and fly just one foot off and your elevation difference could be 50 feet.

RANGO: I would propose that we now have a rough framework for a plan that we can use as a starting point. The next step is to have a meeting of the working group where we would not just talk about how to go about developing a plan, but rather to start putting these things down on paper to generate the plan. Such a meeting could be within a few months pending discussions with headquarters. And what we'd like to do there that we haven't done in soil moisture, is give you a little more preparation on this thing. We'll try to come up with a prepared framework, and then in addition, we'll try to identify areas in which we would like you to contribute, rather than have you all come in and try to recall all the references and so forth off the top of your head. You'll be able to do a little preparatory work, and maybe we'd get more done in the time that we're setting around the table and see if we can't come up with a first cut at such a plan. At that meeting we'll discuss how we would break down into sub working groups to try to generate this thing. And at the end of that meeting we'd have some results that we could give to a contractor to put into a plan format for us that we could then review as a working group. And eventually, within a year say, we would come up with a snowpack properties research plan that we could submit to Headquarters that would document the importance of snow and list the citical research tasks that have to be done. So if that's agreeable to everyone, we'll try to work on the framework and decide on some likely times and locations for the next meeting. Do any of you have further suggestions?

BOYNE: Some of those documents that people are going to send in, I hope you get those circulated.

RANGO: That's correct. We'll try to give you that reference material before the next meeting.

CALABRESE: Would it be useful to send these people a copy of the Soil Moisture Plan?

RANGO: I think so. I'll distribute it in the final form so you can use it as a frame of reference. It gives you an idea of the scope of things we're talking about.

BARNES: Also, I believe the ASVT Snow Project was a good project to show where we are now in terms of the application of areal snow extent determined from satellites. People who may not be familiar with that project may appreciate getting copies of the ASVT final reports.

RANGO: The proceedings are supposedly published by now. They just haven't reached my office yet. The proceedings of that workshop will be available in a few days probably, and these will be sent out as well as the background information. So we'll have a lot of information to get to you before the meeting so you can have some background and then we'll be ready to go. So any final comments on where we stand? This meeting is then adjourned.

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15. Supplementary Notes

16. Abstract

The proceedings of a workshop on Microwave Remote Sensing of Snowpack Properties, held in Fort Collins, Colorado, on May 20-22, 1980, are reported in this NASA Conference Proceedings document. Although visible and infrared data can be used very effectively for measuring snow cover extent in clear weather (see NASA CP-2116), an all-weather remote sensing capability should be developed to provide reliable snow cover data through clouds. Secondly, a remote sensing capability should be developed for the basin-wide measurement of snow water equivalent. The combination of snow area and water equivalent yields the snow water volume stored on the watershed. Such additional capabilities will most likely require an expanded development of remote sensing techniques in the microwave region of the electromagnetic spectrum. Much microwave snow research has been performed, but an organized exchange of research results, technical discussions, and interactions between conventional snow hydrologists and microwave scientists has been lacking. Fourteen scientific papers were presented at this Workshop over a three day period interspersed with numerous discussion sessions. The final discussion at the Workshop on May 22 focussed on the initial development of a microwave snow research plan to assist NASA in formulating their continuing research program. The 14 scientific papers and the transcription of the final discussion session are published in this document.

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