

Proceedings of the Fifth Atmospheric Radiation Measurement (ARM) Science Team Meeting

**March 19-23, 1995
San Diego, California**

April 1996



**United States Department of Energy
Office of Energy Research
Office of Health and Environmental Research
Environmental Sciences Division**



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MASTER

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History and Status of the Atmospheric Radiation Measurement Program - March 1995

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This document contains the summaries of papers presented at the 1995 Atmospheric Radiation Measurement (ARM) Science Team meeting held in San Diego, California. To put these papers in context, it is useful to consider the history and status of the ARM Program at the time of the meeting.

The history of the project has several themes. First, the Program has from its very beginning attempted to respond to most critical scientific issues facing the United States Global Change Research Program. Second, the Program has been strongly coupled to other agency and international programs. Indeed, the Program reflects an unprecedented collaboration among various elements of the federal research community, among the U.S. Department of Energy's (DOE) national laboratories, and between an agency's research program and the related international programs, such as Global Energy and Water Experiment (GEWEX) and TOGA. Next, ARM has always attempted to make the most judicious use of its resources by collaborating and leveraging existing assets and has managed to maintain an aggressive schedule despite budgets that have been much smaller than planned. Finally, the Program has attracted some of the very best scientific talent in the climate research community and has, as a result, been productive scientifically. This introduction covers the first three points—the papers themselves speak to the last point.

Initial Concept

The initial concept for ARM came out of a series of studies that fell under the auspices of the Intercomparison of Radiation Codes in Climate Models (ICRCCM). ICRCCM pointed to several key issues that are now central to the ARM approach and strategy. First, ICRCCM was based on an assertion that one must understand the quality of the physics inside a climate model if one is to understand the quality of the climate model itself. Next, it showed that it is

possible, and in fact necessary, to understand the relatively coarse representations of physics contained in a climate model in terms of a hierarchy of process models. For radiation, this hierarchy ranges from the highly detailed line-by-line codes to the highly parameterized forms of the radiation codes used in climate models. Finally, the hierarchy of models that leads to the parameterizations of processes in climate models must be built on a sound base of experimental verification.

Concurrently with the release of the ICRCCM results, it was becoming clear that the radiative transfer of energy in the atmosphere and the impact of clouds was, and remains, one of the greatest sources of error and uncertainty in the current generation of general circulation models (GCM) used for climate research and prediction. With this as a starting point, DOE proposed a major program targeted at improving the understanding of the role and representation of atmospheric radiative processes and clouds in models of the earth's climate. Initially, the DOE Program focused on the radiative aspects of the climate problem. As the scientific issue was studied in more detail, however, it was obvious that a study of radiative processes associated with clouds could not be decoupled from the problem of representing the processes by which clouds form, are maintained, and dissipate in climate models. As a result, the ARM Program was proposed to the then Committee on Earth Sciences of the Federal Coordinating Council on Science Engineering and Technology with two basic objectives:

- to improve the treatment of radiative transfer in climate models under all relevant conditions
- to improve the treatment of clouds in climate models, including the representation of the cloud life cycle and the prognosis of cloud radiative properties.

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The "Approved" Plan

The ARM Program Plan was subjected to peer review in the fall of 1989. The key element of the proposed ARM effort was to be the Cloud and Radiation Testbed (CART). This user facility was to consist of four to six semi-permanent observational facilities designed to allow detailed investigations of process models used in climate research. These more permanent facilities were to be supplemented with a mobile facility that would allow related measurements to be made at other locations on a campaign-oriented basis. The facility would include a data management and communications system capable of acquiring and quality-controlling site data; acquiring data from sources outside the program; and communicating that data to a Science Team. This Science Team would be selected through a peer review process open to all investigators nationally and internationally.

Based on the peer review, the subcommittee on Global Change Research of the Committee on Earth Sciences approved the Plan, noting several key things about how it should be carried out. First, the scope was broadened beyond radiative transfer to include clouds and cloud processes, a change deemed necessary to adequately address those atmospheric properties important to radiative transfer in the atmosphere and the atmospheres's radiation balance. Next, the Committee recommended that the DOE implementation of this program involve the talents of other federal agencies to the extent possible and that an interagency steering group be formed to assist in that process. Finally, the relevance of ARM to several other climate programs was noted, and DOE agreed to tie its deployment of facilities to the schedules of other national and international programs.

The Early Implementation

The implementation of ARM began in January 1990, proceeding on two coupled but parallel tracks. First, a multi-laboratory team was formed to plan the detailed implementation of the ARM facilities. The second track involved the formation of the Science Team. Because the science drivers were important to the design of the ARM facilities, a series of scientific workshops were held in the spring and summer of 1990 to clarify the scientific foundations of the program. In parallel, a solicitation process was initiated to establish the Science Team.

As these two tracks moved forward, features of the Program emerged. One of the most significant was a pattern of collaboration with other programs. This collaboration was characterized on one hand by a series of joint field campaigns and, on the other, by involvement in program

planning. In the field collaborations, ARM attempted to bring a value-added contribution to another agency's or group's planned effort, while at the same time trying to gain operational experience necessary to guide its own field deployment.

This strategy resulted in collaborations with the Federal Aviation Administration's Winter Icing and Storms Program (WISP) and First ISCCP Regional Experiment (FIRE) activities in Coffeyville, Kansas, and the Azores. In Coffeyville, early ARM concepts were tested in the jointly funded NASA-DOE Spectral Radiance Experiment (SPECTRE). It also led to ARM-fostered projects such as the Boardman-ARM Regional Flux Experiment which tested key aspects of surface and surface flux characterization.

From the standpoint of planning, ARM attempted to gain early involvement in the program planning of other programs that would be evolving in parallel with it. Most notable among these planning collaborations was the GEWEX. One of these joint planning activities culminated in the field deployment of the Pilot Radiation Observation Experiment (PROBE) to Kavieng, Papua New Guinea, as part of TOGA-COARE, in the winter of 1992-3. Again, experience gained during TOGA-COARE has been a crucial influence in ARM planning.

A key convergence between science and facility planning tracks was the selection of a siting strategy for the ARM facilities. This process resulted in the identification of five locales in which ARM should locate its semi-permanent facilities, and a comparable number of secondary locales in which the program should consider shorter, campaign-like activities. The primary locales in the order of their intended occupation were the Southern Great Plains of the United States, the Tropical Western Pacific, the North Slope of Alaska, marine stratus zones of either the Atlantic or Pacific Ocean, and the Gulf Stream.

Budget Realities

While ARM was planned as a decade-long program with a cumulative funding level of almost \$500M, it has always been clear that the annual rate of expenditure would not reach projected levels and that the Program's schedule would be drawn out. This reality has been approached in several ways and needs to be understood in terms of several competing concerns: the cost of acquiring equipment, the tradeoff between capital and operating budgets, and the costs of facility design and deployment versus operating costs.

Early in the Program, capital equipment resources were inadequate to acquire the instrumentation necessary for the first site and the development of the associated data system. As a result, the deployment to the first site was phased, supporting one aspect of the program, the radiative transfer segment, over the cloud life cycle segment. Similarly, the Program sought opportunities to take advantage of existing equipment and data. This approach led, in no small way, to the decision to deploy the first site in the North Central Oklahoma/South Central Kansas area, to take advantage of the existing National Oceanic and Atmospheric Administration (NOAA) profiler and radar facilities and the then-developing Oklahoma Mesonet.

The operational budgets here also lagged, leading to a series of joint development activities. For example, rather than building a new data system for field data acquisition, the Program instead developed a collaboration with the National Center for Atmospheric Research (NCAR), to build the data system around their campaign data management system, now known as Zebra.

Finally, the project has been rescoped annually. This rescoping has resulted in substantive changes including the cancellation of the planned mobile facility, the reduction of planned permanent field sites from five to three, slowed deployment and development of instruments and facilities, sharply fewer than anticipated campaign activities, and delays in the implementation of the ARM Data Archive as a facility readily accessible to the wider scientific community.

Despite budgetary limitations, development of the central facility for the southern Great Plains site began in May 1992, only one month later than originally planned. The initial deployment was meager, a single portable meteorological station borrowed from the National Center for Atmospheric Research. By that fall, however, most of the infrastructure for the instrumentation was in place, and the major equipment was being delivered. Originally planned for completion in about one year, some aspects of the Southern Great Plains facility development are continuing in FY 1995.

In other areas, the initial deployment to the second permanent locale, the Tropical Western Pacific, has been delayed to 1996, and deployment to the third permanent locale, the North Slope of Alaska, continues to be planned for 1997. This schedule is reflective of the impact of the limited budgets allocated to the program. The originally planned deployment schedule called for one site to be completed each year, implying a full deployment of five sites by mid-1997.

Project Status at the Time of the San Diego Meeting

During the period between the Science Team Meetings at Charleston and San Diego, the major changes in the project involved additional instrumentation and facilities at the Southern Great Plains, the firm commitment to deploy the first instrumentation in the Tropical Western Pacific to the island of Manus, Papua New Guinea, and the initiation of planning activity to deploy to the North Slope of Alaska in 1997. Substantial links to external data sources were established and data were routinely being provided through the data system from U.S. Satellites, and the National Weather Service, the Oklahoma Mesonet land, the Kansas agricultural network, for example.

At the Southern Great Plains site, the central facility was largely complete by the end of 1994, but the extended facilities were still only about 50% complete by March of 1995. Work was in progress to establish the remaining extended facilities with much of the instrumentation being in hand. A tentative decision was in place to establish a set of three "intermediate" sites about halfway between the central facility and the boundary facilities. These sites would be comprised of 915-MHz boundary layer profilers with radio acoustic sounding systems to support research objectives in the planetary boundary layer and the routine characterization of site wide boundary layer fluxes. The 915-MHz systems would be deployed in 1996. In addition to routine site operations, intensive operational periods were becoming a regular event at the site, and the list below indicates the IOPs conducted up to the time of the San Diego meeting:

Date	Intensive Operational Period
11/92, 6/93	Investigation of 915-MHz Profilers for PBL Measurements for 4DDA
4/93	IDP Field Evaluation for AERI
6/93	GPS Water Vapor Msmt Evaluation
1/94, 4/94, 7/94, 10/94	Single Column Model IOP

The April, 1994 IOP period was noteworthy as the most complex IOP conducted to date. During April, three separate measurement programs were conducted in parallel, with outstanding success. 1) Sonde operations supported a single-column model data acquisition period; 2) a variety of instruments developed under the Instrument Development Program were deployed to the site for field

testing and participation in a cloud remote sensing IOP along with operational instruments, and 3) the ARM UAV program conducted a series of seven demonstration flights over the site providing confirmation that UAVs are viable data acquisition vehicles and revealed that their low noise environment provides for better data than that acquired from other airborne platforms. The IDP instruments deployed included two Raman lidars, a research grade polarization lidar for cloud microphysics, and three cloud radars, operating at either 35 or 95 Ghz. The cloud remote sensing IOP was also supported by the University of North Dakota Citation, made available through the grant of one of the investigators. While the UAV used for the demonstration flights had a limited altitude capability, the operations proved the value of the platform for scientific data acquisition and paved the way for future field programs.

At the 1993 American Meteorological Society annual meeting, ARM investigators revealed that measurements of satellite, airborne and groundbased radiometric measurements were in disagreement with model predictions by about 40 watts per square meter of solar energy not reaching the ground. During 1994 this became the basis for planning a major campaign to be conducted in 1995 to acquire the data necessary to understand why the measurements and models are in disagreement. The proposed campaign came to be called the ARM Enhanced Shortwave Experiment, or ARESE, and came to revolve around a series of UAV flights using a high altitude vehicle. It is anticipated that the 1996 ARM Science Team Meeting should see some of the data analysis from this program.

By March 1995, the TWP siting strategy that had been proposed at the Charleston meeting led to visits to potential sites in the Tropical Western Pacific locale. As anticipated, Manus Island in Papua New Guinea was confirmed as the best choice as the site for the first observational capability to be deployed. Nauru was selected as the second site from several island locations to the east of Manus; land availability, the island's infrastructure, accessibility, and its relatively "open" ocean location were all elements influencing the decision. Initial discussions with the governments in both locations were encouraging, but left much to be done in 1995, prior to deployment. In parallel, the design of the Atmospheric Radiation and Cloud Station (ARCS) was being refined in accordance with the siting and observational strategy and integration of the facility was beginning at Sandia National Laboratory, Albuquerque. Deployment of the first ARCS to Manus was anticipated for late fall, 1995. The primary path for data to be returned to the ARM Experiment Center remained as periodic shipping of tapes of recorded data. "Health of Station" data would be transmitted periodically through a GOES relay and would be

used for diagnosis and status determination of the various systems comprising the ARCS.

For the North Slope of Alaska, a siting and strategy document was in preparation encompassing several deployment scenarios. The first was an option to participate in the proposed SHEBA experiment in the Spring of 1997 and then to deploy the first permanent observing capability to a location near Barrow, Alaska later in that same year. Several variations from this basic theme were also being discussed based largely on the firmness of plans for SHEBA. The instrumentation used to participate in SHEBA would, at the completion of the SHEBA ice camp deployment, become the core of the initial observational capability at an inland site, probably near the village of Atkasuk, 50 km south of Barrow.

Collaboration with other programs continued to be a hallmark of ARM. Between the Charleston and San Diego meetings, collaboration between GEWEX and ARM continued. During the summer of 1994, ARM sonde operations supported GEWEX and were, in part, supported by GEWEX in order to provide data for periods of time beyond the normal ARM IOP periods. A proposal from the University of Oklahoma to the NOAA program office for the GEWEX Continental Scale International Project (GCIP), was submitted and accepted to provide support for the acquisition and installation of soil water and temperature (SWAT) measurement systems to be installed at each of the ARM extended facilities. The systems were planned for acquisition in FY 1995 and installation at ARM extended facilities in 1996. ARM will operate these instruments as if these were ARM instruments. Discussions were initiated to modify three ARM extended facilities to make them better able to meet the observational needs of the ecological community. Funding for this activity was initially identified, but did not materialize. Plans remain in place to modify ARM extended facilities as necessary should the support to do so come available.

ARM continued to plan towards participation in the Marine Continent Thunderstorm Experiment (MCTEX) as part of the initial deployment to the TWP and the Surface Heat Budget of the Arctic (SHEBA) as part of the initial deployment to the NSA. ARM and EOS continued to plan towards a joint Science Plan pending completion of respective program Science Plans.

The existence of the Southern Great Plains site and its data was attractive to NOAA's Verification of the Origins of Rotation in Tornadoes Experiment (VORTEX) research program. In collaboration with ARM, VORTEX planned for use of a NOAA WP-3 for tornadic research in the vicinity of

the ARM site and provided a number of flight hours for ARM-related research data to support objectives of several ARM Science Team members from NOAA, NASA and the University of Oklahoma. In addition to the normal complement of instrumentation, NOAA agreed to mount a Gerber liquid water probe acquired by ARM. This program was planned for the May 1994 time frame overlapping with the regular ARM IOP period.

Against this broad range of activity, the Science Team began to deal with real data from the SGP site and began to furnish valuable feed back to the instrument and data experts within the infrastructure to improve the quality of the data flow. From a research standpoint, the Science Team began to focus on specific objectives yielding many of the papers presented as posters at the Science Team Meeting and in the extended abstracts presented in this volume. Many of these efforts fall into two of the general measurement strategies that have been extensively discussed in the course of the evolution of ARM; these are the use of single-column models and a focus on the instantaneous radiative flux through the atmosphere; these are predominantly reflected in the IOPs being conducted on a regular basis. A third general area of activity is closely aligned with establishing the lower boundary of the single column model, but is still an area of activity important to several programs interested in interacting with ARM and is sufficiently distinct to warrant being recognized - that is the area of planetary boundary layer behavior and its role in controlling lower boundary heat and moisture fluxes into the single-column model. Approaches previously discussed using hierarchical diagnosis and data assimilation are coming to be more widely recognized as tools or approaches to be applied to single-

column model or instantaneous radiative flux research questions. Drawing from earlier descriptions, these are defined as follows:

- single-column model - Many of the key process models and parameterizations that compose GCM-based climate models appear to be testable by extracting a single vertical array of cells from the model and operating the array in what is referred to as a single-column model. This single-column model retains the subgrid scale physics that must be represented in GCMs, and it offers a promising approach to testing the parameterization of this physics. The CART facility around Lamont, Oklahoma, was designed to support this strategy.
- instantaneous radiative flux - Accurate treatment of radiation is essential in climate models, and testing of radiation transfer models is central to the objectives of ARM. In principle, testing of radiation transfer models calls for complete specification of the state of the atmosphere and the surface. The ICRCCM (Ellingson and Fouquart 1991) has emphasized that the state-of-the-art in radiative modeling cannot be advanced by further model intercomparisons. Progress can be achieved only by supplementing intercomparisons with field observations. CART provides a facility for such observations.

Reference

Ellingson, R., and Y. Fouquart. 1991. The intercomparison of radiation codes in climate models: an overview. *J. Geophys. Res.* 96:8925-8927.

Abbreviations and Acronyms

ADT	anomalous diffraction theory
AERI	atmospheric emitted radiance interferometer
AERI-X	Atmospheric Emitted Radiance Interferometer-Extended Resolution
AGL	above ground level
ARCS	Atmospheric Radiation and Cloud Station
ARESE	ARM Enhanced Shortwave Experiment
ARM	Atmospheric Radiation Measurement
ASRC	Atmospheric Sciences Research Center
ASTEX	Atlantic Stratocumulus Transition Experiment
ASTI	Absolute Solar Transmission Interferometer
AVHRR	advanced very high resolution radiometer
BARFEX	Boardman ARM Region Flux Experiment
BBSSs	Balloon-Borne Sounding Systems
BLP	boundary layer profiler
BNL	Brookhaven National Laboratory
BORCAL	broadband outdoor radiometer calibration
BTAs	back trajectory analyses
CART	Cloud and Radiation Testbed
CCD	charge-coupled device
CCM2	Version 2 of the NCAR Community Climate Model
CCN	cloud condensation nucleus
CIMMS	Cooperative Institute of Mesoscale Meteorological Studies
CIRA	Cooperative Institute for Research in the Atmosphere
CMDL	Climate Monitoring and Diagnostics Laboratory
CN	condensation nuclei
COAMPS	Coupled Ocean/Atmosphere Mesoscale Prediction System
COARE	Coupled Ocean Atmosphere Response Experiment
COR	Coriolis
CPRS	Cloud Profiling Radar System
CRF	cloud radiative forcing
CRM	cloud resolving model
CSU	Colorado State University
DAR	Division of Atmospheric Research
DIAL	infrared differential absorption lidar
DISORT	discrete ordinate radiative transfer
DMSP	Defense Meteorological Satellite Program
DOE	U.S. Department of Energy
DVN	daytime versus nighttime
DWR	dual wavelength ratio
EBBR	Energy Balance Bowen Ratio
EBTs	Equivalent Blackbody Temperatures
ECMWF	European Centre for Medium Range Weather Forecasting
EM	explicit microphysics
EML	Environmental Measurements Laboratory
ENSO	El Nino-Southern Oscillation
ER	equivalent radius

ERBE	Earth Radiation Budget Experiment
ETL	Environmental Technology Laboratory
FASE	FASCODE for the Environment
FCC	fractional cloud cover
FDDA	four-dimensional data assimilation
FDI	field data ingestor
FFT	fast Fourier transform
FIRE	First ISCCP Regional Experiment
FIRE-II	Second ISCCP Regional Experiment
FOV	field-of-view
GCM	general circulation model
GCSS	Global Energy and Water Cycle Experiment Cloud Systems Study
GISS	Goddard Institute for Space Studies
GMS	geostationary meteorological satellite
GOES	Geostationary Operational Environmental Satellite
GSFC	Goddard Space Flight Center
HAcc	horizontal advective acceleration
HAD	horizontal diffusion
HDiv	horizontal divergence
HIS	High-resolution Interferometer Sounder
ICET	Integrated Cumulus Ensemble and Turbulence
ICRCCM	intercomparison of radiative codes in climate models
IMC	ice mass content
IN	ice nuclei
IOP	intensive observation period
IPA	independent pixel approximation
IR	infrared
IRF	Instantaneous Radiative Flux
ISCCP	International Satellite Cloud Climatology Project
ITCZ	intertropical convergence zone
JFD	joint frequency distribution
KF	Kain-Fristch (deep convection model)
LBLRTM	Line-by-Line Radiative Transfer Model
LCL	lifting condensation levels
LDRD	Laboratory Directed Research and Development
LDRs	linear depolarization ratios
LES	large-eddy simulation
LFC	level of free convection
LIAR	lidar/radiometer
LLJ	low-level jet
LST	local solar time
LW	longwave
LWC	liquid water content
LWP	liquid water path
MAPS	Mesoscale Analysis and Prediction System

MAS	MODIS Airborne Simulator
MBL	marine boundary layer
MC	Monte Carlo (simulation)
MCS	mesoscale convection system
MFRSR	multifilter rotating shadowband radiometer
ML	mixed layer
MLS	mid-latitude summer
MPL	Micro Pulse Lidar
MPL	Marine Physical Laboratory (Sun and Thorne article only, p. 331)
MSL	mean sea level
MWR	microwave water radiometer
NASA	National Aeronautics and Space Administration
NCAR	National Center for Atmospheric Research
NDVI	normalized-difference vegetative index
NIGEC	National Institute for Global and Environmental Change
NOAA	National Ocean and Atmospheric Administration
NREL	National Renewable Energy Laboratory
NWS	National Weather Service
OLR	outgoing longwave radiation
OLS	Operation Line Scanner
OML	ocean mixed layer
PBL	planetary boundary layer
PCMDI	Program for Climate Model Diagnosis and Intercomparison
PDF	probability distribution function
PDL	Polarization Diversity Lidar
PGF	pressure gradient force
PIR	precision infrared radiometer
PMS	particle measuring system
PNNL	Pacific Northwest National Laboratory
PRF	pulse repetition frequency
PROBE	Pilot Radiation Observation Experiment
PRT	precision radiation thermometer
PW	precipitable water
QME	Quality Measurement Experiment
RAMS	regional atmospheric modeling system
RASSs	Radio Acoustic Sounding Systems
RCM	Regional Circulation Model
RCMs	radar coded messages
RCS	Remote Cloud Sensing
RH	relative humidity
RH _c	critical relative humidity
R _n	net radiometer
RRTM	rapid radiative transfer model
RSS	rotating shadowband spectroradiometer
RT	radiation transport
RTNEPH	Real-Time Nephelometer
RW	required warming
RWPs	Radar Wind Profilers

Abbreviations and Acronyms

SAGE	Stratospheric Aerosol and Gas Experiment
SCMs	single column models
SEPM	stochastic entraining parcel model
SGP	Southern Great Plains
SIROS	solar and infrared observing system
SNR	signal-to-noise ratio
SORTI	Solar Radiance Transmission Interferometer
SPECTRE	Spectral Radiance Experiment
SRL	Scanning Raman Lidar
SRRB	Surface Radiation Research Branch of NOAA
SSM/T1	Special Sensor Microwave Temperature Sounder
SSM/I	Special Sensor Microwave Imager
SSM/T2	Special Sensor Microwave Water Vapor Sounder
SST	sea surface temperature
SW	shortwave
TAdv	temperature advection
TAO	Tropical Atmospheric Ocean
TEM	trajectory ensemble model
TKE	turbulent kinetic energy
TOA	top of the atmosphere
TOGA	Tropical Ocean Global Atmosphere
TOVS	TIROS Operational Vertical Sounder
TP	tropospheric profiler
TRES	Tomsk Radiation Experiment in Siberia
TWP	Tropical Western Pacific
USSA	U.S. Standard Atmosphere
UT	Universal Time
UW	University of Wisconsin
VAD	Velocity-Azimuth Display
VDU	vertical diffusion
WPDA	Wind Profiler Demonstration Array
WSI	Whole Sky Imager
WWCB	Weekly Weather and Crop Bulletin (Department of Agriculture Publication)

Contents

History and Status of the ARM Program - March 1995	iii
Abbreviations and Acronyms	ix
Session Papers	
A Boundary-Layer Cloud Study Using Southern Great Plains Cloud and Radiation Testbed (CART) Data <i>B. Albrecht, G. Mace, X. Dong, W. Syrett, and T. Ackerman</i>	1
Explicit Simulation of a Midlatitude Mesoscale Convection System <i>G.D. Alexander and W.R. Cotton</i>	5
A Comparison of Ground and Satellite Observations of Cloud Cover to Saturation Pressure Differences During a Cold Air Outbreak <i>R. J. Alliss and S. Raman</i>	9
Radiative Heating in Global Climate Models <i>F. Baer, N. Arsky, and K. Rocque</i>	13
Evidence for Island Effects and Diurnal Signals in Satellite Images of Clouds Over the Tropical Western Pacific <i>S.A. Barr-Kumarakulasinghe, R.M. Reynolds, and P.J. Minnett</i>	19
Analytical Study of the Effects of the Low-Level Jet on Moisture Convergence and Vertical Motion Fields at the Southern Great Plains Cloud and Radiation Testbed Site <i>X. Bian, S. Zhong, C.D. Whiteman, and S.A. Stage</i>	23
Initial Analyses of Surface Spectral Radiance Between Observations and Line-by-Line Calculations <i>P.D. Brown, S.A. Clough, N.E. Miller, T.R. Shippert, D.D. Turner, R.O. Knuteson, H.E. Revercomb, and W.L. Smith</i>	29
Cloud Classification Using Whole-Sky Imager Data <i>K.A. Buch, Jr., C.-H. Sun, and L.R. Thorne</i>	33
Broken-Cloud Enhancement of Solar Radiation Absorption <i>R.N. Byrne, R.C.J. Somerville, and B. Subasilar</i>	39
Toward Improving the Representation of Anvil Cirrus Effects in General Circulation Models (GCMs) <i>H.-N.S. Chin and M.M. Bradley</i>	43
The Role of Aerosols in Cloud Drop Parameterizations and Its Application in Global Climate Models <i>C.C. Chuang and J.E. Penner</i>	49
Performance of Radar Wind Profilers, Radiosondes, and Surface Flux Stations at the Southern Great Plains (SGP) Cloud and Radiation Testbed (CART) Site <i>R.L. Coulter, B.M. Lesht, M.L. Wesely, D.R. Cook, D.J. Holdridge, and T.J. Martin</i>	53

Using Remotely Sensed Planetary Boundary Layer Variables as Estimates of Aerally Averaged Heat Flux <i>R.L. Coulter, T.J. Martin, and D.J. Holdridge</i>	59
Modeling the Summertime Arctic Cloudy Boundary Layer <i>J.A. Curry, J.O. Pinto, and K.L. McInnes</i>	63
Observed and Simulated Temperature Dependence of the Liquid Water Path of Low Clouds <i>A.D. Del Genio and A.B. Wolf</i>	69
Heterogeneous Surface Fluxes and Their Effects on the Southern Great Plains (SGP) Cloud and Radiation Testbed (CART) Site <i>J.C. Doran, Q. Hu, J.M. Hubbe, J.C. Liljegren, W.J. Shaw, S. Zhong, and G.J. Collatz</i>	73
Data Assimilation of a Ten-Day Period During June 1993 Over the Southern Great Plains Site Using a Nested Mesoscale Model <i>J. Dudhia and Y.-R. Guo</i>	77
Comparison of Mean Properties of Simulated Convection in a Cloud-Resolving Model with Those Produced by Cumulus Parameterization <i>J. Dudhia and D.B. Parsons</i>	83
Integrated Cumulus Ensemble and Turbulence (ICET): An Integrated Parameterization System for General Circulation Models (GCMs) <i>J.L. Evans, W.M. Frank, and G.S. Young</i>	89
Drizzle Production in Stratocumulus <i>G. Feingold, A.S. Frisch, B. Stevens, and W.R. Cotton</i>	93
Diagnostics from a 1-D Atmospheric Column <i>J.M. Flately and G. Mace</i>	97
Island Based Radar Microwave Radiometer Measurements of Stratus Cloud Parameters During the Atlantic Stratocumulus Transition Experiment (ASTEX) <i>A.S. Frisch, C.W. Fairall, J.B. Snider, D.H. Lenschow, and S.D. Mayer</i>	101
A Study of Monte Carlo Radiative Transfer Through Fractal Clouds <i>C. Gautier, D. Lavallec, W. O'Hirok, P. Ricchiazzi, and S.R. Young</i>	105
Simulation of the Low-Level Jet by General Circulation Models <i>S.J. Ghan</i>	111
Prediction of Cloud Droplet Number in a General Circulation Model <i>S.J. Ghan and L.R. Leung</i>	113
Evaluation of a Stratiform Cloud Parameterization for General Circulation Models <i>S.J. Ghan, L.R. Leung, J. McCaa, and X. Zou</i>	117
Implementation of Raman Lidar for Profiling of Atmospheric Water Vapor and Aerosols at the Southern Great Plains (SGP) Cloud and Radiation Testbed (CART) Site <i>J.E.M. Goldsmith, F.H. Blair, and S.E. Bisson</i>	121

Variability of Radiatively Forced Diurnal Cycle of Intense Convection in the Tropical West Pacific <i>W.M. Gray, J.D. Sheaffer, and W.B. Thorson</i>	125
Feasibility of Tropospheric Water Vapor Profiling Using Infrared Heterodyne Differential Absorption Lidar <i>C.J. Grund, R.M. Hardesty, and B.J. Rye</i>	129
Retrieving Moisture Profiles from Precipitable Water Measurements Using a Variational Data Assimilation Approach <i>Y.-R. Guo, X. Zou, and Y.-H. Kuo</i>	133
Coupled Interactions of Organized Deep Convection Over the Tropical Western Pacific <i>X. Hong and S. Raman</i>	139
Radiative Transfer in Atmosphere-Sea Ice-Ocean System <i>Z. Jin, K. Stamnes, W.F. Weeks, and S.-C. Tsay</i>	143
Evaluating Aerosol Indirect Effect Through Marine Stratocumulus Clouds <i>Z.N. Kogan, Y.L. Kogan, and D.K. Lilly</i>	147
A General Circulation Model (GCM) Parameterization of Pinatubo Aerosols <i>A.A. Lacis, B.E. Carlson, and M.I. Mishchenko</i>	151
Aerosol Measurements at 60 m During the April 1994 Remote Cloud Study Intensive Operating Period (RCS/IOP) <i>R. Leifer, B. Albert, H.N. Lee, R.H. Knuth, S.F. Guggenheim, and L. Kromidas</i>	155
Aerosol Measurements at the Southern Great Plains Site: Design and Surface Installation <i>R. Leifer, R.H. Knuth, S.F. Guggenheim, and B. Albert</i>	159
Comparison of Satellite-Derived and Observer-Based Determinations of Cloud Cover Amount at the Southern Great Plains Cloud and Radiation Testbed Site <i>Y.P. Liaw, D.R. Cook, D.L. Sisterson, and W. Gao</i>	163
Numerical Simulations of Altocumulus with a Cloud Resolving Model <i>S. Liu and S.K. Krueger</i>	167
A Simple Formula for Determining Globally Clear Skies <i>C.N. Long, A.T. George, G.G. Mace, and T.P. Ackerman</i>	173
Diffusion and Scattering in Multifractal Clouds <i>S. Lovejoy, B. Watson, D. Schertzer, S. Pecknold, P. Silas, and G. Brosamlen</i>	177
Examination of Objective Analysis Precision Using Wind Profiler and Radiosonde Network Data <i>G.G. Mace and T.P. Ackerman</i>	183

A Comparison of Radiometric Fluxes Influenced by Parameterized Cirrus Clouds with Observed Fluxes at the Southern Great Plains (SGP) Cloud and Radiation Testbed (CART) Site <i>G.G. Mace, T.P. Ackerman, and A.T. George</i>	185
A One-Year Cloud Climatology Using Data from the Southern Great Plains (SGP) Site Micropulse Lidar <i>G.G. Mace, T.P. Ackerman, J. Spinhirne, and S. Scott</i>	187
Comparisons of Cloud Ice Mass Content Retrieved from the Radar-Infrared Radiometer Method with Aircraft Data During the Second International Satellite Cloud Climatology Project Regional Experiment (FIRE-II) <i>S.Y. Matrosov, A.J. Heymsfield, R.A. Kropfli, and J.B. Snider</i>	189
Raman Lidar Measurements of Water Vapor and Aerosols During the Atmospheric Radiation Measurement (ARM) Remote Cloud Sensing (RCS) Intensive Observation Period (IOP) <i>S.H. Melfi, D.O'C. Starr, D. Whiteman, R. Ellingson, R.A. Ferrare, and K.D. Evans</i>	193
Technical Progress Report: Completion of Spectral Rotating Shadowband Radiometers and Analysis of Atmospheric Radiation Measurement Spectral Short-Wave Data <i>J. Michalsky and L. Harrison</i>	199
Rotating Shadowband Radiometer Development and Analysis of Spectral Shortwave Data <i>J. Michalsky, L. Harrison, Q. Min, J. Schlemmer, W. Berkheiser III, and C. Zhou</i>	205
Effect of Particle Nonsphericity on Bidirectional Reflectance of Cirrus Clouds <i>M.I. Mishchenko, W.B. Rossow, A. Macke, and A.A. Lacis</i>	211
Radiative Properties of Ice Clouds <i>D.L. Mitchell, D. Koracin, and E. Carter</i>	215
RRTM: A Rapid Radiative Transfer Model <i>E.J. Mlawer, S.J. Traubman, and S.A. Clough</i>	219
Instrument Development for Atmospheric Radiation Measurement (ARM): Status of the Atmospheric Emitted Radiance Interferometer-Extended Resolution (AERI-X), the Solar Radiance Transmission Interferometer (SORTI), and the Absolute Solar Transmission Interferometer (ASTI) <i>F. Murcray, T. Stephan, and J. Kusters</i>	225
Fractional Integration and Radiative Transfer in a Multifractal Atmosphere <i>C. Naud, D. Schertzer, and S. Lovejoy</i>	227
Direct Shortwave Forcing of Climate by Anthropogenic Sulfate Aerosol: Sensitivity to Particle Size, Composition, and Relative Humidity <i>S. Nemesure, R. Wagener, S.E. Schwartz</i>	231
Development and Testing of an Aerosol-Stratus Cloud Parameterization Scheme for Middle and High Latitudes <i>P.Q. Olsson, M.P. Meyers, S. Kreidenweis, and W.R. Cotton</i>	235

Airborne Studies of Submicron Aerosol in the Troposphere Over West Siberia <i>M.V. Panchenko, V.E. Zuev, B.D. Belan, S.A. Terpugova, V.V. Pol'kin, and A.G. Tumakov</i>	241
The Optical Properties of Equatorial Cirrus in the Pilot Radiation Observation Experiment <i>C.M.R. Platt, S.A. Young, P. Manson, G.R. Patterson, and S. Marsden</i>	245
Atmospheric Emitted Radiance Interferometer (AERI): Status and the Aerosol Explanation for Extra Window Region Emissions <i>H.E. Revercomb, R.O. Knuteson, F.A. Best, T.P. Dirkx, R.G. Dedecker, R. Garcia, B.A. Whitney, W.L. Smith, and H.B. Howell</i>	253
The Lidar Dark Band: An Oddity of the Radar Bright Band Analogy <i>K. Sassen, R.E. McIntosh, and S.M. Sekelsky</i>	259
Interaction of Clouds, Radiation, and the Tropical Warm Pool Sea Surface Temperatures <i>N. Schneider, G.J. Zhang, T.P. Barnett, V. Ramanathan, U. Lohmann, and E. Roeckner</i>	263
Evaluation of Cloud Prediction and Determination of Critical Relative Humidity for a Mesoscale Numerical Weather Prediction Model <i>N.L. Seaman, Z. Guo, and T.P. Ackerman</i>	273
A Shallow Convection Parameterization for the Non-Hydrostatic MM5 Mesoscale Model <i>N.L. Seaman, J.S. Kain, and A. Deng</i>	277
Dual-Wavelength Millimeter-Wave Radar Measurements of Cirrus Clouds <i>S.M. Sekelsky, J.M. Firda, and R.E. McIntosh</i>	279
Remote Cloud Sensing Intensive Observation Period (RCS-IOP) Millimeter-Wave Radar Calibration and Data Intercomparison <i>S.M. Sekelsky, J.M. Firda, R.E. McIntosh, E.E. Clothiaux, G.G. Mace, and T.P. Ackerman</i>	283
Balloon-Borne Radiometer Profiler: Field Observations <i>W.J. Shaw, C.D. Whiteman, G.A. Anderson, J.M. Alzheimer, J.M. Hubbe, and K.A. Scott</i>	287
Satellite Data Sets for the Atmospheric Radiation Measurement (ARM) Program <i>L. Shi and R.L. Bernstein</i>	291
FASCODE for the Environment (FASE) <i>H.E. Snell, J.-L. Moncet, G.P. Anderson, J.H. Chetwynd, S. Miller, and J. Wang</i>	293
Cirrus and Aerosol Lidar Profiler - Analysis and Results <i>J.D. Spinhirne, V.S. Scott, J.A. Reagan, and A. Galbraith</i>	297
Aspects of the Quality of Data from the Southern Great Plains (SGP) Cloud and Radiation Testbed (CART) Site Broadband Radiation Sensors <i>M.E. Splitt and M.L. Wesely</i>	301
The Multi-Filter Rotating Shadowband Radiometer (MFRSR) - Precision Infrared Radiometer (PIR) Platform in Fairbanks: Scientific Objectives <i>K. Stamnes and E. Leontieva</i>	307

Observations of a Cold Front With Strong Vertical Undulations During the ARM RCS-IOP <i>D. O'C. Starr, D.N. Whiteman, S.H. Melfi, A.R. Lare, R.A. Ferrare, B. Demoz, K.D. Evans, G. Mace, K. Sassen, S.E. Bisson, and J.E.M. Goldsmith</i>	311
Summertime Low-Level Jets Over the Great Plains <i>D.J. Stensrud and S. Pfeifer</i>	317
Microphysical Structure of Simulated Marine Stratocumulus: Effects of Physical and Numerical Approximations <i>B. Stevens, W.R. Cotton, and G. Feingold</i>	321
Size Distributions of Boundary-Layer Clouds <i>R. Stull, L. Berg, H. Modzelewski, and K. Schrieber</i>	327
Inferring Spatial Cloud Statistics from Limited Field-of-View, Zenith Observations <i>C.-H. Sun and L.R. Thorne</i>	331
Longwave Scattering Effects on Fluxes in Broken Cloud Fields <i>E.E. Takara and R.G. Ellingson</i>	335
Comparison of Seasonal and Zonal Patterns of the Direct and Indirect Radiative Forcing of Climate by Aerosols <i>R. Wagener and S.E. Schwartz</i>	341
Treatment of Cloud Radiative Effects in General Circulation Models <i>W.-C. Wang, M.P. Dudek, X.-Z. Liang, M. Ding, L. Zhu, E. Joseph, and S. Cox</i>	343
Clouds, Radiation, and the Diurnal Cycle of Sea Surface Temperature in the Tropical Western Pacific <i>P.J. Webster, C.A. Clayson, and J.A. Curry</i>	349
Observations of Regional and Local Variability in the Optical Properties of Maritime Clouds <i>A.B. White and C.W. Fairall</i>	355
Solar Semidurnal Tidal Wind Oscillations Above the CART Site <i>C.D. Whiteman and X. Bian</i>	359
A Characteristic Scale in Radiation Fields of Fractal Clouds <i>W. Wiscombe, R. Cahalan, A. Davis, and A. Marshak</i>	363
The Great Plains Low-Level Jet (LLJ) During the Atmospheric Radiation Measurement (ARM) Intensive Observation Period (IOP) - 4 Simulations of Land Use Pattern Effect on the LLJ <i>Y. Wu and S. Raman</i>	367
Clouds and Snowmelt on the North Slope of Alaska <i>T. Zhang, K. Stamnes, and S.-A. Bowling</i>	373
A Case Study of the Great Plains Low-Level Jet Using Wind Profiler Network Data and a High Resolution Mesoscale Model <i>S. Zhong, J.D. Fast, X. Bian, S. Stage, and C.D. Whiteman</i>	377

Some Results of an Experimental Study of the Atmospheric Aerosol in Tomsk: A Combined Approach <i>V.V. Zuev</i>	381
Albedo and Transmittance of Inhomogeneous Stratus Clouds <i>V.E. Zuev, E. I. Kasyanov, G. A. Titov, and S.M. Prigarin</i>	389
Simulation of Solar Radiative Transfer in Cumulus Clouds <i>V.E. Zuev and G.A. Titov</i>	393
Absorption of Solar Radiation in Broken Clouds <i>V.E. Zuev, G.A. Titov, and T.B. Zhuravleva</i>	397
Author Index	401

A Boundary-Layer Cloud Study Using Southern Great Plains Cloud and Radiation Testbed (CART) Data

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Introduction

Boundary layer clouds—stratus and fairweather cumulus—are closely coupled to the water and energy budgets of land surfaces. This coupling involves the radiative impact of the clouds on the surface energy budget and the strong dependence of cloud formation and maintenance on the turbulent fluxes of heat and moisture in the boundary layer. The continuous data collection at the Southern Great Plains (SGP) Cloud and Radiation Testbed (CART) site provides a unique opportunity to study components of the coupling processes associated with boundary layer clouds and to provide descriptions of cloud and boundary layer structure that can be used to test parameterizations used in climate models. But before the CART data can be used for process studies and parameterization testing, it is necessary to evaluate and validate data and to develop techniques for effectively combining the data to provide meaningful descriptions of cloud and boundary layer characteristics. In this study we use measurements made during an intensive observing period (IOP) conducted at the SGP CART site in April 1994 for data validation and technique development. During this period we consider a case where low-level stratus were observed at the site for about 18 hours. This case is being used to examine the temporal evolution of cloud base, cloud top, cloud liquid water content, surface radiative fluxes, and boundary layer structure. A method for inferring cloud microphysics from these parameters is currently being evaluated.

Results

The clouds observed during this study were associated with a weak cold front that passed over the site. During this period the flow was dominated by a large high pressure center that moved from the Great Lakes at 1200 UTC 22 April to the Carolina coast by 1200 UTC 24 April. At about 0800 UTC (0200 LST) on 22 April 1994 a shallow layer of cool, moist air moved southwest over the CART site. This front was associated with a wind shift and a 3 °C decrease in temperature. Shallow stratus clouds were first observed at

the time of the frontal passage and persisted for another 18 hours. The initial analysis presented here focuses on a four-hour period near the beginning of the event and a five-hour period near the end.

Although the energetics of the clouds that were observed in the event may be more strongly controlled by the surface heat fluxes than marine boundary layer clouds, the mean thermodynamic structure is well characterized by a mixed layer structure. The boundary layer deepened slightly during the day and the potential temperature jump across the inversion increased from about 2 K during the initial period to about 6 K during the second observing period. Since the above-inversion air was relatively moist, during the entire event equivalent potential temperature increased with height across the inversion. Future work will consider the detailed evolution of the boundary layer structure using wind profiler and radio-acoustic sounding system measurements.

The cloud characteristics for the event were obtained from the CART (Belfort) ceilometer, microwave radiometer, narrow band infrared (IR) (Heimann) radiometer and the Penn State 94-GHz radar. For this initial analysis the 94-GHz reflectivity data were analyzed subjectively to estimate cloud base and top at 10-minute intervals. These estimates are shown in Figure 1 and indicate a cloud thickness of 200-400 meters during the first analysis period with clouds thinning to about 200 meters during the second period. The cloud-top heights agree well with the inversion heights obtained from the soundings. The cloud-base heights from the ceilometer (Figure 2) are correlated with the radar cloud-base heights, but are systematically about 100 m higher than the radar estimates. As the cloud thins during the second analysis period, the ceilometer cloud bases are about 200 meters higher than the radar cloud bases and are occasionally higher than the cloud top. The cloud-top estimates, however, may only be accurate to about 75 meters. Furthermore, estimates of cloud-base heights from radar reflectivity are also uncertain. But as the cloud continues to thin after 2400 UTC (not shown here), the ceilometer no longer detects a cloud even though returns are still obtained from the radar.

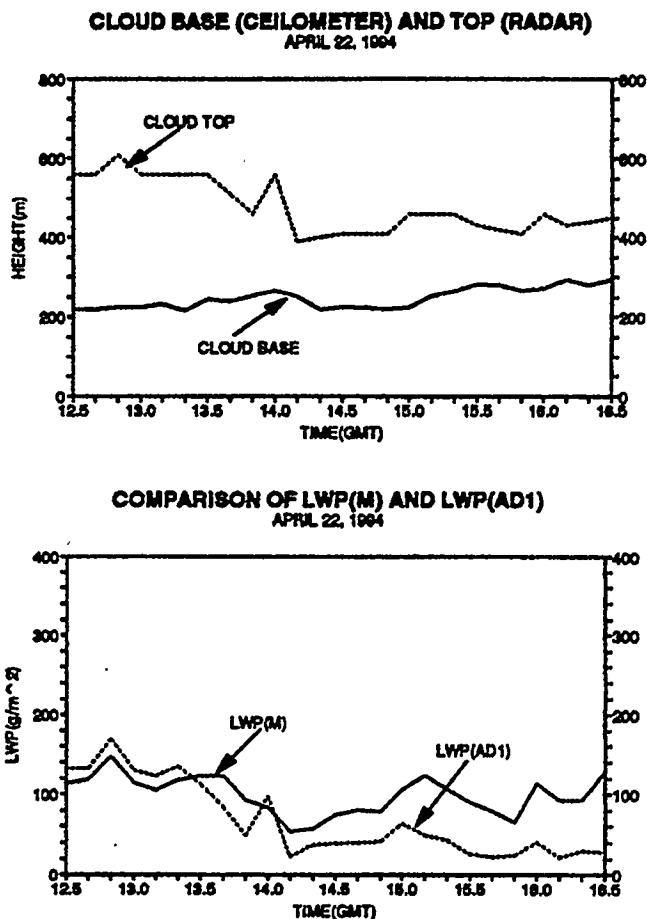
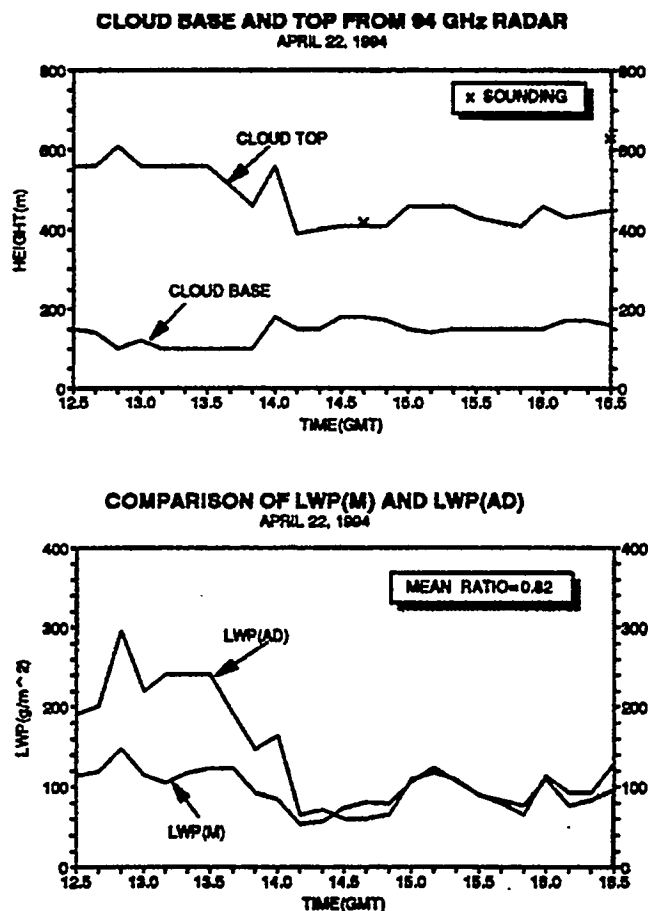


Figure 1. Cloud base and cloud top from 94-GHz reflectivities. Liquid water path (LWP) from microwave radiometer. Adiabatic LWP calculated using cloud-top and -base heights.

Figure 2. Cloud base from CART ceilometer. Liquid water path (LWP) from microwave radiometer. Adiabatic LWP calculated using cloud-top and -base heights.

The cloud thicknesses from the radar and the ceilometer were used to calculate the adiabatic liquid water path. A comparison of the calculated liquid path with liquid water path retrieved from the microwave radiometer is also shown in Figures 1 and 2. A positive clear-sky bias of 46gm^{-3} was removed from the microwave data. The adiabatic values obtained using the cloud radar to estimate cloud base show reasonable correlation with the measured values. For the thicker cloud observed during the beginning of the first analysis period, the observed water path is substantially less than the adiabatic values. From 1410 UTC to 1630 UTC the observed values are very close to adiabatic. During the second analysis period it is also found that the deeper clouds have liquid water paths that are less than the adiabatic value, but more closely match the adiabatic value when the cloud thins to about 200 meters. The adiabatic liquid water path

calculated using the radar cloud tops and ceilometer cloud bases is almost always less than the observed liquid water path. These results further indicate that the ceilometer may underestimate cloud-base heights for these relatively thin clouds. The causes for this discrepancy are being explored. Although the clouds observed during this period are relatively thin, they have a substantial impact on the downward solar radiation observed at the surface. Although the downward shortwave flux shows considerable variability, the IR emission remains relatively constant as indicated by the IR radiometer cloud-base temperatures. Work is in progress to retrieve cloud microphysical properties by using the cloud boundaries, liquid water path, and surface solar flux in conjunction with a two-stream radiative transfer model.

Conclusions

Data collected at the SGP CART site during an IOP in April 1994 were used to study low-level stratus clouds and their effects on the surface radiation budget. Although the analysis focuses on a one-day case associated with persistent stratus observed during the April 1994 IOP, the emphasis is on the synergism of the diverse data sets collected at the site and the development of strategies for routinely characterizing cloud, boundary layer, and radiation conditions from the CART data for low-cloud situations.

Although the clouds during this period were relatively thin, estimates of cloud-base height, cloud-top height, and liquid water content provide a consistent description of the evolution of the clouds and one that may be very useful for model verification and process studies. Work is in progress to couple the surface radiative fluxes to the observed cloud

properties. Although this case study provides a prototype for how the CART analyses can be effectively used to produce cloud and radiation characterizations useful for testing climate model parameterizations, additional effort is needed to derive these cloud properties routinely from the CART data. Furthermore, the data from the CART site should be carefully evaluated to determine any systematic biases in ceilometer cloud-base heights and microwave radiometer estimates of liquid water path. The utility of using a millimeter radar for the definition of the properties of boundary layer clouds is clearly documented. Work is in progress to refine objective techniques for defining cloud-top heights. Routine observations from millimeter wavelength radars would be invaluable for studying boundary layer clouds and identifying multi-layered cloud systems that might complicate the interpretation of bulk measures of cloud properties such as liquid water path and surface solar fluxes.

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Explicit Simulation of a Midlatitude Mesoscale Convective System

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Introduction

We have explicitly simulated the mesoscale convective system (MCS) observed on 23-24 June 1985 during PRE-STORM, the Preliminary Regional Experiment for the Stormscale Operational and Research and Meteorology Program (Cunning 1986). Stensrud and Maddox (1988), Johnson and Bartels (1992), and Bernstein and Johnson (1994) are among the researchers who have investigated various aspects of this MCS event. We have performed this MCS simulation (and a similar one of a tropical MCS; Alexander and Cotton 1994) in the spirit of the Global Energy and Water Cycle Experiment Cloud Systems Study (GCSS), in which cloud-resolving models are used to assist in the formulation and testing of cloud parameterization schemes for larger-scale models (Browning et al. 1993).

In this paper, we describe 1) the nature of our 23-24 June MCS simulation and 2) our efforts to date in using our explicit MCS simulations to assist in the development of a GCM parameterization for mesoscale flow branches. The paper is organized as follows. First, we discuss the synoptic situation surrounding the 23-24 June PRE-STORM MCS followed by a discussion of the model setup and results of our simulation. We then discuss the use of our MCS simulations in developing a GCM parameterization for mesoscale flow branches and summarize our results.

Synoptic Situation and Observations

The PRE-STORM 23-24 June MCS formed under classic synoptic conditions. At 1200 UTC 23 June a cold front trailed from a deep surface low near Hudson Bay, with the front becoming stationary across the central United States. By 0000 UTC 24 June the Nebraska surface low had deepened a bit with a dryline extending south from it into western Kansas. A stationary front snaked its way south-eastward into the low and east-northeastward out of the low (see Figure 1 of Stensrud and Maddox 1988). At this time, low-level air south of the front was hot and moist, with

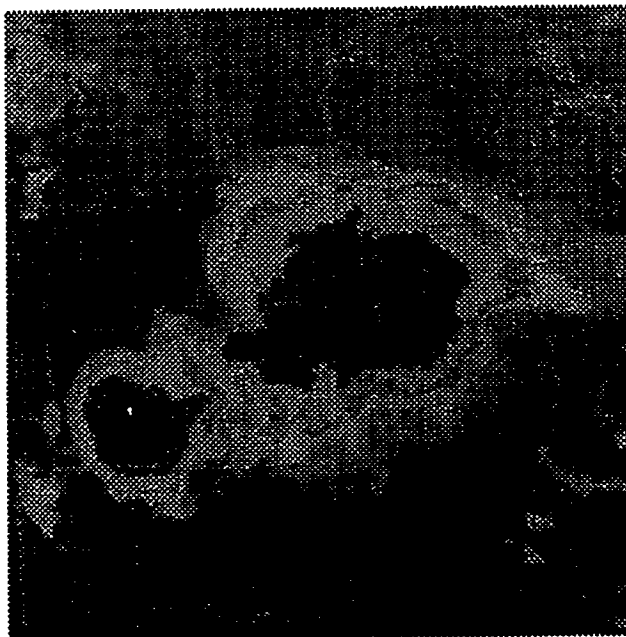


Figure 1. Infrared satellite imagery of the 23-24 June PRE-STORM MCS at 0400 UTC 24 June 1985. The portion of the PRE-STORM MCS that we simulate is near the intersection of Iowa, Missouri, and Illinois.

surface temperatures as high as 39°C and surface dewpoints as high as 24°C. Meanwhile, at 850 mb, a strong southerly jet (maximum wind speeds $> 15 \text{ ms}^{-1}$) over the southern Plains states provided a continuing supply of warm, moist air. Convective cells first formed around 1900 UTC 23 June along the dryline and front. Convective cells in northern Kansas and southern Nebraska moved toward the east-southeast; those in central and southern Kansas moved toward the south. By 0000 UTC 24 June, convection in the northeastern part of the area had consolidated into a large MCS in eastern Nebraska and Iowa along and to the south of the front (courtesy of an old outflow boundary). Another area of thunderstorms was parked over west-central Kansas along the dryline, and eventually blossomed into a smaller MCS. Figure 1, a 0400 UTC infrared satellite image, shows

the smaller MCS over western Kansas, and the larger one over northern Missouri and southern Iowa. Our fine grid zeroes in on the latter MCS.

Simulation

A discussion of the model setup is provided in this section. The discussion includes information on timestepping, boundary conditions, and the grid setup.

Timestepping

Prognostic model variables in the Regional Atmospheric Modeling System (RAMS) (Pielke et al. 1992) include the three velocity components; the perturbation Exner function; the ice-liquid potential temperature; the total water mixing ratio; and the mixing ratio of rain droplets, snowflakes, pristine ice crystals, graupel particles, and aggregates. The bulk hydrometeors have prescribed exponential size distributions. Prognostic equations use a time-splitting technique—allowing the model to explicitly compute on a small timestep those terms governing sound waves and to compute on a long timestep those terms governing other processes. Horizontal time differencing for long time steps is a flux conservative form of second order leapfrog (Tripoli and Cotton 1982). The long timesteps are 30, 15, and 7.5 seconds on grids with 25-km, 8.333-km, and 2.083-km horizontal grid spacing respectively; short timesteps are one-fifth this long.

Boundary Conditions

Lateral boundary conditions are the Klemp and Wilhelmson (1978a,b) radiative type, in which the normal velocity component specified at a lateral boundary is effectively advected from the interior assuming a specified propagation speed. We also use a Davies nudging condition, which causes model data at and near the lateral boundaries to be forced toward available observations. At the top boundary, we use a rigid lid, in concert with a Rayleigh friction absorbing layer. The latter damps gravity wave and other disturbances which approach the top boundary. At the lower boundary, we provide the model with topographic data, which has a horizontal spacing of 10 minutes latitude/longitude on Grid #1 and 30 seconds latitude/longitude on the other grids. A vegetation type data set from the National Center for Atmospheric Research with 11 primary vegetation types at a 5-minute latitude/longitude resolution is interpolated onto the model grids and then converted to the vegetation classification used in the model (the model recognizes 18 vegetation types). In addition, horizontally variable soil

moisture is based on the soil moisture analysis in the United States Department of Agriculture publication, Weekly Weather and Crop Bulletin (WWCB). The WWCB soil moisture index data was manually transferred to a latitude/longitude gridded data set at 1° resolution. This data set was then filtered and interpolated onto the model grid where it was converted into a soil moisture percentage.

Grid Setup

We initialize the model with the National Meteorological Center analyses, which provide the horizontal wind components, temperature, and relative humidity at 2.5° latitude/longitude intervals and at the mandatory pressure levels. We start the simulation at 1200 UTC (0600 LST over the MCS region) 23 June 1985. The simulation actually uses two grid setups. Prior to 0000 UTC, we run the model with three grids, with horizontal grid spacings of 75 km, 25 km, and 8.333 km. After 0000 UTC, we add another fine grid (2.083 km), but eliminate the coarse 75-km grid. Thus, after 0000 UTC, the model's horizontal grid spacing on the three grids is 25 km, 8.333 km, and 2.083 km. Hereafter, we call these grids Grid #1, #2, and #3, respectively (Figure 2). We use 32 vertical levels, stretched from a spacing of 175 m near the surface to 1000 m at the model top (~21 km). During the first 9 hours of the simulation, we use only the two coarsest grids. We activate Grid #3 at the 9-hour mark and Grid #4 at the 10-hour mark. Between

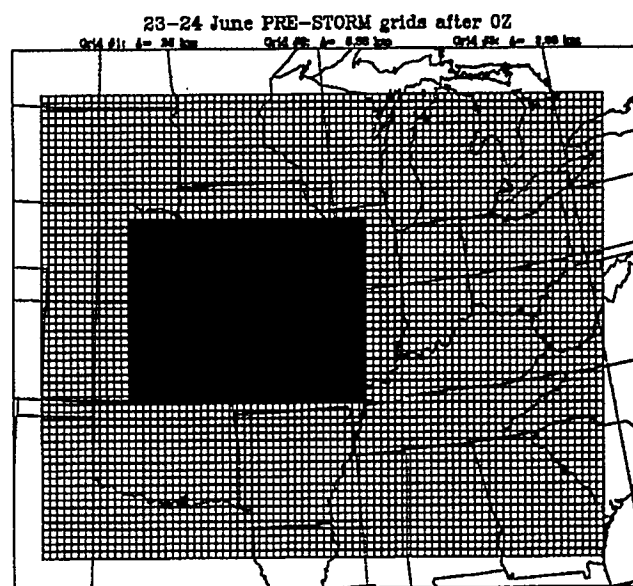


Figure 2. Grid setup for the PRE-STORM simulation after 0000 UTC 24 June 1985. The horizontal grid spacing is 25, 8.333, and 2.083 km on Grids #1, #2, and #3, respectively.

1900 and 2230 UTC, we use the Level 2.5w convective parameterization scheme (Weissbluth and Cotton 1993) on the 8.333-km grid; thereafter, we turn it off on all grids. Thus, we explicitly simulate the PRE-STORM 23-24 June 1985 MCS (no convective parameterization) between 2230 and 0400 UTC, writing analysis files to disk every 15 minutes. All of the model results we show are from Grid #4 between 0000 UTC and 0400 UTC 24 June 1985.

Results

The RAMS simulation captures many of the key features of the 23-24 June 1985 MCS. At 2000 UTC 23 June, an east-west oriented line of parameterized convection had developed in east-central Iowa along the simulated frontal boundary. The overall pattern of parameterized convection agreed well with observed convective activity at this time. Shortly thereafter, at 2230 UTC, a new band of convection developed in extreme southeast Nebraska, while the convection in Iowa intensified. At this time, simulated rainfall rates in south-central Iowa exceeded 5 cm^{-1} , in good agreement with observations. Storm reports in this region indicate over 12 cm of rainfall in Madison, Clark, and Warren counties of Iowa in the late afternoon. For the next several hours (until the end of the simulation, at 0400 UTC), in agreement with observations, the simulated convection propagated southeast at about 30 knots. Comparison of simulated convection with contemporaneous radar observations shows generally good agreement between the simulation and observations, although the simulated convection does not extend quite as far west as the actual convection. Figure 3 shows the 500-mb condensate mixing ratio at 0400 UTC, corresponding to the time of the satellite image in Figure 1.

Analysis of Simulation Results

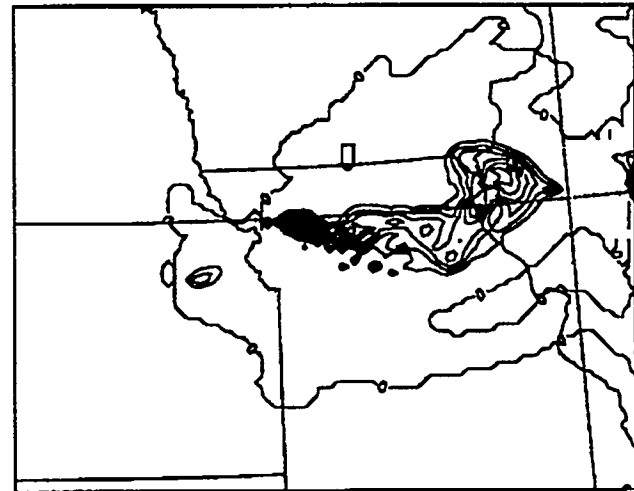
We intend to use output from both of our explicit MCS simulations (this one and a tropical MCS simulation) to calibrate a CGM convective parameterization scheme which accounts for mesoscale flow branches (i.e., mesoscale updrafts and downdrafts). The parameterization scheme is still under development, although its framework is largely complete. The convective part of the scheme is the Arakawa-Schubert parameterization (e.g., Arakawa and Cheng 1993) modified to account for the effects of convective downdrafts following Johnson (1976). In order to account for a more physically realistic coupling between cumulus convection and associated mesoscale cloudiness, the scheme employs a prognostic closure (as opposed to a quasi-equilibrium closure), as described by Randall and Pan (1993). The parameterized convection provides condensed water, ice, and water vapor which drives a parameterization

PRE-STORM

23-24 June 1985

Grid #2

From .0000E+00 To .1180E-01 By .7000E-03



Condensate MR

p = 500 mb t = 400 UTC

Figure 3. Condensate mixing ratio at 500 mb on Grid #3 at 0400 UTC 24 June 1985.

for the large-scale effects of mesoscale circulations associated with the convection. We conditionally sample mesoscale updrafts and mesoscale downdrafts within our explicit simulations in order to calibrate the shape and magnitude of the vertical profiles of mesoscale ascent, descent, phase transformations, and eddy flux convergences of moisture and entropy within parameterized mesoscale updrafts and downdrafts. Our scheme accounts for the effect of organized convection on the large-scale wind field by employing a cumulus momentum parameterization which includes the convective-scale pressure-gradient force, following Wu and Yanai (1994).

A key part of our parameterization, naturally, will be to determine when and where it will be necessary to activate the mesoscale part of the scheme. We have hypothesized that potential vorticity (PV) could be used as a marker of MCS activity. In an attempt to explore this possibility of using PV as a marker in the decision-making process of when and where to activate the MCS parameterization scheme, the vertical integration of area averaged PV variance ($\overline{q'q'}$) was calculated over a mid-tropospheric layer for the output from an explicit simulation of the 10-11 June 1985 PRE-STORM squall line for the time period while the squall line evolved from its initial stages to the beginning of its dissipating stage. The PV variance

evolved from a small positive value at the initial stage to the maximum value when the convection was the strongest. The center of the maximum was located to the north end of the squall line and highly correlated with the squall line thereafter. The above results are encouraging in that PV variance has reflected the life-cycle of the squall line from the preliminary analysis of one case. Further analysis and study of more cases are under way in order to determine whether it is possible to use PV as a marker.

Conclusions

We have used RAMS to perform a three-dimensional simulation of the PRE-STORM 23-24 June 1985 mesoscale convective system. RAMS successfully captures the observed characteristics of the MCS, including temporal initiation, geographic location, speed and direction of motion of the system, and spatial arrangement of the convective cells and stratiform region. We intend to use output from both of our explicit MCS simulations (this one and a tropical MCS simulation) to tune a convective parameterization scheme which includes mesoscale effects.

Acknowledgments

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A Comparison of Ground and Satellite Observations of Cloud Cover to Saturation Pressure Differences During a Cold Air Outbreak

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Introduction

The role of clouds in the atmospheric general circulation and the global climate is twofold. First, clouds owe their origin to large-scale dynamical forcing, radiative cooling in the atmosphere, and turbulent transfer at the surface. In addition, they provide one of the most important mechanisms for the vertical redistribution of momentum and sensible and latent heat for the large scale, and they influence the coupling between the atmosphere and the surface as well as the radiative and dynamical-hydrological balance.

In existing diagnostic cloudiness parameterization schemes, relative humidity is the most frequently used variable for estimating total cloud amount or stratiform cloud amount. However, the prediction of relative humidity in general circulation models (GCMs) is usually poor (Slingo 1980). Even for the most comprehensive GCMs, the predicted relative humidity may deviate greatly from that observed, as far as the frequency distribution of relative humidity is concerned. Recently, there has been an increased effort to improve the representation of clouds and cloud-radiation feedback in GCMs, but the verification of cloudiness parameterization schemes remains a severe problem because of the lack of observational data sets.

In this study, saturation pressure differences (as opposed to relative humidity) and satellite-derived cloud heights and amounts are compared with ground determinations of cloud cover over the Gulf Stream Locale (GSL) during a cold air outbreak.

Methodology

Saturation level pressures (p^*) (i.e., lifting condensation level) are derived from ETA model analyses. For unsaturated parcels, p^* is the intersection of the dry adiabat through the parcel temperature and the constant q line through the dewpoint temperature. For cloudy saturated parcels, it is the

intersection of the moist adiabat through the parcel temperature with the q isopleth now equal to the total water mixing ratio (vapor + liquid). Another very useful parameter is the saturation pressure difference (SPD) which has been used to distinguish layers that are in and above the convective boundary layer (Betts and Albrecht 1987). Following Betts (1982), it is defined as

$$\text{SPD} = p^* - p \quad (1)$$

(the difference between air parcel pressure, p , and saturation level pressure, p^*). The usefulness of this parameter has also been shown in Alliss and Raman (1995c). In this study, however, the degree of subsaturation of a given layer will be documented and compared with available surface-satellite observations of cloud cover. Negative values of SPD are directly related to subsaturation since $q < q_s$ (where q_s = saturation mixing ratio). Thus SPD may be viewed as the "potential" for parcel saturation.

The CO_2 slicing technique is used to calculate cloud top pressures and effective cloud amounts (Wylie and Menzel 1989). This technique has also been used to show seasonal and diurnal variations over the GSL (Alliss and Raman 1995a,b). For a complete description of both the satellite and SPD methods in estimating cloud fraction, the reader is referred to Alliss (1995).

Results

Since ETA model analyses/forecasts and satellite CO_2 radiances are available over the Western Atlantic Ocean the composite cloud fraction algorithms are used to test the performance during a cold air outbreak observed on 6 February 1995. Geostationary Operational Environmental Satellite (GOES-I) radiances are obtained to evaluate a composite cloud fraction for several locations in the GSL at 1800 UTC. Saturation pressure differences obtained from a six-hour forecast valid at 1800 UTC, 6 February 1995 are

also used to estimate composite cloud fractions. On this day, the GSL is dominated by high pressure with an offshore flow of cold Polar air. Satellite imagery indicates the familiar cloud free zone between the Coast and Gulf Stream core. Employing the satellite and SPD algorithm to data at RDU, GSO, and HAT indicates clear sky at all locations, which agrees with surface reports. Figure 1 shows profiles of SPD at four locations. The four profiles are separated by 1 degree of longitude along 34 north latitude. The first profile (34N/78W) is nearest ILM. This profile shows extremely dry conditions with an average SPD of -203 mb. Thus, according to the algorithm, clear skies would be identified. ILM reports clear skies at 1800 UTC. The corresponding mean relative humidity for this layer is 27%. The second profile (34N/77W) indicates a moistening of the atmosphere below 900 mb compared with profile one.

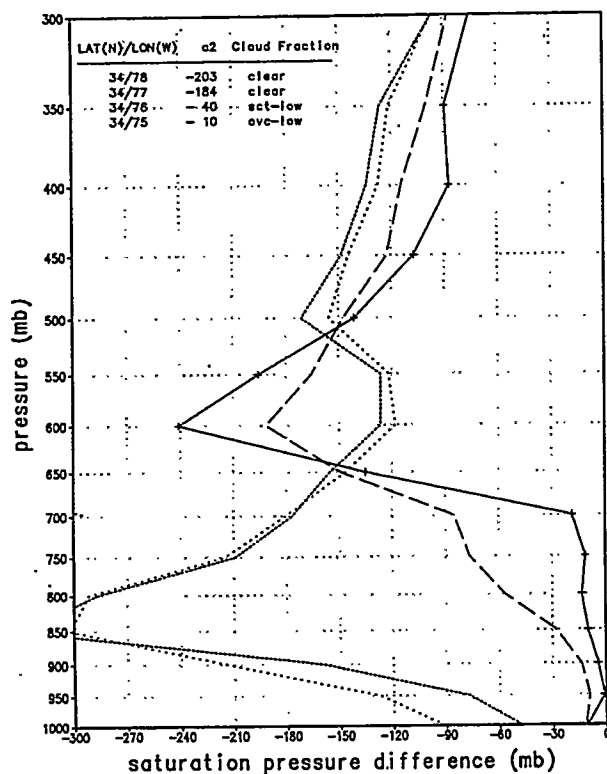


Figure 1. ETA model six-hour forecasts of SPD during the cold air outbreak observed on 6 February 1995 1800 UTC. The dotted line indicates the profile at 34N/78W, short dash (34N/77W), long dash (34N/76W) solid line (34N/75W). The value a2 is defined as the mean SPD in the 1000-700 mb layer.

Although surface SPDs are approaching saturation, the layer as a whole remains too dry to support cloudiness ($a_2 = -184$ mb). This position is approximately between the coastline and the Gulf Stream core. Profile three (34N/76W) indicates that intense modification of the airmass has begun. The mean SPD has increased from -184 mb to -40 mb, indicating scattered low clouds are probably present. For reference, the mean relative humidity of the layer is 78%. The profile at 34N/75W lies just east of the Gulf Stream core and indicates nearly saturated conditions from the surface to 700 mb. The mean SPD in this layer is approximately -10 mb. In this case the algorithm predicts a low overcast cloud within the 1000-700 mb layer. A mean relative humidity of 99% is also calculated at this time. According to satellite imagery, considerable low clouds are present. No clouds are observed above 700 mb as the atmosphere remains very dry and stable.

The CO₂ slicing technique is used to determine a composite cloud fraction at the same four locations. Table 1(a-d) shows the histograms of satellite determinations. Entries in Table 1 are percentage values for a given category of all observations within the 50-km by 50-km grid over the given ground site. By applying the percentage values in Table 1 to thresholds (not shown) the composite reports are defined. Over ILM (34N/78W), 100% of the satellite observations indicate clear skies (Table 1a). This agrees with both ground reports and the SPD determinations. Clear skies are also indicated at 34N/77W as the satellite again reports 100% clear skies (Table 1b). At location three (34N/76W), the satellite detects the presence of some low clouds (Table 1c). In this case, only 30% of observations indicate clear sky; therefore a low opaque cloud cover is indicated. Of the 70% of low opaque clouds the satellite reports, the majority of the cloud top pressures are near 900 mb. Finally, at location four (34N/75W), the satellite sees 100% low opaque cloud cover with the average cloud top pressure near 750 mb (Table 1d). Unfortunately there were no surface reports over the ocean which contained cloud cover observations, therefore no direct validation could be made. However, based on a qualitative comparison with visible imagery, the two techniques appear to be performing well.

Acknowledgments

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Table 1. Histograms of satellite cloud determinations at (a) ILM, (b) 34N/77W, (c) 34N/76W and (d) 34N/75W at 1800 UTC, 6 February 1995.

(a) Effective Cloud Amount				
Cloud Top Pressure (CTP)	≤ 0.33	≤ 0.65	≤ 0.95	≤ 1.0
CTP < 400	0	0	0	0
400 $\leq$ CTP < 700	0	0	0	0
700 $\leq$ CTP < 1000	x	x	x	0
CTP = 1000	100	x	x	x
Satellite reports clear skies.				
(b) Effective Cloud Amount				
Cloud Top Pressure (CTP)	≤ 0.33	≤ 0.65	≤ 0.95	≤ 1.0
CTP < 400	0	0	0	0
400 $\leq$ CTP < 700	0	0	0	0
700 $\leq$ CTP < 1000	x	x	x	0
CTP = 1000	100	x	x	x
Satellite reports clear skies.				
(c) Effective Cloud Amount				
Cloud Top Pressure (CTP)	≤ 0.33	≤ 0.65	≤ 0.95	≤ 1.0
CTP < 400	0	0	0	0
400 $\leq$ CTP < 700	0	0	0	0
700 $\leq$ CTP < 1000	x	x	x	0
CTP = 1000	100	x	x	x
Satellite reports low opaque cloudiness.				
(d) Effective Cloud Amount				
Cloud Top Pressure (CTP)	≤ 0.33	≤ 0.65	≤ 0.95	≤ 1.0
CTP < 400	0	0	0	0
400 $\leq$ CTP < 700	0	0	0	00
700 $\leq$ CTP < 1000	x	x	x	100
CTP = 1000	0	x	x	x
Satellite reports low opaque cloudiness.				

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Radiative Heating in Global Climate Models

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Climate Prediction and Radiative Heating

Climate model dynamics are driven by external and internal forcing. The primary forces affecting the thermal field are long wave radiative (LWR) heating, short wave radiative (SWR) heating, and convection (cumulus, etc.). These forcing effects are cycled through the thermal field to the motion field by nonlinear transfer. The model dependent variables, in particular temperature (T), moisture (Q), and especially clouds, evolve in time in a Global Climate Model (GCM) and thereby determine the subsequent forcing. If the dependent variables are not accurately calculated in space and time, the predicted forcing functions will be adversely affected. As integration time proceeds, such inaccuracies will lead to systematic errors in the prediction of climate.

Since it is neither possible nor advisable to study the effects of all these forcing functions simultaneously, we focus here on LWR heating and attempt to assess how sensitive this force is to its required input variables. This forcing is determined in GCMs by a LWR heating algorithm. Such algorithms compute heating rate (HR) profiles from the profiles of T, Q, clouds, and minor constituents. The HRs depend on the vertical structure of the input variables in addition to the physics built into the algorithm.

Sensitivity of LWR Algorithms to Input Variables

We tested the sensitivity of various algorithms taken from GCMs. The following seven algorithms were chosen: Canadian Climate Center (J.-P. Blanchet), ECMWF (J.-J. Morcrette), NCAR (J. Kiehl), Colorado State University (D. Randall), University of Maryland (R. Ellingson), Recherche en Prevision Numerique (L. Garand), and the National Meteorological Center (K. Campana). These models represent a cross section of GCMs which include LWR heating algorithms. For reference we selected the line-by-line model of Clough (1992) which is highly accurate under clear sky conditions. The algorithms were tested on a variety of data profiles to cover different

geographic regions and seasons. The standard McClatchey et al. (1971) and a selection of Phillips et al. (1988) soundings were used for clear sky conditions. For tests under cloudy sky conditions, we used soundings taken from the Atmospheric Radiation Measurement (ARM) Cloud and Radiation Testbed (CART) Great Plains site in Oklahoma.

Figure 1 shows heating rates generated by all seven LWR algorithms for the mid-latitude summer (MLS) and mid-latitude winter (MLW) McClatchey et al. data sets under clear sky conditions and compared to the line-by-line (CLO) calculation. Both the results for 30 and 18 vertical levels are presented. Note that even for this statistically averaged data, the models tend to vary by as much as one-half degree K per day, independent of the number of levels chosen. It is noteworthy that the model HRs cluster about the reference CLO results.

The Phillips et al. data allow us to look more closely at individual soundings and their impact on the HRs. Using individual sounding Phillips et al. data as input to the algorithms for 100 soundings in each season and regional grouping, the mean HRs generated by each model were averaged and standard deviations on each level were calculated. Results show that on average the models do not conform well to the reference (CLO) in the tropics, but do better in MLW. The effect of variability amongst the individual Phillips et al. soundings within a data set on HRs was also tested and showed that the standard deviations of HRs derived from the soundings for each of the algorithms considered were substantial and also differed amongst the models.

The variability amongst the algorithms is implicit in the construction of the LWR model physics and is related to the input parameters as described above. The sensitivity to moisture is highlighted by comparing HRs from the algorithms using McClatchey et al. MLS and MLW data with and without moisture. Although almost all the cooling disappears in the troposphere without moisture, the variability amongst the models remains.

Inclusion of clouds exacerbates this variability. Since clouds are difficult to measure let alone predict in a GCM,

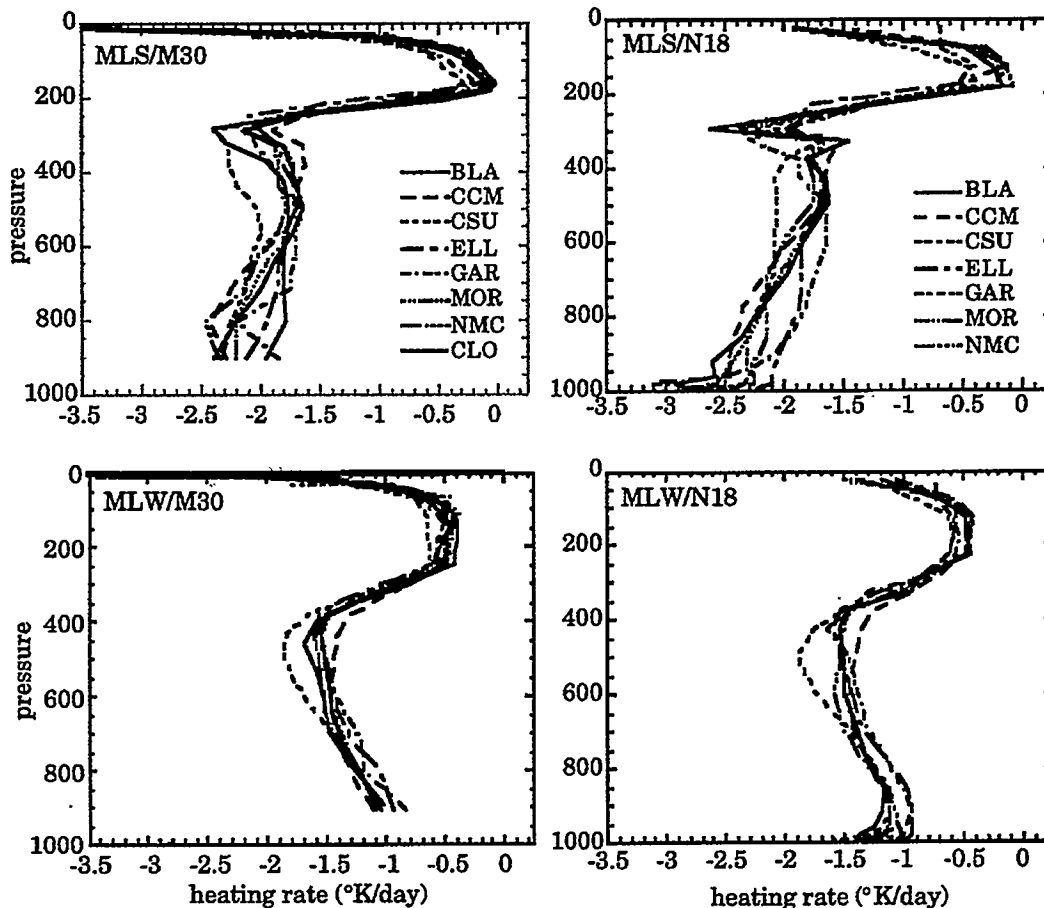


Figure 1. Heating rates generated by the seven tested algorithms and the reference algorithm (CLO) for the McClatchey data sets representing mid-latitude summer and mid-latitude winter.

the LWRM response to clouds was first tested by inserting them artificially at given levels and with fixed intensity. We studied the response to single level clouds (thin clouds) by inserting them one at a time at the lowest eleven model levels of the 30-level distribution. The resulting HR profiles show extreme sensitivity to clouds and overwhelm the effects of temperature and moisture in the vicinity of the cloud.

The algorithms were subsequently tested under real cloud conditions. We chose to perform this test with data taken from the May 1994 ARM archives at the Oklahoma site. Figure 2 shows the heating rates produced for two cloudy soundings as representative of a number of soundings tested. To demonstrate the effect of clouds on the HRs, we show the HRs which are produced by the algorithms for the same cloud structure but with the sounding data (T and Q) taken

from the McClatchey et al. MLS data set (Figure 2, panel d). The cloud structure of the data samples presented are included in the figure.

Our observations considering a number of cloudy soundings are the following. For clear skies, the profiles depend exclusively on the soundings which differ from Oklahoma (clear) to MLS/MC, and the algorithms differ from one another as seen previously. For only thin low clouds (one level) the cloud effect is clearly seen locally and the algorithms' output begins to differ from one another. Concurrently, the sounding data tend to become less important. As the clouds become thicker and/or move higher in the troposphere, the effects just noted become amplified (Figure 2) to the extent that each algorithm gives a unique profile different from all others.

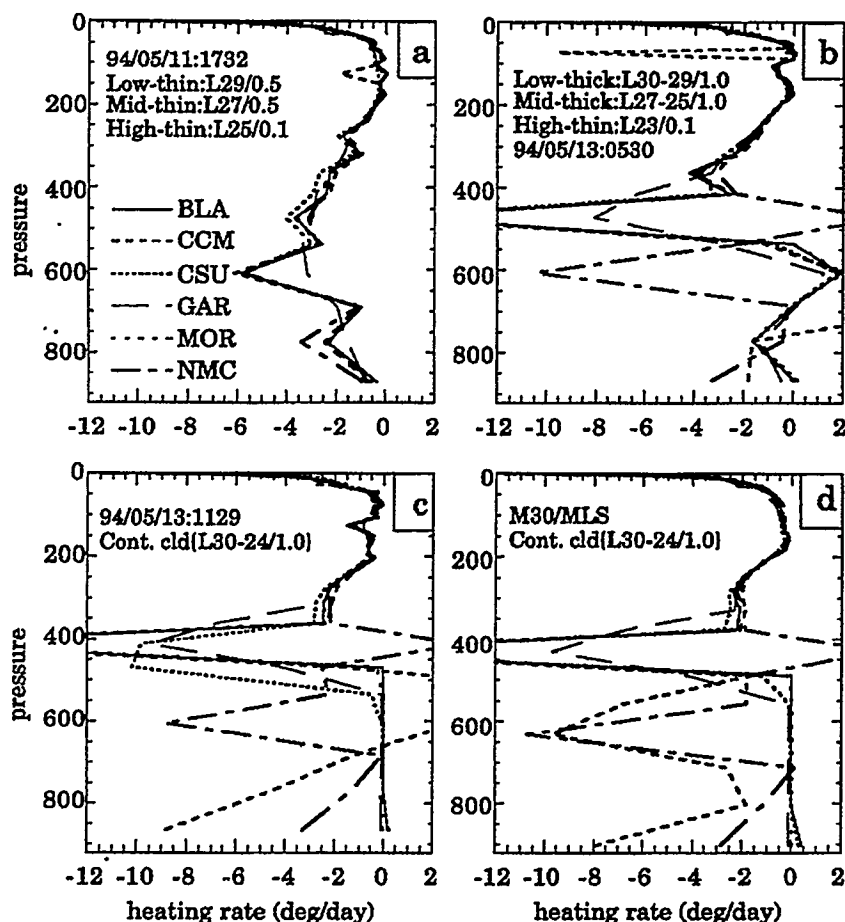


Figure 2. Heating rate profiles generated by six algorithms from observed data including clouds. Note cloud level/density (lxx/xx.x).

The Affect on GCMs by HRs Generated from LWR Algorithms

LWR algorithms calculate vertical profiles of HRs at each horizontal point in a GCM, and these are calculated periodically in time. Thus they produce three dimensional fields. These HR fields represent the appropriate LWR forcing which directly determines the temperature tendencies. The temperature field predicted from these tendencies modifies the wind field tendencies by nonlinear interaction in space. The predicted wind field then modifies the temperature tendencies by nonlinear advection. Thus the impact of the HRs is spread in time and space to all the variables (T, Q, clouds). These modified variables are then used to recalculate new HRs.

Three-dimensional HR fields from GCMs were used to demonstrate heating rate sensitivity to GCM characteristics. Model output of LWR HR fields at each archived time were averaged over 60-day wintertime (Jan-Feb) periods for comprehensive statistics. Models available for this analysis included

- NCAR CCM1: R15, T42; 12 levels; climatology
- NCAR CCM2: R15, T42, T106; 18 levels; climatology
- NCAR CCM2: T42; 18 levels; AMIP.

Previous observations from this study noted the following:

- *Truncation* plays an important role. Considering both the CCM1 and the CCM2, the cooling distribution is radically different for all space scales from one truncation run to another, other factors remaining identical.
- HRs are dependent on *model changes*, including changes to the incorporated LWR algorithm. Comparing output from the CCM1 and CCM2 runs with the same time interval chosen for each model and beginning with identical initial conditions, the HR output is substantially different.
- *Surface forcing* in the form of sea-surface temperature (SST) does not show a significant impact on the evolution of the T, Q, and cloud fields in a model and the resulting HRs generated by the LWR algorithm. Two wintertime periods (Jan-Feb 1987 and Jan-Feb 1983) from the CCM2 AMIP archive and the CCM2 wintertime climatology were compared. Note that the AMIP run has observed SST and the climatology run has climatological SST. The differences in HR statistics of the two AMIP samples were not significant, nor did they differ substantially from the run with climatological SST.

Reference to Observations

In general, there are no observed measurements of LWR heating in the atmosphere. Thus comparisons of model results to observations cannot be made. However, one could take the observed data during the AMIP period which have been archived and calculate instantaneous HR fields using the LWR algorithm in the GCM. Unfortunately observational cloud information is not available and, as seen earlier in this study, clouds are essential to the algorithm. As an alternative, we used the GCM to generate clouds from the observed data and used those clouds to generate HRs in the model. Specifically, we introduced observed data into the CCM2 as initial conditions for each day of the AMIP period (Jan-Feb 1987) and ran the model for 36 hours. We used the HR fields which the model developed at this time and defined them as "observations" (our approximation) to compare with the output of various AMIP model runs.

GCM Intercomparisons

To compare the predictions of various GCMs, we collected data fields from three individual models: the NCAR CCM2 model, the Colorado State University (CSU) Model, and the NASA/GSFC GEOS model, as well as the observations

noted above. The data were taken from the AMIP runs of each model and averaged for the period of Jan-Feb 1987.

Global maps were available at three levels (850 mb, 500 mb, 250 mb) for the field variables long wave HRs, cloud fraction, temperature, and moisture. To give a flavor of the differences, we present on Figure 3 the heating rate patterns of the models at 250 mb and the clouds at 500 mb.

Differences amongst the GCMs in all variables considered were pronounced. The CCM2 appears a bit closer to observations, probably because the same LWRM algorithm is used. The temperature and moisture fields are least different amongst the GCMs, but also do not compare well. The differences in the cloud fields amongst the GCMs are probably the primary factor leading to the large differences in HRs. Note that no detailed cloud maps were available for the GEOS-1. These results confirm our previous observations about clouds. However we did not know to what extent the GCMs would differ from one another.

Conclusions

LWR algorithms from various GCMs vary significantly from one another for the same clear sky input data. This variability becomes pronounced when clouds are included. We demonstrate this effect by intercomparing the various models' output using observed data including clouds from ARM/CART data taken in Oklahoma.

The LWRM algorithms play a vital role in GCMs insofar as they redistribute the HRs produced by their LWRM algorithm. Our analyses indicate that the LWR heating in the GCM depends

- significantly on model truncation
- significantly on model construction (including the LWR algorithm)
- less on surface heating effects
- most notably on clouds and their parameterization.

Intercomparison of AMIP statistics from several GCMs indicates substantial differences in model HR output under identical input conditions. We thus conclude that GCMs at the very least need better cloud parameterization and probably better LWRM algorithms.

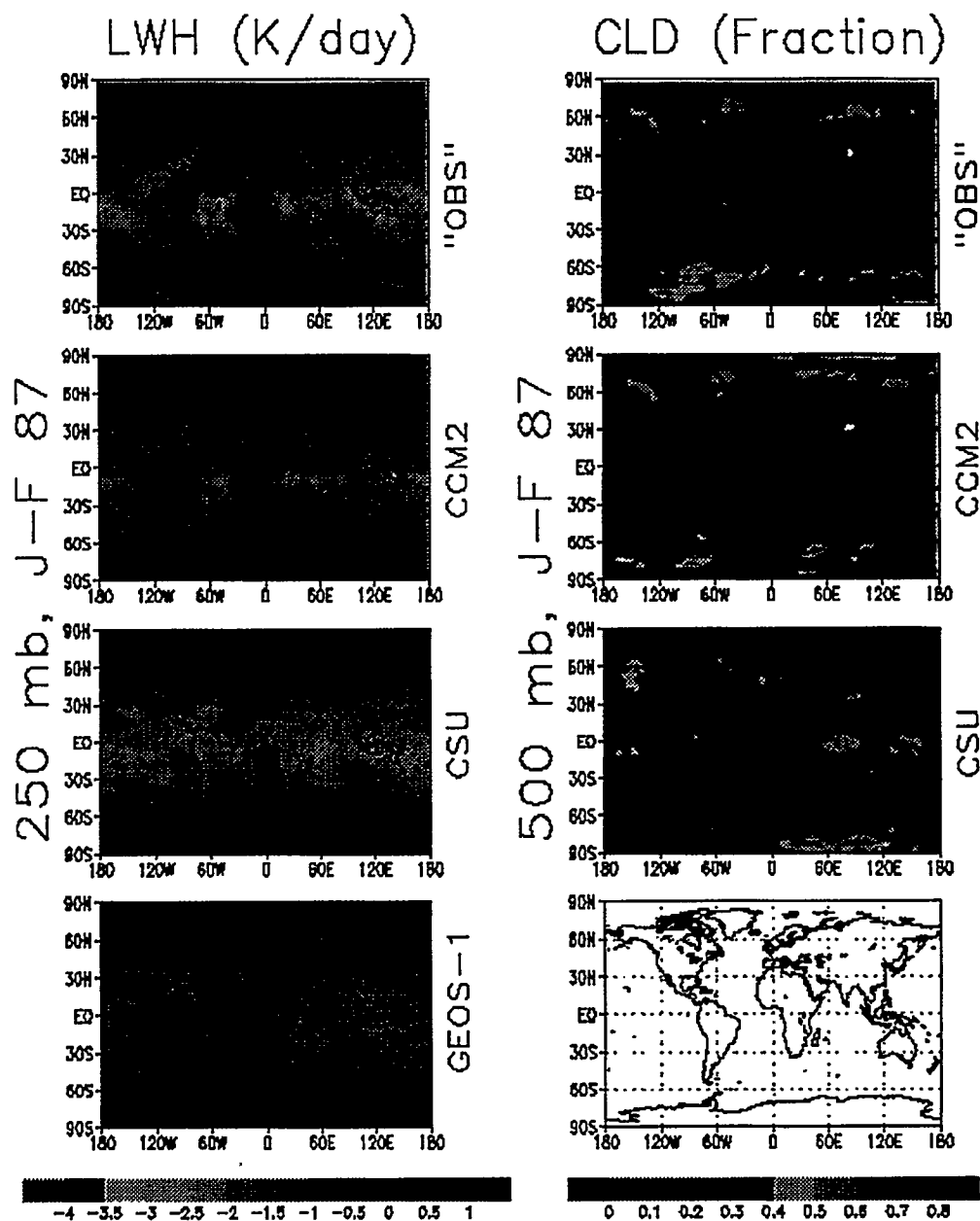


Figure 3. Maps of averaged HRs and clouds for the CCM2, CSU, and GEOS models and corresponding 'obs.' See text for details.

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Evidence for Island Effects and Diurnal Signals in Satellite Images of Clouds Over the Tropical Western Pacific

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Introduction

Instruments to measure atmospheric radiation and ancillary meteorological variables will be set up on Manus Island as the first site of the tropical western pacific (TWP) locale of the Atmospheric Radiation Measurements (ARM) program. Manus is in the "warm pool" region of the TWP. This region is critical in establishing global atmospheric circulation patterns and is a primary energy source for the Hadley and Walker cells (Webster 1994; Skinner and Trapper 1994; Keenan et al. 1989). The myriad islands and enclosed seas in the immediate vicinity of Manus have been referred to as the "maritime continent" (Ramage 1968), which has the deepest convective activity in the world (Skinner and Trapper 1994; Keenan et al. 1989). Manus is in a region having a global impact on climate and where island effects on clouds are likely to be important. In this preliminary analysis we have sought evidence of island effects in the cloud fields around Manus and have studied the variability of the diurnal cycles of cloud cover over Manus and over other islands and areas of open sea in the region.

Methods

The data we have used are hourly images from the thermal infrared channel ($\lambda \sim 11 \mu\text{m}$) of the GMS-4 (Meteorological Satellite Center 1989), at 10-km resolution for the time interval of November 1992 to February 1993. These data are now available on CD-ROM (Japanese Pacific Climatic Studies 1993). Hovmuller diagrams are used here to depict the longitudinal variation in time of the infrared channel brightness temperatures (T_b) at 2.1°S , the latitude of Manus (Figure 1). From this, and other evidence, it is clear that

there were three distinct regimes during this period. Two regimes represent "undisturbed conditions" (December 1-18, 1992, and January 1-31, 1993), and one represents "disturbed conditions" during which a Madden-Julian event occurred (December 19-31, 1992) (Madden and Julian 1971).^(a) For these time periods we created images of average brightness temperature fields and diurnal cycles of probabilities of the occurrence of low, mid-level, and high clouds. These cloud divisions are determined using only infrared brightness temperature ranges: $T_b < 273^\circ\text{K}$, $220^\circ\text{K} < T_b < 273^\circ\text{K}$, and $T_b < 220^\circ\text{K}$. These are not the thermodynamic temperatures as they are not corrected for emissivity effects, nor for the intervening atmosphere. A comparison was made of diurnal cycles of cold clouds over the center of the Bismarck Sea and over three islands of varying size, Manus, Luf (Hermit Islands) and Mussau (Figure 2, Table 1).

Results and Discussion

The Hovmuller diagrams show a regular series of lower cloud temperatures around 147°E (lighter color), especially during December 1-7 and January 12-17 (Figure 1). These regular occurrences of convection are suppressed (or at least masked) during the periods when the cloud fields are associated with larger scale meteorological features. This suppression of diurnal cloud formation by large-scale features has been identified over Melville Island by Skinner and Tapper (1994), who ascribe the suppression of diurnal convection to both the reduction of the land-sea temperature differences resulting from the diminished surface insolation.

(a) D. Waliser, 1994, personal communication.

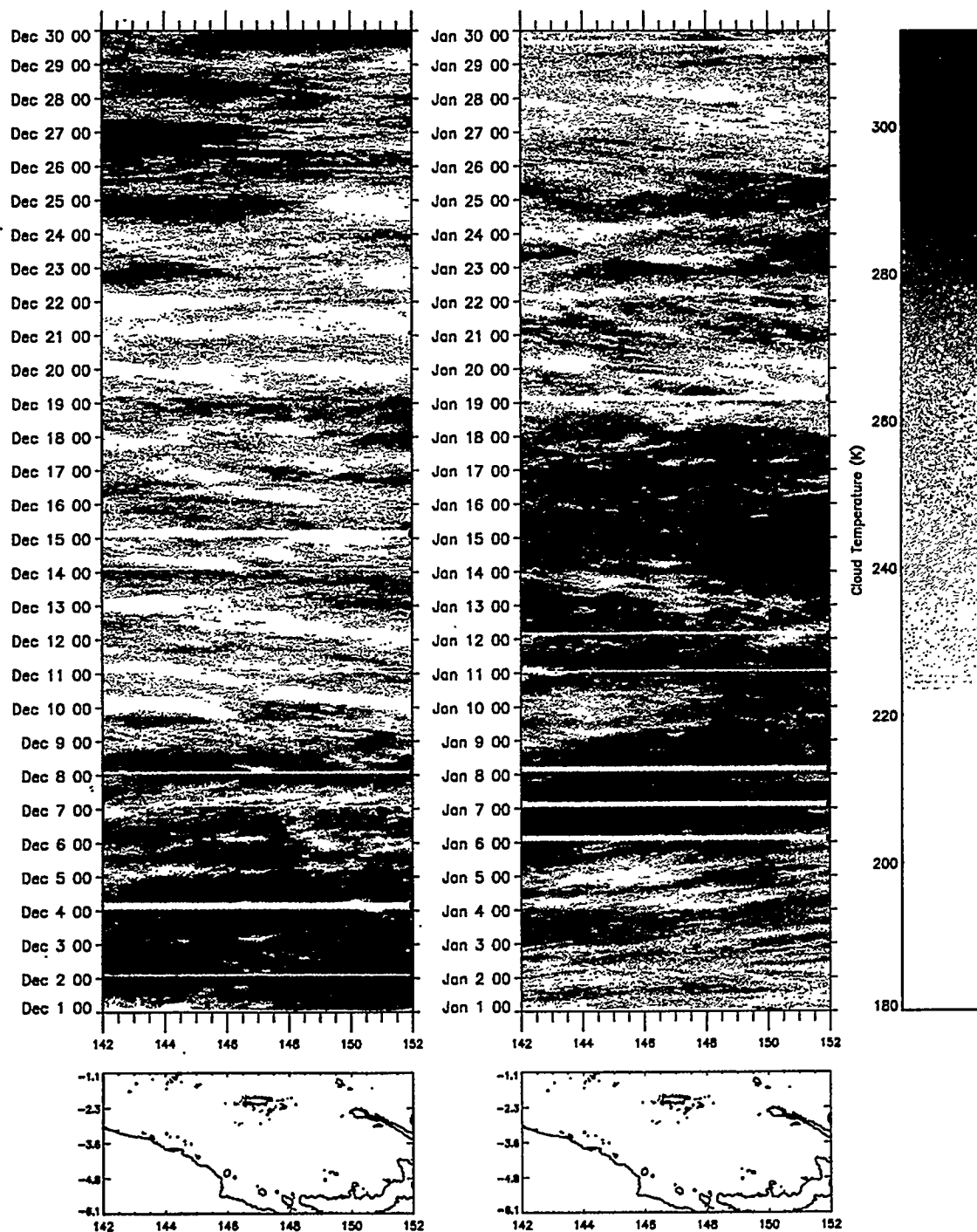


Figure 1. Hovmuller diagrams of satellite infrared channel brightness temperatures measured along the line of latitude 2.1°S, through the island of Manus, for the months of December 1992 and January 1993. The diurnal influence of Manus on the clouds is most apparent in conditions where the surrounding area is cloud free.

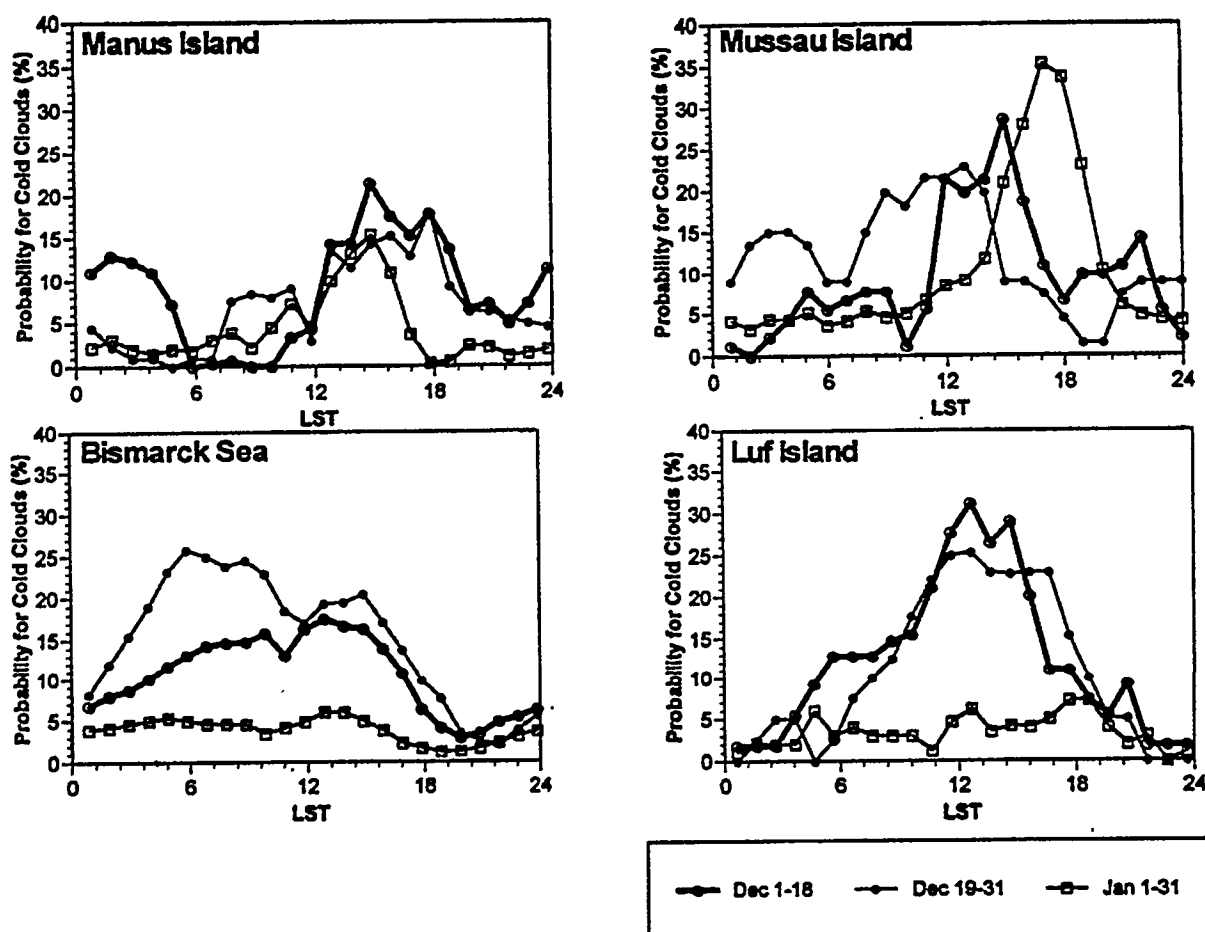


Figure 2. Diurnal cycles of the probability of the occurrence of high clouds ($T_b < 253^\circ\text{K}$) over Manus, Mussau and Luf Islands, and over the Bismarck Sea. Diurnal cycles for the three regimes discussed in the text are shown.

Table 1. Islands used in this study.				
Island	Location	Topography ^(a) (m)	Area ^(b) (km ²)	Perimeter ^(b) (km)
Manus	2.1S, 147.0E	718	1925	475
Luf	1.5S, 145.0E	111-244	7	20
Mussau	1.4S, 149.6E	651	346	102

(a) Defense Mapping Agency (1992).
(b) Calculated from coastline data of the World Vector Shoreline.

In the presence of extensive cloud cover and the inhibition of sea-breezes by the strong large-scale surface winds.

The average temperatures over Manus are lower than the surrounding area as are the probabilities of low clouds. The probabilities of mid clouds and high clouds are increased over the island. The diurnal cycles for the cold, mid-level, and high clouds surrounding $2^\circ \times 2^\circ$ have approximately the same shape but with somewhat reduced amplitudes as that over Manus island, and this appears to reflect the island's influence on the surrounding region (approximately 100 km). The shape of the diurnal cycles over and around Manus are different from those over the Bismarck Sea (Figure 2).

The diurnal cycles of the cold clouds show that Manus Island changes from having two peaks, one early in the morning and one in the afternoon for cold clouds in December 1-18, to a single peak in the afternoon during January (Figure 2). Mussau, an island a sixth the size of Manus

has three times the probability of high clouds over it during the evening. Even Luf Island, a small island has a higher probability of high clouds over it compared with Manus.

These results indicate that convective activity over even small islands can be significant. The possibility of small islands inducing considerable convection has also been observed elsewhere.^(a) Published studies indicate that cloud structure over an island is primarily controlled by its size and relief (Gray and Jacobson 1977), with large islands typically showing morning and evening maxima and small islands only a morning maximum in their diurnal cycles of rainfall and convection. Our preliminary analysis shows that the characteristic forms of the diurnal convection may be determined also by the influence of large-scale meteorology, with the same island displaying single or double peaks dependent on the larger-scale atmospheric features.

Conclusions

Based on our study, the following conclusions can be reached:

- Island effects are discernable in satellite thermal infrared images.
- Diurnal cycles of clouds are sometimes different over islands than over the open sea.
- Diurnal effects are influenced, probably suppressed, by larger-scale cloud fields or meteorological structures.
- Island effects will have consequences on the analysis of ARM surface radiometric and meteorological measurements as these are localized and subject to small, island-scale influences. When used with measurements covering the larger scale features, such as from satellite and aircraft, there is the potential for systematic discrepancies resulting from scale-mismatch.

(a) J. Simpson, 1994, personal communication.

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Analytical Study of the Effects of the Low-Level Jet on Moisture Convergence and Vertical Motion Fields at the Southern Great Plains Cloud and Radiation Testbed Site

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Introduction

The Atmospheric Radiation Measurement (ARM) Southern Great Plains (SGP) Cloud and Radiation Testbed (CART) is located in a region that is strongly affected by a prominent meteorological phenomenon--the Great Plains Low-Level Jet (LLJ). Observations (Means 1952; Pitchford and London 1962; Parish et al. 1988; Frisch et al. 1992) have shown that the LLJ plays a vital role in spring and summertime cloud formation and precipitation over the Great Plains. An improved understanding of the LLJ characteristics and its impact on the environment is necessary for addressing the fundamental issue of development and testing of radiational transfer and cloud parameterization schemes for the general circulation models (GCMs) using data from the SGP CART site.

A climatological analysis of the summertime LLJ over the SGP has been carried out using hourly observations from the National Oceanic and Atmospheric Administration (NOAA) Wind Profiler Demonstration Network and from the ARM June 1993 Intensive Observation Period (IOP). The hourly data provide an enhanced temporal and spatial resolution relative to earlier studies which used 6- and 12-hourly rawinsonde observations at fewer stations.

Data Description

The wind profiler network consists of 31 stations scattered mostly across the Great Plains. The densest part of the network is in Oklahoma and Kansas (Figure 1). This station distribution is nearly ideal for LLJ observations, as the maximum frequency and development of the LLJ occur here (Bonner 1968). In this study, hourly averaged wind profiles were used to provide data to maximum heights of 19 km, with vertical resolution of 250 m. Analyses were performed using a mean climatological day that was constructed from

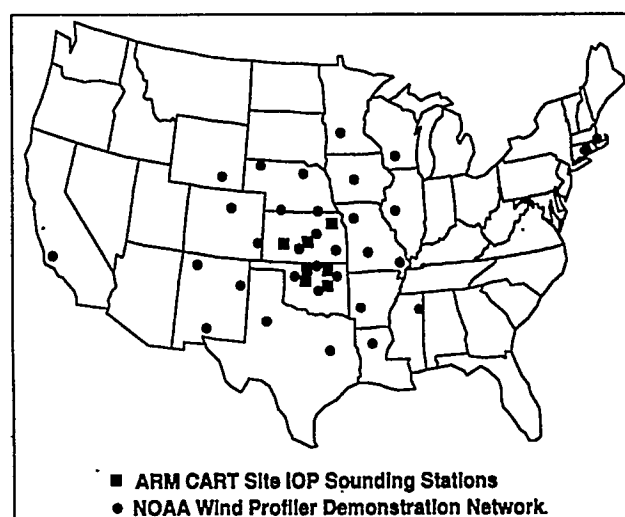


Figure 1. The NOAA wind profiler demonstration network and the sounding sites during the ARM June 1993 IOP.

hourly wind data obtained over the 62-day period-of-record from June 15 to July 15, 1991, and from June 15 to July 15, 1992.

LLJ Wind Characteristics

The longitudinal variation of mean winds is shown in Figure 2, which presents the 62-day-average vertical wind profiles for 1400 CST and 0200 CST at stations aligned W-E near the latitude of the Oklahoma-Kansas border. The mean wind is from the south and southwest in the lower atmosphere and veers gradually with height to blow from the west and southwest. The strongest winds are found between 95 to 98°W longitude. Winds in the daytime are relatively uniform with height due to strong vertical turbulent mixing, but at night between 95 to 98°W, a wind maximum appears

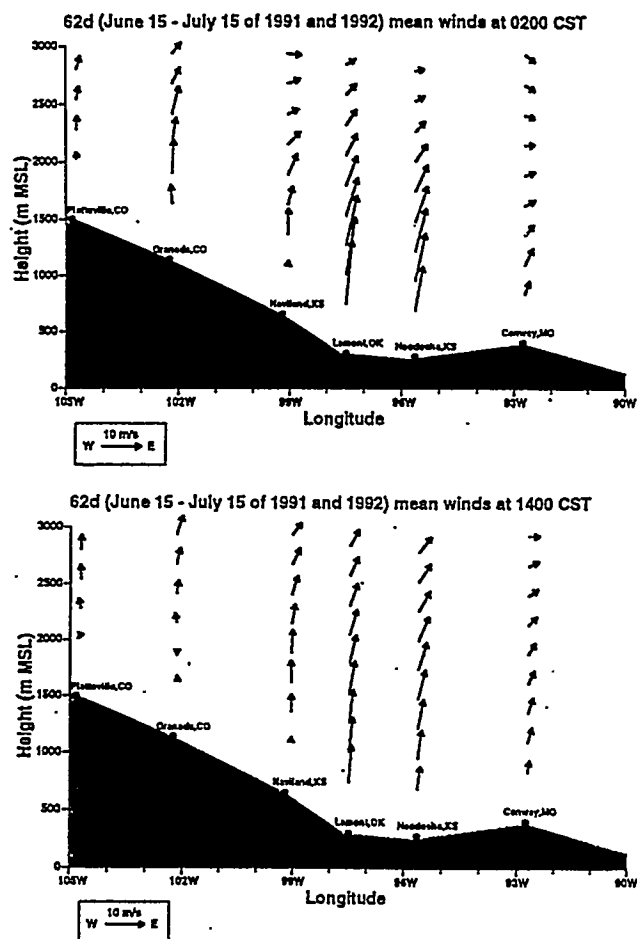


Figure 2. Vertical wind profiles at 1400 and 0200 CST averaged over the 62-day period from 15 June to 15 July in 1991 and 1992 at wind profiler stations aligned W-E near the latitude of the Oklahoma-Kansas border.

below 1000 m and the wind speed decreases rapidly with height, forming the LLJ. As the LLJ strengthens, its direction remains nearly constant.

The diurnal changes in wind speed and the latitudinal variation of the boundary-layer winds are documented in Figure 3, which shows the southerly wind component for a climatological mean day at Lamont, Oklahoma, and Palestine, Texas. Boundary-layer winds undergo a diurnal oscillation at both sites, with a nighttime maximum occurring around midnight and a daytime minimum occurring in the afternoon. However, the oscillation at Lamont leads that at Palestine by about 2 hours, suggesting that the timing of the LLJ has a latitudinal dependence. This earlier phasing at the higher latitudes is a feature of

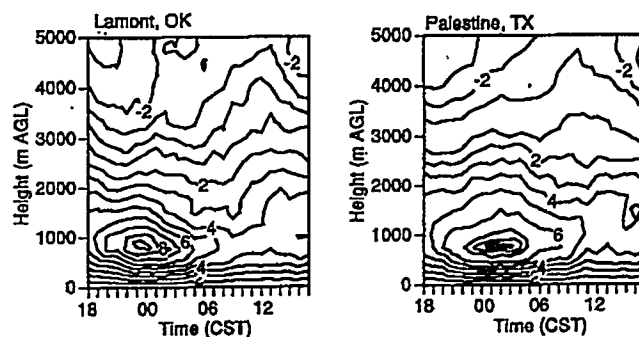


Figure 3. Time-height cross section of the south-north component of mean wind speed at Lamont, Oklahoma and Palestine, Texas.

observations at other sites (not shown), and is consistent with Blackadar's (1957) inertial oscillation theory.

The conclusion that the inertial oscillation seems to play an important role in the evolution of the LLJ is also suggested by the hodographs in Figure 4, which show a smooth clockwise turning of mean wind with time at all stations, consistent with predictions of the inertial oscillation theory.

Vertical Motion and Moisture Convergence

LLJ-induced mean vertical motion and moisture convergence are estimated for the ARM SGP CART site using data from soundings released at the CART site during the two-week ARM IOP in June 1993. Figure 5 shows the two-week-average vertical motions and the time rate of change of the mixing ratio for all hours of the day. The low-level moisture increased rapidly after sunset as the jet intensified, advecting moist air from the Gulf of Mexico into the south central Great Plains. Meanwhile, strong rising motions developed at night due to the low-level convergence caused by the LLJ. This rapid rise of warm, moist air acts as a triggering mechanism for the development of convective clouds and thunderstorms frequently observed in this region.

Frequency of the LLJ

The diurnal variation of LLJ frequency is shown in Figure 6 for stations around the CART site. At all stations, the LLJ frequency exhibits a strong diurnal oscillation. Although the frequencies at each hour differ a little from one station to another, the frequency distributions are nearly the same at all stations. The frequency increases dramatically from a

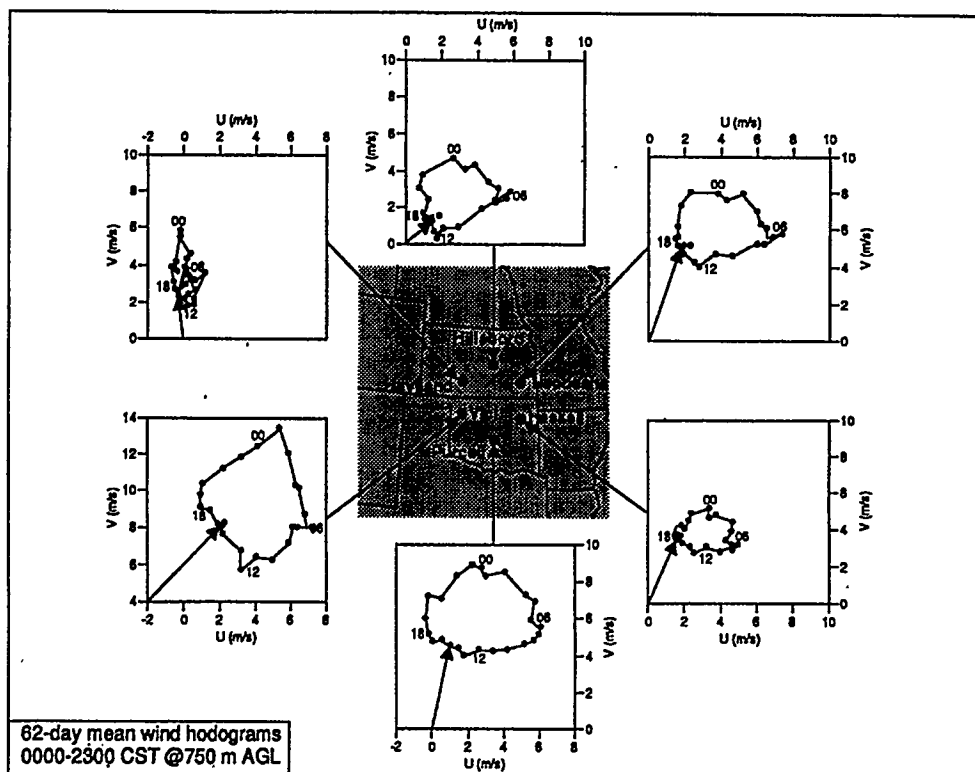


Figure 4. Hodographs of mean wind at six stations in Oklahoma and Kansas.

few percent at 18 CST to 40-60% by midnight. The LLJ frequency, then, decreases rapidly from 06 CST to noon and remains low throughout the afternoon. Notice that the daytime frequency is small but not zero. The daytime LLJ appears to be associated with passages of fronts or squall lines and is not resolved by large-scale models.

Conclusion

The climatological analyses show the ARM SGP CART site is frequently under the influence of the Great Plains LLJ in

the summer season. The mean wind direction of the LLJ is from the south-southwest and the mean wind speed is between 10 to 15 m/s. The LLJ frequency exhibits a strong diurnal oscillation, with large values during nighttime and small values during daytime. This strong southerly jet produces low-level moisture convergence and rising motions at the top of the boundary layer and helps to trigger the nocturnal development of convective clouds and precipitation in this region. The earlier phasing of the LLJ oscillation at the higher latitudes and the smooth clockwise turning of the mean wind with time at each station suggest that the inertial oscillation plays an important role in the evolution of the LLJ.

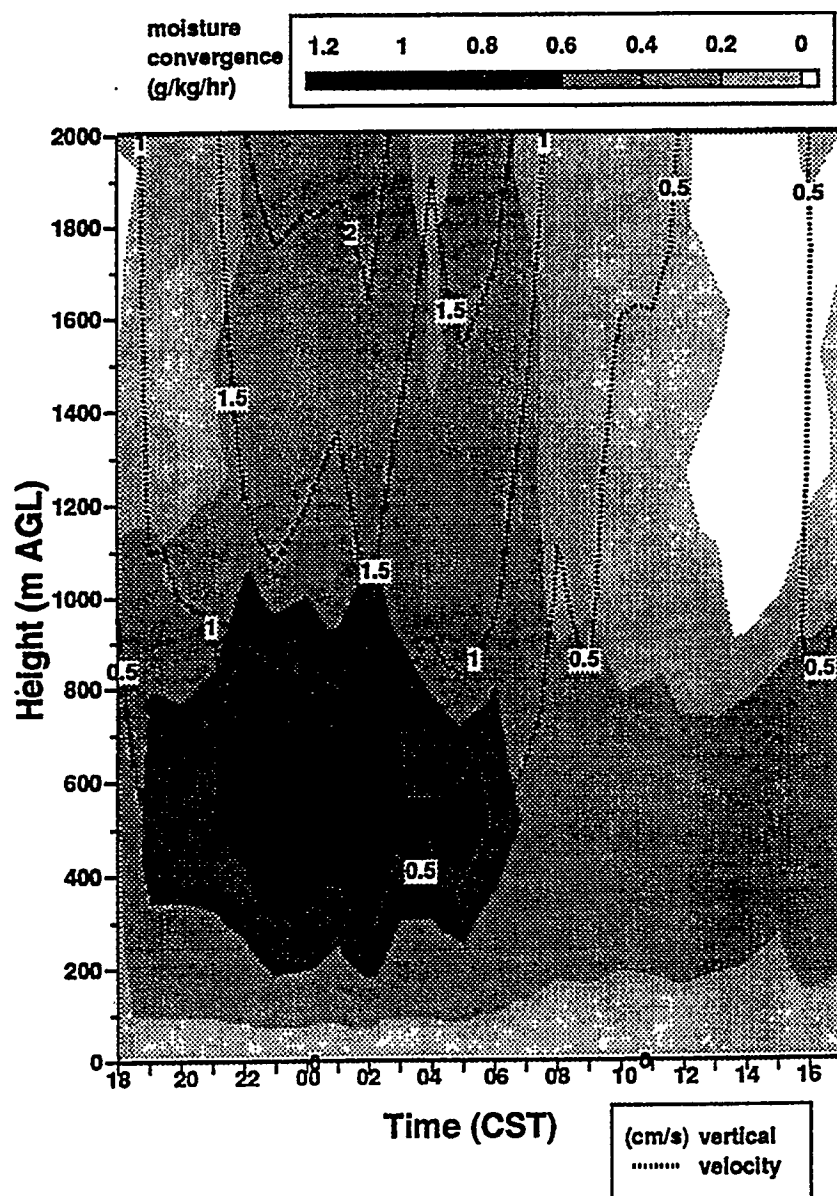


Figure 5. Two-week averaged vertical velocity and the time rate of change of the mixing ratio for all hours of the day at the CART site.

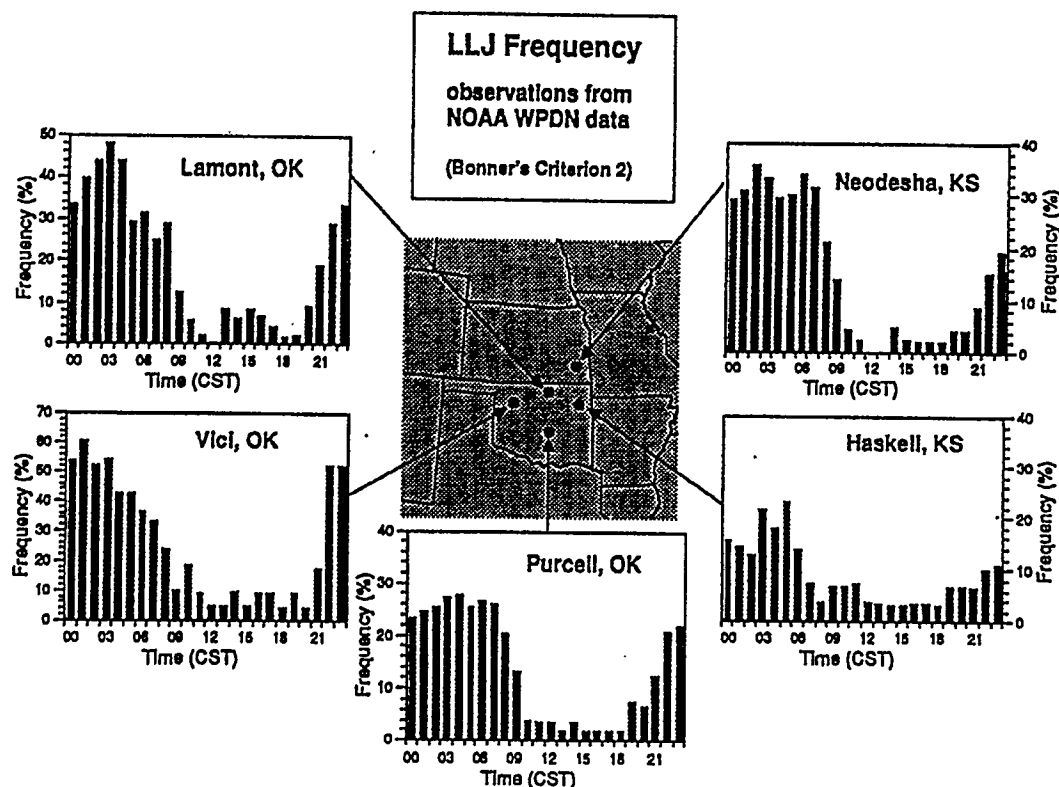


Figure 6. Diurnal variation of the LLJ frequency at selected stations.

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Initial Analyses of Surface Spectral Radiance Between Observations and Line-by-Line Calculations

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Introduction

The evaluation and improvement of radiative transfer calculations are essential to attain improved performance of general circulation models (GCMs) for climate change applications. A Quality Measurement Experiment (QME) is being conducted to analyze the spectral residuals between the downwelling longwave radiance measured by the University of Wisconsin Atmospheric Emitted Radiance Interferometer (AERI) and spectral radiance calculated by the Line-By-Line Radiative Transfer Model (LBLRTM). The three critical components of this study are 1) the assessment of the quality of the high resolution AERI measurements, 2) the assessment of the ability to define the atmospheric state in the radiating column, and 3) the evaluation of the capability of LBLRTM. Validations have been performed on spectral radiance data, obtained from April 1994 through July 1994, through the analysis of the spectral residual differences between measured and calculated radiances. These residuals are analyzed by both spectral interval and physical process. The results are archived as a function of time, enabling the retrieval of specific data and facilitating investigations of diurnal effects, seasonal effects, and longer-term trends. While the initial focus is restricted to clear-sky analyses, efforts are under way to include the effects of clouds and aerosols. Plans are well formulated for the extension of the current approach to the shortwave. An overview of the concept of the QME is described by Miller et al. (1994), and a detailed description of this study is provided by Clough et al. (1994).

Spectral Radiance Observations and Calculations

Measurements of downwelling radiance at the surface are obtained from the AERI instrument at the Southern Great Plains (SGP) Cloud and Radiation Testbed (CART) central

facility in Lamont, Oklahoma. The AERI is a zenith viewing instrument with a spectral resolution of 0.5 cm^{-1} (wavenumber value to first zero of unapodized spectrum). Radiance calculations are performed using LBLRTM (Clough et al. 1981), which includes a full water vapor continuum model (Clough et al. 1989; CKD) that has been updated to provide improved agreement with atmospheric observations (CKD_2.1). The atmospheric column is partitioned into 54 layers, chosen to accurately model the transfer of radiation to the surface. Quality controlled radiosonde data are ingested directly into the model at all reporting levels from which the temperature and water vapor column amounts are calculated. This layering also facilitates the calculation of cooling rates, which are important for other aspects of the Atmospheric Radiation Measurement (ARM) Program.

Definition of the Atmospheric State

A spectral validation is currently performed for each radiosonde release, four times per day under standard conditions and eight times daily during Intensive Observation Periods (IOP). Temperature and water vapor profiles are obtained from the radiosonde observations. A constant ozone profile has been derived from a retrieval based upon an April 23, 1994, AERI spectral radiance measurement. The characterization of atmospheric state from radiosonde measurements is a major concern, not only from the point of view of accuracy, but from the point of view of spatial and temporal displacement from the radiating column for atmospheres with variability.

Spectral Validations

Spectral residuals are obtained by subtracting LBLRTM calculated results from the measured AERI spectra. The

residuals are statistically analyzed using two approaches: 1) by spectral domain, similar to those from GCM bands; and 2) by physical process, as determined by assigning each spectral element to a physical process (e.g., spectral elements associated with H_2O lines). Figure 1a shows an AERI spectrum taken in June 1994. Channel 1, spanning $550\text{--}1800\text{ cm}^{-1}$, is the current region of principal emphasis because of its importance for climate related studies. Radiance in the saturated regions of CO_2 from $630\text{--}705\text{ cm}^{-1}$ and of H_2O from $1350\text{--}1800\text{ cm}^{-1}$ are dominated by characteristics of the Planck function associated with the atmosphere near the instrument. H_2O lines provide the dominant structure in the window regions at $800\text{--}980\text{ cm}^{-1}$ and $1080\text{--}1200\text{ cm}^{-1}$, and O_3 lines are dominant in the $980\text{--}1080\text{ cm}^{-1}$ region. The strong effects of the underlying water vapor continuum between spectral lines is seen throughout the $800\text{--}1200\text{ cm}^{-1}$ window region.

The spectral residuals for this observation, as shown in Figure 1b, are typical of those observed at the CART site within the time period of this analysis, from April 7 to July 30, 1994. Integration of the residuals over the spectral domain of Figure 1 provides a result of $1.7\text{ W}/(\text{m}^2\text{ sr})$ in zenith radiance. An estimate of the difference in flux between the observed and model values for the atmospheric window region may be obtained by multiplying the integrated residual value by five. These large residuals in the window region do not exhibit spectral behavior characteristic of errors in the water vapor continuum.

The QME products include statistical results by spectral domain as well as by physical process. As shown in the expanded plot in the lower left of Figure 1a, spectral elements in the regions denoted by the sharp spikes are

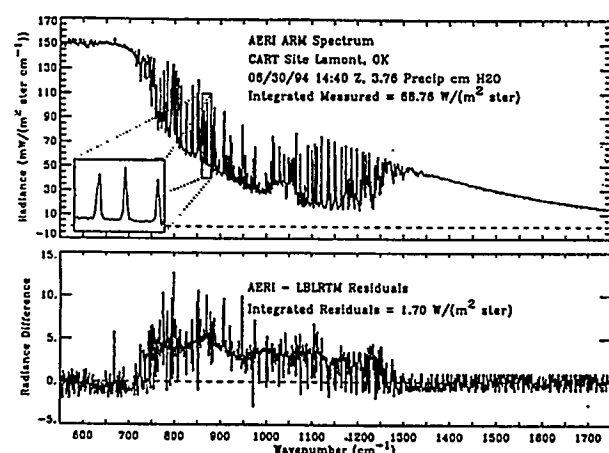


Figure 1. (a) AERI observed radiances and (b) spectral radiances between observed and LBLRTM calculated radiances.

associated with molecular lines; spectral elements in the regions between the lines, referred to as "transparent" regions, reflect the radiative effects of the water vapor continuum, aerosols and clouds.

An important objective of the QME is to control the quality of the data set. Problems associated with the loss of the liquid N_2 coolant in the instrument, resulting in the degradation of sensitivity in the detector, occurred during this time period. The affected spectra are characterized by a significant increase in measurement noise, and have been excluded from the analyses.

Initial Analyses

Figure 2 shows the equivalent brightness temperature in the channel 1 window versus the equivalent brightness temperature in the channel 2 window. The window brightness temperatures in channels 1 and 2 are determined from the average radiance across the spectral intervals $1142\text{--}1147\text{ cm}^{-1}$ and $2506\text{--}2511\text{ cm}^{-1}$, respectively. Circles refer to clear-sky cases, while stars denote non-clear cases. This information is contained in the QME cloud product platform and is currently obtained from the cloud base heights given by the Belfort ceilometer and the National Aeronautics and Space Administration Micro-Pulse Lidar (MPL), as well as the total liquid water in the line of sight of the microwave radiometer (MWR). The dotted line indicates the values at which the brightness temperatures in the channel 1 and channel 2 windows are equal. The stars along this line in Figure 2 are associated with warm, opaque clouds. Cloudy cases with lower window temperatures are associated with

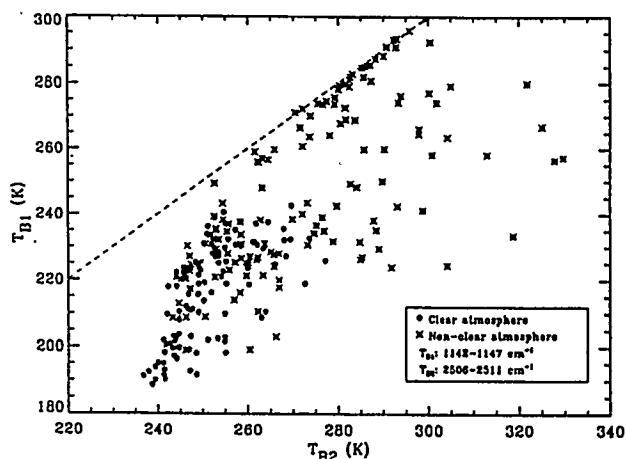


Figure 2. Channel 1 window brightness temperatures as a function of channel 2 window brightness temperatures for clear and non-clear atmospheres.

high, cold, thin clouds. The higher brightness temperatures in the channel 2 window with a high degree of variability are caused by solar scattering from clouds and aerosols. Efforts are currently under way to assess the extent to which solar radiation may be entering the AERI light pipe directly.

Analyses of diurnal variation can be explored, as shown in Figure 3, where integrated spectral differences are analyzed as a function of time of day in Greenwich Mean Time (GMT). There is a difference of six hours between GMT and Lamont local time during daylight savings time. Only clear-sky cases are plotted, distinguished by month. The columns indicate the daily radiosonde launch times, with 3 Z, 8 Z, 11 Z, and 23 Z reflecting the additional radiosonde launches during the April and July 1994 IOPs. The systematic difference from day to night on the order of $1.5 \text{ W}/(\text{m}^2 \text{ sr})$ is currently unexplained.

A plot of integrated spectral residuals by physical process as a function of water vapor amounts is shown in Figure 4. Only clear-sky cases are shown and cover a wide range of precipitable water, from 0.5 to 4.0 precipitable centimeters. The dominant processes in this spectral region are those associated with absorption in the "transparent" regions between the lines, which include the continuum, aerosols and thin cloud, and the spectral elements associated with water vapor lines and ozone lines. The sum of these differences in Figure 4 reflect 98% of the total difference in channel 1. Although a constant ozone profile is used, the differences associated with the ozone region of Figure 4c, on the order of $0.3 \text{ W}/(\text{m}^2 \text{ sr})$, are remarkably good considering the seasonal variation of O_3 . It is of interest that the differences associated with H_2O lines, as shown in

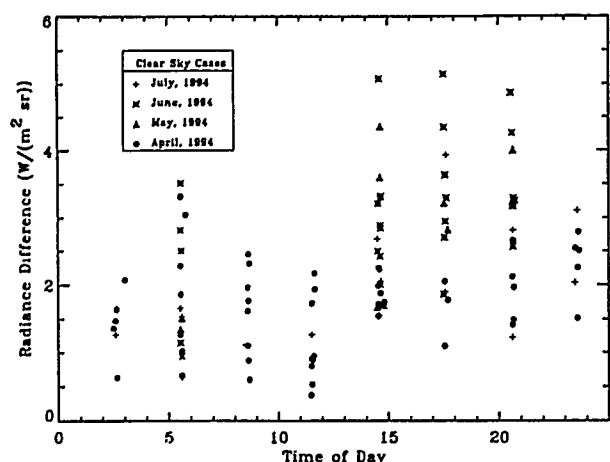


Figure 3. Channel 1 integrated radiance differences as a function of time of day.

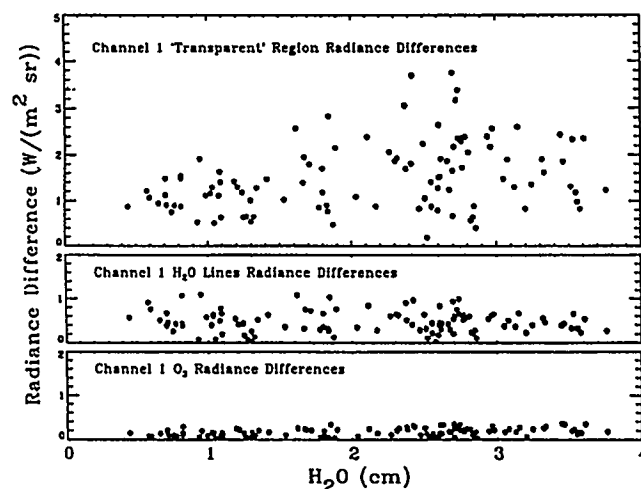


Figure 4. Integrated spectral residuals by physical process as a function of radiosonde observed water vapor amounts.

Figure 4b, are not correlated with water vapor, though there is an apparent small linear correlation between the differences associated with the transport regions with the water vapor column as shown in Figure 4a. Most importantly, the lack of a quadratic correlation between the differences associated with the transparent regions and water vapor indicates that these differences are not attributable to the water vapor continuum. A linear correlation is more consistent with the hygroscopic growth of aerosols and the consequent increase in the aerosol radiative effects.

Conclusions

The AERI/LBLRTM QME has been effective in assessing problems associated with the measurement of downwelling spectral radiance, the characterization of the atmosphere in the radiating column, and the line-by-line calculations. Issues including the loss of liquid N_2 fill in the AERI dewar and the switch from nominal mode to research mode in the processing of the radiosonde data have been identified and addressed. There are two important conclusions that can be drawn from this initial analysis of the QME products: 1) the residuals in the important channel 1 window for the clear sky are anomalously large corresponding to $\sim 10 \text{ W}/\text{m}^2$ in flux and 2) the specification of atmospheric state for the radiating column to the requisite accuracy is a difficult but tractable task.

Clear-sky analyses show that the integrated radiance differences for validations of AERI measurements taken at the ARM SGP Central Facility are significantly higher than

spectral validation between AERI measurement and the LBLRTM calculations from Coffeyville, Kansas, during the Spectral Radiance Experiment (SPECTRE) in November and December 1991. The integrated spectral radiance differences for SPECTRE are on the order of $0.5 \text{ W/(m}^2 \text{ sr)}$ while those from ARM are $2.0 \text{ W/(m}^2 \text{ sr)}$. We have established that this difference in the ARM validations is not attributable to the water vapor continuum. Radiative effects of aerosols or thin clouds may play a role, but the extinctions required appear to be too large. Another possible contributor to the residuals is the AERI calibration at low radiances. Inclusion in the calculation of the effects of HNO_3 and improvements in the halocarbon mixing ratios are expected to provide some reduction in the residuals, but nothing approaching the required amount.

With respect to the task of adequately specifying the atmospheric state, the anticipated availability of Raman lidar and Radio Acoustic Sounding System (RASS) observations is expected to provide improvement in the water vapor and temperature profiles, respectively, particularly with respect to the spatial and temporal sampling issue. Absolute calibration of both instruments remains an important and difficult task. It is clear that more information on aerosols is essential; at an absolute minimum, a value for the extinction in the visible is required.

This initial analysis strongly supports the QME as a strategy to assess our current capability to measure and model longwave radiation in the study of climate change. Before the approach can become fully effective, an explanation for the ubiquitous and anomalous residuals in the window regions must be resolved. Work is under way toward the inclusion of cloud and aerosol effects in the model calculations and in the calculation of atmospheric cooling rates from $10\text{--}3020 \text{ cm}^{-1}$ for single column model applications. In addition, analyses of validations in the microwave region between the MWR and LBLRTM are currently being implemented. Future plans include the assessment of potential improvement in atmospheric state measurement from the RASS and Raman Lidar, and the extension of the analysis to solar spectra, as measured at the central facility by the University of Denver Solar Radiance Transmission

Interferometer and the Absolute Solar Transmission Interferometer, and to the shortwave spectral region.

Acknowledgments

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Cloud Classification Using Whole-Sky Imager Data

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Introduction

Clouds are one of the most important moderators of the earth radiation budget and one of the least understood. The effect that clouds have on the reflection and absorption of solar and terrestrial radiation is strongly influenced by their shape, size, and composition. Physically accurate parameterization of clouds is necessary for any general circulation model (GCM) to yield meaningful results. The work presented here is part of a larger project that is aimed at producing realistic three-dimensional (3D) volume renderings of cloud scenes based on measured data from real cloud scenes. These renderings will provide the important shape information for parameterizing GCMs. The specific goal of the current study is to develop an algorithm that automatically classifies (by cloud type) the clouds observed in the scene. This information will assist the volume rendering program in determining the shape of the cloud.

Much work has been done on cloud classification using multispectral satellite images (e.g., Ebert 1992; Garand 1986; Inoue 1987; Lamei et al. 1993; Lee et al. 1990; Smotroff 1993; Yool and Brandley 1992). Most of these references use some kind of texture measure to distinguish the different cloud types and some also use topological features (such as cloud/sky connectivity or total number of clouds). A wide variety of classification methods has been used, including neural networks, various types of clustering, and thresholding. The work presented here uses binary decision trees to distinguish the different cloud types based on cloud feature vectors.

The images used in this study were obtained from two whole-sky imagers (WSIs) located at White Sands, New Mexico, during the months of July, September, and October 1989 and May 1992. These ground-based imagers are spaced approximately 5 km apart and, together, provide a stereo view of the sky. The data acquired from these imaging systems can be used in a wide variety of applications including cloud base height determination and cloud-type occurrence statistics. These applications can be augmented by automatic determination of the type of cloud present in a cloud scene.

WSI data present some unique challenges to cloud classification for several reasons. First, the image is not a flat representation of the cloud scene. Instead, equal distances in the image are equal angles in the scene. Therefore, while the angular resolution of the data ($\approx 1/3^\circ$) is constant, the length resolution decreases from the center to the edge of the image. Also, due to the perspective of ground observations, the center of the image views the bottom of the clouds but the image edge views either the bottom (for clouds of large spatial extent) or the side (for modular or broken clouds). Another challenging aspect of the WSI data is the relatively limited field of view. Compared with satellite data, the WSI cloud scene is a very local one. Many of the clouds extend beyond the edge of the image and, therefore, very few clouds are seen in their entirety. This precludes such features as perimeter measures, shape measures, etc., from being used in the classification.

Classification Method

The objective of this work is to automatically classify the different types of clouds observed in WSI data. To do this we use binary decision trees (Breiman et al. 1984) to classify each pixel in the cloud scene as either clear sky or one of four cloud types: altocumulus, cirrus, cumulus, or stratus. Binary decision tree algorithms use a set of training data, for which the cloud class is known, and generate a hierarchy of thresholds based on a series of measured features. For each point in the training sample, n different features are measured. These measurements are placed in an n -dimensional space, and successive splits (which are equivalent to thresholds on individual feature values) are used to separate the points into different regions or nodes. At every level, each region is individually analyzed to find the optimal split (i.e., the split which most reduces some measure of the node impurity). In the resulting tree structure, each node has an associated misclassification rate estimate which can be used to evaluate the relative performance of the tree and also the confidence of the final classification.

The features described here are all measured for each pixel in the WSI data. Most of the features are computed as an average over a neighborhood around each pixel. The resulting binary decision tree, then, actually classifies individual

pixels as belonging to one of four cloud classes or the clear sky class. This result can be generalized to give the most likely class to which each cloud or cloud image belongs.

The features used in this study can be grouped into three categories: texture measures, position information, and pixel brightness. One of the texture measures is the standard deviation of the image brightness computed over a neighborhood around each pixel. The remaining texture measures are computed using the LAWS kernels (Laws 1980). These 5×5 matrix kernels are used to quantify the amount of 25 different types and orientations of texture in the cloud scenes and are derived by taking the outer product of combinations of two of the following five vectors: $L5 = [1 \ 4 \ 6 \ 4 \ 1]$, $E5 = [-1 \ -2 \ 0 \ 2 \ 1]$, $S5 = [-1 \ 0 \ 2 \ 0 \ -1]$, $W5 = [-1 \ 2 \ 0 \ -2 \ 1]$, and $R5 = [1 \ -4 \ 6 \ -4 \ 1]$. The letter names for the vectors refer to the type of texture that each vector is sensitive to, namely Level, Edge, Spot, Wave, and Ripple, respectively. For example, the kernel $L5E5$ is the outer product of the vectors $L5$ and $E5$, and $E5L5$ is its transpose, the outer product of the vectors $E5$ and $L5$.

The method used to obtain a measure of the texture energy in the neighborhood of each pixel closely follows that found in Laws (1980). First, each cloud scene is convolved with one of the 5×5 texture kernels. Next, the absolute value of the result is averaged over a neighborhood around each pixel to give a measure of the texture energy at each point. The nature of the WSI camera systems gives rise to two important distance measures in the acquired images. First is the distance from the center of the image, which is equivalent to the zenith angle. Since the linear (not angular) resolution of the image is degraded as the zenith angle increases, the texture of the clouds will also change with increasing zenith angle. Second is the difference between the sun angle and viewing angle, which we approximate as the distance from the center of the occulor. These two distance measures are also included as features used in the classification. Finally, the pixel brightness of the acquired image is also included as a feature.

The texture measures, as outlined above, only use information from a small neighborhood relative to the sizes of the clouds. In order to include the available large-scale information in the classification, the resolution and sample density of the data are recursively reduced to form a pyramid-type structure following Burt (1984). In this pyramid structure, each successive level represents low-pass filtered copies of the original image with an additional decrease in sampling density and a corresponding decrease in physical size (each level being one-half as big as the preceding level). The same sized kernels and neighborhoods (in pixels) are used to compute the texture energy at each

level with the result that each level contains information collected over a larger extent than the previous level. After the texture measures are computed on the reduced images, they are expanded by interpolation (Burt 1984) to the original size so that information from different spatial resolution levels can be used simultaneously at each point. Four levels were used for this work with the original image being level 0, with a diameter of about 450 pixels, and level 3 being eight times smaller in size than level 0, with a diameter of about 56 pixels.

Of the features outlined above, there is no way to determine *a priori* which ones are more important to the final classification and which ones can be left out without introducing errors. Furthermore, it would be very computationally expensive to grow the decision tree using all the features and let the algorithm choose the best ones. Therefore, an iterative process was used to estimate which of the above features allowed the optimal tree to be grown. First, the training data were classified using each feature independently and the resulting resubstitution error rates compared (the resubstitution error rate is the error incurred when the data that were used to grow the decision tree are classified by the decision tree). This error rate can be non-zero because in some instances the training data do not warrant the amount of splitting required to produce a zero error rate. The five features that gave the lowest error rates were then passed on to the next step. The training data were classified using one of the best five features from the previous step and all the other features, taking one at a time so that each tree was grown using two features. The five trees with the lowest error rates were then passed on to step three. To each pair of features from step two, all other features were added one at a time and a tree with three features grown. This procedure was repeated until the nine most important features were discovered. This is actually an estimate of the most important features. To determine the absolute best features, an exhaustive search would have to be made where every combination of nine features would be used to classify the training data and the resubstitution error rates compared for all trees. This is an enormous task for even a small number of features.

Results

Table 1 lists the number of images and pixels for each cloud type used in the training data set. The method outlined above was used to determine the nine most important features for classifying the training data. The resubstitution error rate of the resulting decision tree was 6.4%. The nine features used for the classification, in order from most to least important, are

Table 1. The composition and size of the training data and test data sets.

Cloud Class	Number of Training Images	Number of Training Points	Number of Test Images	Number of Test Points
Alto cumulus	3	80,197	1	32,264
Cirrus	5	153,760	1	52,347
Cumulus	5	114,955	2	75,235
Stratus	5	187,394	2	152,595
Clear sky	16	469,152	4	125,606

1. the LAWS feature L5L5 for level 2
2. the distance from the center of the occulor
3. the distance from the center of the image
4. the local standard deviation at level 0
5. the local standard deviation at level 1
6. the pixel brightness
7. the LAWS feature L5E5 for level 3
8. the LAWS feature L5R5 for level 3
9. the LAWS feature E5L5 for level 3

Several things should be noted from this list. First, the four LAWS texture measures included are for levels 2 and 3. This implies that these important textures are more large-scale textures, probably inter-cloud rather than intra-cloud. Also note that the transpose pair L5E5 and E5L5 are included, indicating that this particular texture occurs with more than one orientation. Second, the local standard deviation features are for levels 0 and 1, indicating that intra-cloud texture is also important for the classification. Finally, both distance measures are very important features, indicating the strong dependence of cloud appearance on position within the image.

The resubstitution error rate is a measure of how well the decision tree classifies the training data. But a more strict performance measure of the decision tree is the error rate resulting from classifying data not included in the training set. (Table 1 lists the number of images and pixels for each cloud type used in this test data set.) This error rate is the misclassification error rate, which is defined as the

percentage of class I points that are misclassified as class j , with $j \neq I$. The misclassification error rate would be zero for a perfect classification. It is also instructive to generate a confusion matrix C_{ij} which gives the fraction of class I points that are classified as class j . This matrix would be diagonal for a perfect classification and shows which classes are confused most often.

Since the classification is performed on the individual pixels in each test image, the output from the decision tree algorithm is a very noisy image. Several methods can be used to clean up the raw output and produce a more usable and informative result. For the present, a k-nearest neighbor filter is used to remove the noise in the raw classified images. The class of each point is changed to the most probable class in a 31×31 neighborhood around that point. This has the effect of absorbing small islands of one class into the larger regions of the surrounding class. Visual inspection of both the raw classified and the filtered images shows that the basic trends are maintained but the noise is greatly reduced. Figure 1 shows one example of the classification result with Part a showing the cloud scene, Part b showing the classification result before filtering, and Part c showing the result of filtering with the k-nearest neighbor filter. In Parts b and c, black denotes clear sky pixels, dark gray denotes cirrus cloud pixels, medium dark gray denotes stratus cloud pixels, medium light gray denotes alto cumulus cloud pixels, and light gray denotes cumulus cloud pixels. Table 2 gives the confusion matrix for the filtered results which has an overall misclassification rate of 39%.

Conclusions/Discussion

The decision tree grown using the training data has a very low resubstitution error rate (6.4%), indicating that the features used to perform the classification represent the

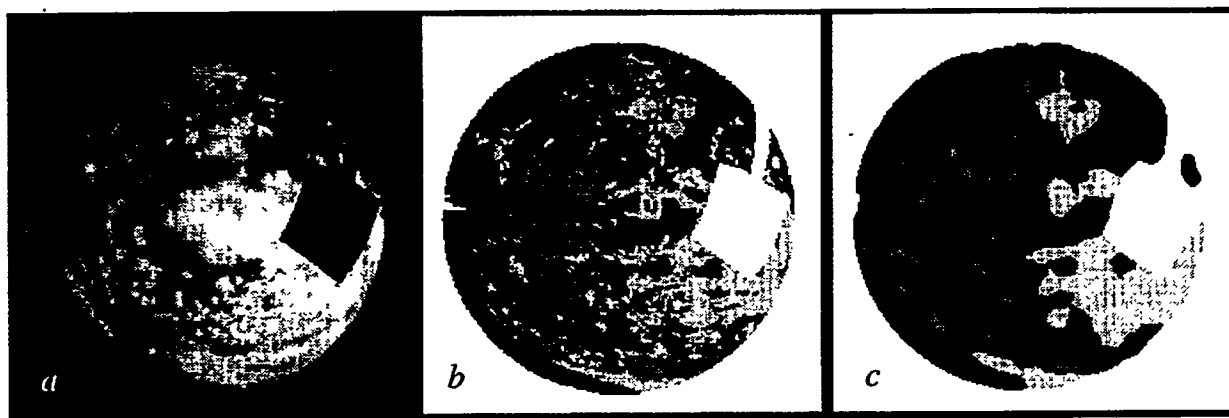


Figure 1. An example of the classification results for one of the test data images. Part a shows the WSI cloud scene. Part b shows the classification result before filtering. Part c shows the classification result after processing with the k-nearest neighbor filter. In Parts b and c, the five gray levels, starting with black and increasing to light gray, correspond to pixels classified as clear sky, cirrus cloud, stratus cloud, altocumulus cloud, and cumulus cloud, respectively.

Table 2. Confusion matrix for the test data after filtering with the k-nearest neighbor filter.					
True Class	(%) altocumulus	(%) cirrus	(%) cumulus	(%) stratus	(%) clear sky
Altocumulus	43	19	12	10	17
Cirrus	7	35	21	24	12
Cumulus	8	14	54	14	11
Stratus	4	12	12	46	26
Clear sky	1	0	1	0	98
Overall misclassification rate = 39%					

training data well. The misclassification error rate is higher (39%) for several reasons. First, there is the possibility that the training data are not representative of the test data. This difficulty can be overcome by including more samples in the training data, thus supplying the tree growing algorithm with information about a larger range of clouds in each cloud class. With the immense variability that exists in clouds, this is almost certainly a contributing factor in this work and also one of the hardest to overcome. Second, when the test data classification results are analyzed, several common errors that contribute to misclassified pixels are easily recognized. The two most obvious are problems in correctly determining the

class of pixels near the edge of the image and near the occulter. This is not an unexpected result since the resolution is the poorest near the edge of the image, which greatly reduces the sensitivity of the texture measures, and the brightness is generally the highest near the occulter, which can tend to make different cloud types appear similar. Another factor contributing to the high misclassification error rate is related to the use of the k-nearest neighbor filter. Passing this filter over the classified images does decrease the misclassification error rate (for the test data, the error rate was reduced from 45% to 39%), but it does so without any knowledge specific to the geometry or physics of cloud fields. Another method that

does use some basic rules about typical cloud sizes, cloud occurrence statistics etc., would probably give a more dramatic reduction in error rate. This assertion is motivated by the fact that, in the test data classification results, the correct cloud class is present somewhere within the region occupied by almost every cloud. If this correct class could be identified by some rule-based decision algorithm, then another filtering operation with the correct cloud class preferentially weighted could be used with an associated decrease in the error rate.

This study has succeeded in developing an algorithm that can classify pixels in WSI data according to cloud type through the use of binary decision trees. The classification is fairly accurate when tested on data other than those used for training, and analysis of the results has suggested several ways of improving the overall accuracy of the classification. Future work on this project will be focused on two main areas. First, we will assemble a training data set that is more representative of the cloud scenes collected by the WSIs and thereby improve the accuracy of the initial classification step. Second, we will develop a more physically based process of determining the actual cloud class from the output of the decision tree algorithm and thereby improve the accuracy of the second classification step.

Acknowledgments

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Broken-Cloud Enhancement of Solar Radiation Absorption

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Abstract

Two papers recently published in *Science* have shown that there is more absorption of solar radiation than estimated by current atmospheric general circulation models (GCMs) and that the discrepancy is associated with cloudy scenes. We have devised a simple model which explains this as an artifact of stochastic radiative transport.

We first give a heuristic description, unencumbered by mathematical detail. Consider a simple case with clouds distributed at random within a single level whose upper and lower boundaries are fixed. The solar zenith angle is small to moderate; this is therefore an energetically important case. Fix the average areal liquid water content of the cloud layer, and take the statistics of the cloud distribution to be homogeneous within the layer. Furthermore, assume that all the clouds in the layer have the same liquid water content, constant throughout the cloud, and that apart from their droplet content they are identical to the surrounding clear sky. Let the clouds occupy on the average a fraction p_{cl} of the volume of the cloudy layer, and let them have a prescribed distribution of sizes about some mean. This is not a fractal distribution, because it has a scale. Cloud shape is unimportant so long as cloud aspect ratios are not far from unity. Take the single-scattering albedo to be unity for the droplets in the clouds. All of the absorption is due to atmospheric gases, so the absorption coefficient at a point is the same for cloud and clear sky. Absorption by droplets is less than a 10% effect in the numerical stochastic radiation calculations described below, so it is reasonable to neglect it at this level of idealization.

Now, consider the effect of a variation in the scale of the size distribution of the clouds on the spatial distribution of solar radiation. There are two fixed sizes in the problem: the photon mean free path and the layer thickness. When the cloud size and distribution scale is much smaller than a photon mean free path, the effect of the clouds is as if the droplets that compose them were uniformly distributed. Neither cloud fraction nor cloud size matters, and the problem reduces to a plane-parallel one. Scattering is maximized, because the self-shielding is minimized; the photon path length is also maximized. Absorption is maximized because it is monotonically related to path length.

As the scale size becomes larger, inhomogeneities occur in the scene, and the top of the atmosphere albedo drops. Eventually thin spots and holes develop in the cloud layer; transmission through it increases as a result. But absorption does not respond so rapidly with increased cloud size as does transmission. Bright clouds are seen from the ground in almost all of the 2π solid angle regardless of the observer's exact ground location, an indication that the cloud and below-cloud volume is still full of photon tracks. Imagine a clear column cut into the clouds, such that sunlight can stream through. The additional light that reaches the ground through this column does not have the increase in path length that results from multiple scattering, but absorption in the column is still larger than calculated by a clear-sky model. This is because the column is full of laterally-traveling photons which have emerged from the surrounding clouds. An increase in the scale of the size distribution causes an increase in intermittency of occurrence (brokenness) of cloud in the fixed thickness layer assumed here. Thus absorption, which is dependent on path length, remains

near its (maximum) small-cloud value as cloud size increases, even as transmission of the solar beam increases.

Very large clouds become so localized that they lose effectiveness as light sources for each other (and for ground observers). A fraction $(1-p_{\text{cld}})$ of observers would see no clouds, while a fraction p_{cld} would be under a cloud extending to their horizons. This is the situation accurately described by the fractional cloud cover (FCC) model, a weighted combination of two plane-parallel calculations, one clear and one overcast. This arrangement of liquid water minimizes the effect of scattering, because many of the droplets are hidden deep in the interior of the large cloud, and thus shielded from the solar radiation. The incloud scattering (and increase in path length) that takes place is confined to the surface of the cloud. The cloud-cloud scattering is small, because the solid angle subtended by other clouds is small, vanishing in the limit of very large but very rare clouds. The absorption thus falls as cloud size increases. Once the cloud-cloud scattering becomes negligible, FCC describes the situation adequately; there are no further cloud size effects.

Thus, in this simplistic model the absorption and albedo vary from maximum to minimum, but the absorption varies more slowly as the cloud scale size increases.

Plane parallel theory holds for small cloud size scales, and FCC holds for large ones. For intermediate cloud sizes, absorption is anomalously enhanced relative to the predictions of either of those models, but is captured by a stochastic model.

Anomalous absorption is manifested largely in the near infra-red part of the solar spectrum because, although the relative increase in path length occurs for all wavelengths, there is very little absorption in the visible solar spectrum.

Absorption responds more slowly than does albedo as inhomogeneities appear in a cloud layer. This is predicted by the simple model given above and confirmed by numerical calculations. We give a translation of the result into the observational quantities of Cess et al. (1995) and Ramanathan et al. (1995).

Figure 1 plots the coefficient β derived from our calculations against cloud fraction and cloud size. It shows the existence of the decrease implied by our model. The values quoted by Cess et al. (1995) for β are about 0.8 for the GCMs and about 0.6 for the observations. We find values below 0.6 for the case of small dense widely scattered clouds (the region of low cloud volume fraction and 1 km sizes). We find $\beta = 0.7$ for small or large cloud sizes, rather

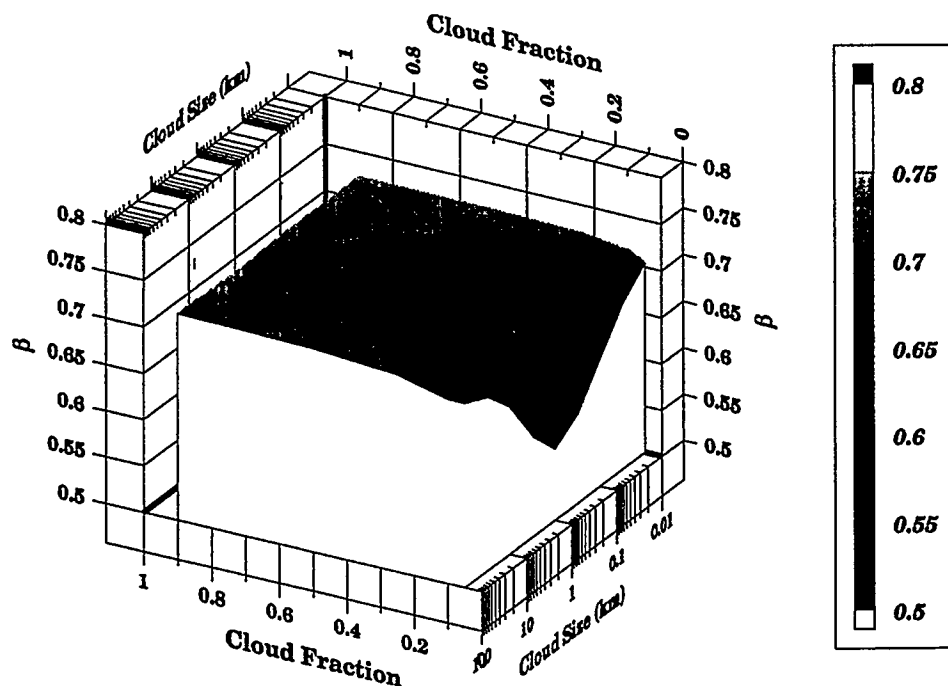


Figure 1. Cess coefficient, β , for a range of cloud sizes and fractional coverage.

than the expected 0.8, which may be due to the parameters we used in the example. The net result is that the change in β that we find, due to stochastic effects (about 0.1) is half that reported by Cess et al. (1995).

The increase in f_s , the measure used by Ramanathan et al. (1995), is shown in Figure 2. This figure summarizes the results of many computer runs. The same conditions that minimize β produce values of f_s greater than 1.3, as opposed to the 1.1 we find when we neglect broken cloud effects by taking the clouds to be very small or very large. Again, our calculation produces about half the reported effect.

The numerical example presented here does not change β or f_s to the extent reported, but it does illustrate the existence of a one-sided bias toward smaller β (larger f_s) and accounts for a significant part of the discrepancy between observations and conventional models. To ascertain the magnitude of this contribution, more research needs to be done using observed profiles and taking long-term averages.

This paper has described what we claim to be a general trend tending to support the anomalous absorption

observations, but has provided numerical results for only one atmospheric profile, only one solar angle, only one cloud liquid water loading, only one surface albedo, and (for the most part) only one cloud layer at only one height and thickness. Future work should explore the variation of the effect with these parameters and provide a weighted estimate of the net result, allowing a direct comparison with the observations.

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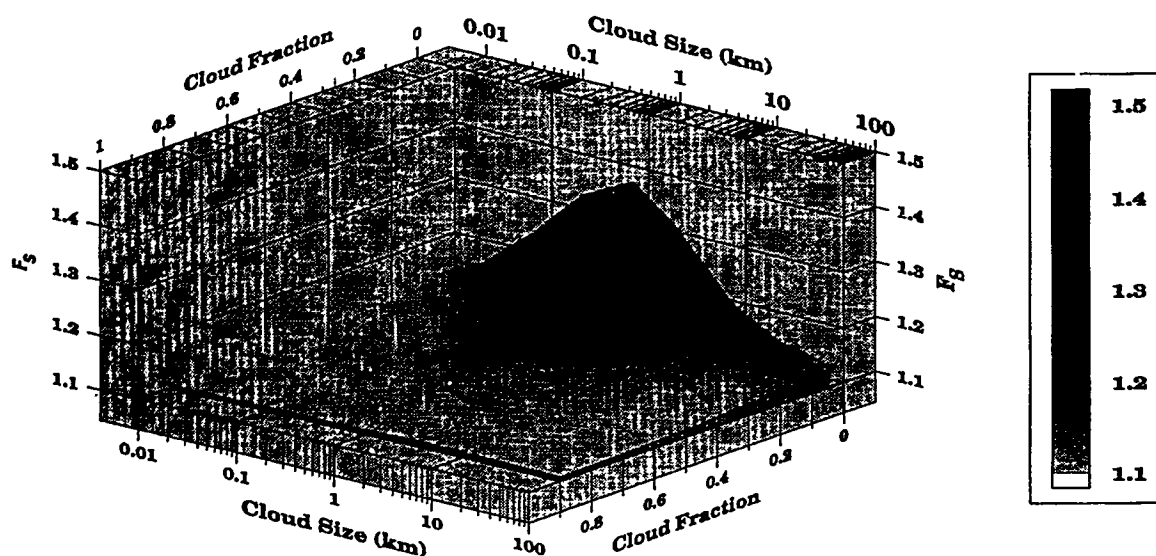


Figure 2. Ramanathan coefficient, f_s , for a range of cloud sizes and fractional coverage.

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Toward Improving the Representation of Anvil Cirrus Effects in General Circulation Models (GCMs)

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Introduction

Cloud-radiation feedback has been identified as the most important factor limiting general circulation models (GCMs) from further progress in climate change research (Cess et al. 1989). It is also regarded as a major uncertainty in estimating the impact of greenhouse gases on climate simulations (Schlesinger and Mitchell 1987). Many GCMs show high sensitivity to the treatment of clouds and cloud radiative properties (e.g., Randall et al. 1989; Mitchell et al. 1989). Therefore, a better understanding of cloud-radiation feedback on the large-scale environment is absolutely essential to improve the representation of cloud processes in GCMs.

To this end, a cloud model with enhanced model physics is used to study the impact of cloud-radiation interactions on mesoscale convective systems (MCSs). Case studies representing a variety of convective systems are important to generalize the overall effects of anvil clouds on the large-scale environment. Our primary interest is limited to MCSs in an environment with substantial wind shear, such as squall-line systems, because they have longer lifetimes and wider coverage to impact the earth radiation budget and climate.

Recent studies (Chin 1994; Chin et al. 1995 and many others) indicated that the vertical shear of the environmental wind plays an important role in the formation of the anvil cloud through the tilting of MCSs. However, this process has not been represented in GCMs. A detailed investigation on the formation of anvil clouds and their relationship to cumulus portions of MCSs would help develop a better cloud parameterization for use in GCMs.

Comparisons between simulated midlatitude (Oklahoma) and tropical (Global Atmospheric Research Program Atlantic Tropical Experiment) squall-line systems are presented in this study. Two important issues are addressed through these comparisons. First, what factors cause the

differences between midlatitude and tropical anvil clouds? Second, do these differences have climatic implications to improve our climate forecasting ability?

Model and Initialization

The model used is an extension of Chin and Ogura's (1989) two-dimensional cloud model, which is non-hydrostatic and fully compressible. The major improvements include ice microphysics and radiation transfer schemes for longwave (LW) and shortwave (SW). The modified parameterizations of ice microphysics and radiation can simulate midlatitude and tropical squall-line systems with prominent anvils and realistic mesoscale structures. The radiation schemes can distinguish the impacts of hydrometeor phase, size, and shape on cloud optical properties.

Refer to Chin (1994) and Chin et al. (1995) for the details of model physics, grid structures, and initial conditions of both midlatitude and tropical cases. Based on Weisman and Klemp (1984), the bulk Richardson numbers of midlatitude and tropical soundings (74 and 52, respectively) favor the development of multicellular storms.

The model is initialized by a warm, moist bubble and a horizontally homogeneous sounding. However, the tropical case also includes the large-scale ascent. Our sensitivity tests indicate that an organized tropical oceanic squall-line system cannot develop without the imposed large-scale ascent.

Results

To estimate the degradation in GCM accuracy caused by omitting individual processes, we conducted a series of simulations for midlatitude broken-line and tropical oceanic squall-line systems. These simulations use different combinations of radiation and microphysics complexity, ranging from no radiation and only liquid microphysics to both LW and SW radiation with full liquid and ice

microphysics. To validate our results, we compared the simulations with observations (Smull and Houze 1987; Houze 1977).

Impact on Dynamic and Microphysical Structures

The general patterns of the dynamic and microphysical structures of the simulated midlatitude and tropical squall-line systems (Figures 1 and 2) were very similar to the observed features, that include the upshear tilting of convective cores, the broad stratiform precipitation region, the strong slant rear-to-front inflow with the outflow above and below, a bright melting band, and a well-defined transition zone in the radar reflectivity.

Our results show that both ice-phase microphysics and radiation have little influence on the multicellular character of the modeled midlatitude and tropical squall-line systems. Although both simulated systems exhibit strong similarities in the dynamic and microphysical structures, different environmental conditions, such as the presence or absence of the jet wind profile and thermodynamical stability, still

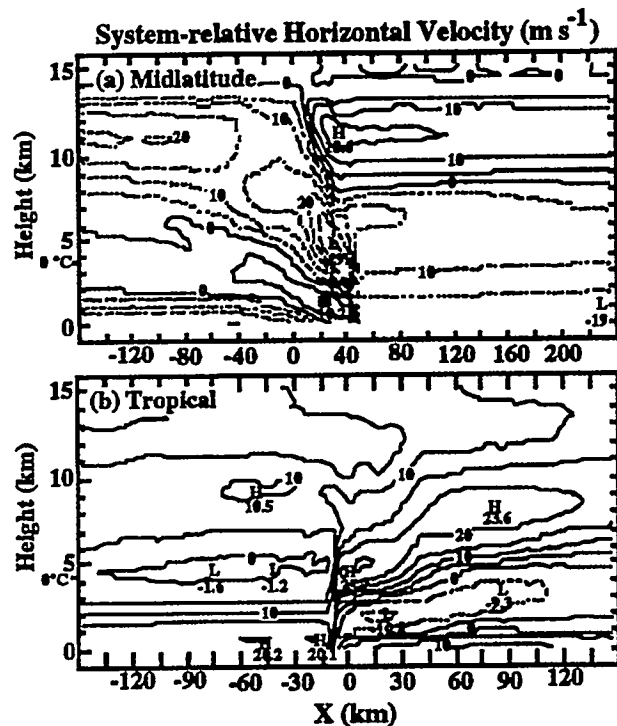


Figure 1. Temporally averaged cross-sections of system-relative horizontal velocities at the mature stage of the simulated squall-line systems: (a) midlatitude case and (b) tropical case.

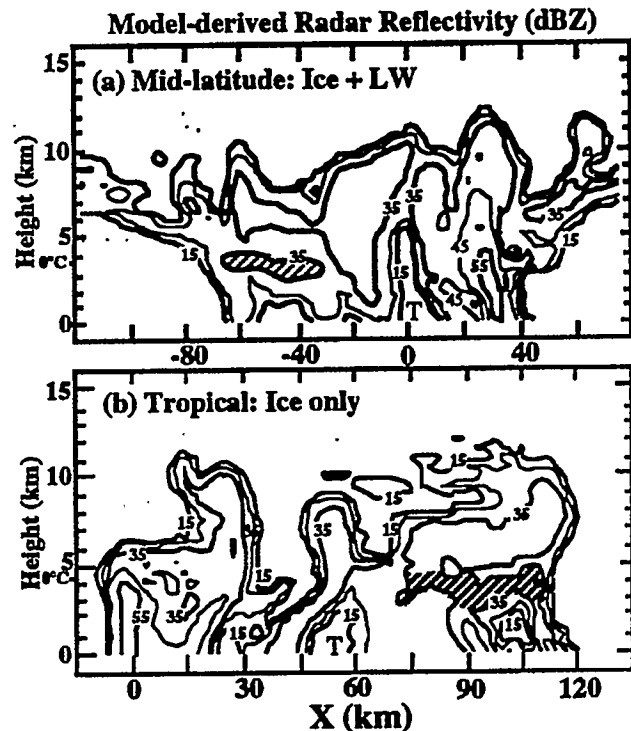


Figure 2. Instantaneous model-derived radar reflectivities near the late mature stage of simulated squall-line systems. Shaded area represents the bright melting band and T marks the location of the transition zone: (a) midlatitude case and (b) tropical case.

cause substantial differences. These differences include a weaker upper-level leading and stronger upper-level trailing outflow, and a stronger and broader trailing anvil cloud in the tropical squall-line system.

Impact on Moisture and Water Budget

To assess the contribution of moisture (i.e., water vapor) transport by deep convection to the formation of simulated anvil clouds, the moisture budget is calculated for both simulated tropical and midlatitude squall-line systems (Figure 3). For the direct comparison, the values shown in the tropical and midlatitude moisture budgets are normalized by total surface precipitation (P_0) of the tropical case.

Interestingly, moisture transport from the lower tropical and midlatitude convective regions (below 3.6 km) is about the same; this is because the very moist low-level tropical atmosphere compensates for the weaker convective instability in the tropics. In addition, the weaker and lower deep

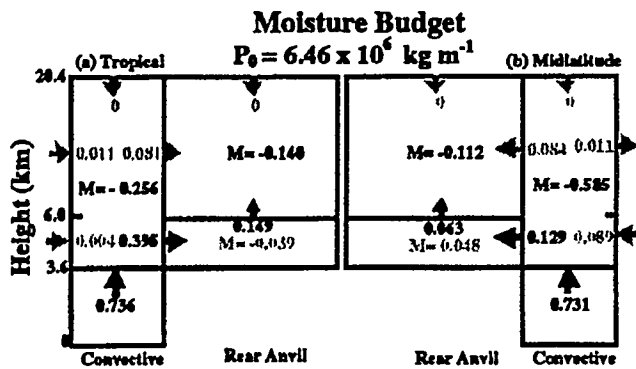


Figure 3. The moisture budgets at the mature stage of the simulated squall-line systems: (a) tropical case and (b) midlatitude case. All quantities are normalized by total surface precipitation of the tropical control run (P_0). M represents the net depletion by microphysical processes. Small arrows represent horizontal and vertical moisture transport.

convection in the tropical case depletes much less moisture through net condensation/deposition at the upper levels of the convective region (above 3.6 km). As a result, more moisture at the upper levels is transported rearward to form the anvil cloud in the tropical case. In addition, the larger upper-level moisture depletion by condensation/deposition and stronger vertical moisture transport by mesoscale lifting at 6 km in the tropical anvil region are an indication of stronger mesoscale ascent associated with the tilting structure of the simulated squall-line system. Similar features are also found in moisture transport at the lower level for the mesoscale descent. As a whole, these moisture budgets indicate that tropical deep convection serves as a more efficient engine pumping low-level moisture into the upper-level anvil cloud than its midlatitude counterpart.

The water (i.e., condensate) budget for the midlatitude case indicates that hydrometeor transport by deep convection is the primary contributor to the water source of the anvil cloud at the mature stage (66%; defined as the ratio of horizontal transport to the sum of horizontal transport and microphysical production). The rest of the water budget in the midlatitude anvil (34%) is contributed by the mesoscale lifting associated with the tilting convective system in the sheared environment (Figure 4a). In contrast, deep convection contributes only 40% to the water budget of the tropical anvil and becomes a secondary source (Figure 4b). The dominant contribution of the water budget in the tropical anvil (60%) comes from microphysical production (M) in response to the mid-level easterly jet, that enhances the mesoscale circulation and the resulting lifting.

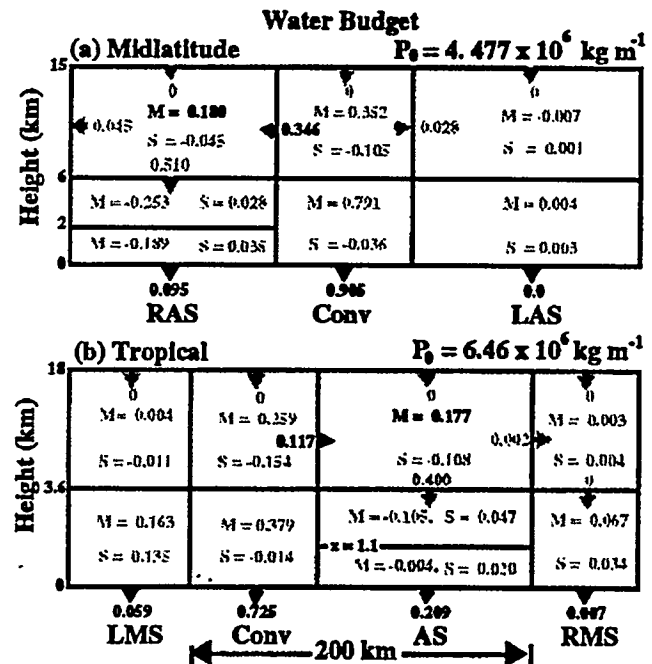


Figure 4. The water budget for the mature stage of the simulations with ice microphysics and without radiation: (a) midlatitude squall-line system over the central subdomain (200 km) and (b) tropical squall-line system over the whole domain (3560 km). All quantities are normalized by the total surface precipitation (P_0). M represents the net microphysical production; S , the storage term; arrows, horizontal and vertical transports; LAS and RAS, leading and rear anvil stratus; Conv, the convective region; LMS and RMS, leading and rear marine stratus; and AS, the anvil stratus.

Without the ice phase, the contribution of hydrometeor transport by deep convection to the water budget of anvil clouds is substantially reduced (not shown). In addition, the contribution of microphysical production in the tropical case without the large-scale ascent (not shown) indicates that large-scale ascent only makes a trivial contribution to the convective region and anvil cloud. However, it does play a significant role in maintaining the marine stratus.

Impact on Cloud Optical Properties

The cloud LW and SW radiative forcings, averaged over the areas of different stratiform cloud regions, are used to assess the impact of stratiform clouds on the radiation budget of the local earth-atmosphere system. The cloud radiative forcing is defined as $F_{\text{clear}} - F_{\text{cloudy}}$, where F is the upward flux at the top of the model atmosphere. As shown in

Figure 5, the cloud SW radiative forcings for all types of stratiform clouds decrease with increasing solar zenith angle, and the tropical anvil cools the local earth-atmosphere system more effectively than the marine stratus and the midlatitude anvil.

Based on Manabe and Moller (1961), the daily mean of the cloud SW radiative forcings for different simulated stratiform clouds (the marine stratus and the anvil cloud) during the summer is calculated to compare with its LW counterpart (Table 1). These SW calculations account for the latitudinal variations of daytime fractional length (f) and daily mean of solar zenith angle (ζ); f and ζ are 0.517 and 54.1° for the tropical case, and 0.637 and 58.6° for the midlatitude case, respectively.

Radiation calculations show that the simulated tropical anvil cloud has larger cloud LW (200 vs. 20 W m^{-2}) and SW radiative forcings (-290 vs. -248 W m^{-2}) than the tropical

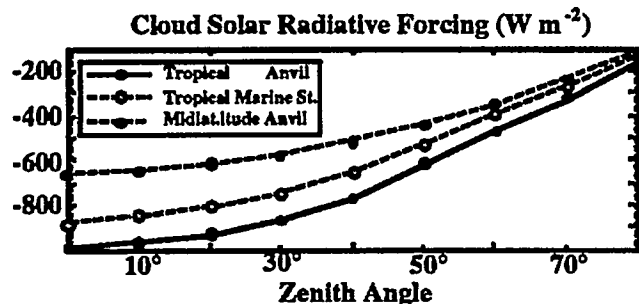


Figure 5. SW cloud forcing averaged over the area of different stratiform clouds at varied zenith angle.

Table 1. Cloud radiative forcing (W m^{-2}) for different stratiform clouds.

Cloud Type/Radiation	LW	SW ^(a)	Total
Tropical Anvil ^(b)	+200	-290	-90
Tropical Marine St. ^(b)	+20	-248	-228
Midlatitude Anvil ^(c)	+190	-236 ^(d) (-212)	-46

(a) Overbar represents the summer-time daily mean, based on Manabe and Moller (1961).

(b) Surface albedo = 0.05 and effective size of ice crystals = 30 μm .

(c) Surface albedo = 0.2 and effective size of ice crystals = 75 μm .

(d) Without considering the latitudinal variations of day-time length and solar zenith angle.

marine stratus. This result contradicts a general belief that low/high-level clouds act to cool/warm the earth-atmosphere system. Nevertheless, the optically thick anvil cloud was proposed to act as a thermodynamical regulator to limit sea surface temperature (Ramanathan and Collins 1991) using the Earth Radiation Budget Experiment data. Although the anvil cloud has a smaller total (LW + SW) cloud radiative forcing than the marine stratus, the cloud LW and SW radiative forcings of the anvil cloud are much larger than their marine stratiform counterparts. The climatic effects of the anvil cloud are, however, significant because the LW radiative forcing is largely confined to the atmosphere and its SW counterpart is mainly felt at the surface (Ramanathan 1987). Thus, anvil clouds must be appropriately represented in GCMs to improve cloud radiative feedback on large-scale climate.

Further, simulated tropical oceanic and midlatitude continental anvils have comparable cloud LW radiative forcings (200 vs. 190 W m^{-2}). The comparison of cloud SW radiative forcings between tropical oceanic and midlatitude continental anvils (-290 vs. -236 W m^{-2}) also shows that the tropical oceanic anvil plays a more important role than the midlatitude continental anvil in reducing the effect of SW insolation on the earth-atmosphere system. Without considering the latitudinal variations of f and ζ , this difference is increased from 54 to 78 W m^{-2} by lowering the SW forcing of midlatitude continental anvil to -212 W m^{-2} . Based on these calculations, the larger SW radiative forcing of the tropical oceanic anvil is mainly attributed to a smaller surface albedo and smaller ice crystal size with both factors about equally contributing to the difference, while the variation of f and ζ can cancel out this difference by about one third.

Summary and Discussion

Our results show that the tilting structure of MCSs in an environment with substantial wind shear makes an important contribution to the water budget of anvil clouds, particularly the tropical anvil due to the impact of the mid-level easterly jet on enhancing the mesoscale circulation. This implies that a cumulus parameterization would considerably underestimate the cloud heating and drying effects of the tropical squall-line system if the mesoscale organization of MCSs was not considered. Comparisons of water budgets between mid-latitude and tropical anvils also have an important implication for climate research due to the strong SW radiative forcing of the tropical anvil. The moisture budgets also indicate that tropical deep convection serves as a more efficient engine pumping low-level moisture into the upper-level anvil cloud than its midlatitude counterpart.

The high sensitivity of the dynamical and microphysical structures of the simulated squall-line systems to wind-shear profiles (jet or non-jet) suggests that momentum transport by clouds in different types of sheared environments may provide an important physical basis linking GCM-resolvable variables with the mesoscale lifting associated with the tilting structure of convective systems. Unfortunately, this tilting structure cannot be resolved by GCM grids and no current cumulus parameterization scheme represents these sub-GCM-grid processes and their resulting feedback on the large-scale environment. Therefore, parameterizing the large-scale effects of these sub-GCM-grid processes associated with the tilting structure of MCSs is a crucial issue that must be addressed to improve the representation of cloud-radiation feedback on large-scale climate.

Acknowledgments

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The Role of Aerosols in Cloud Drop Parameterizations and Its Application in Global Climate Models

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Introduction

The characteristics of the cloud drop size distribution near cloud base are initially determined by aerosols that serve as cloud condensation nuclei and the updraft velocity. We have developed parameterizations relating cloud drop number concentration to aerosol number and sulfate mass concentrations and used them in a coupled global aerosol/general circulation model (GCM) to estimate the indirect aerosol forcing. The global aerosol model made use of our detailed emissions inventories for the amount of particulate matter from biomass burning sources and from fossil fuel sources as well as emissions inventories of the gas-phase anthropogenic SO_2 . This work is aimed at validating the coupled model with the Atmospheric Radiation Measurement (ARM) Program measurements and assessing the possible magnitude of the aerosol-induced cloud effects on climate.

Global Emissions Inventory

In order to provide a global understanding of the effects of aerosols on clouds, one must first understand the global concentrations of the different aerosol components or types. This task consists of the development of a source emissions inventory for aerosols and for gas-phase species that form aerosols. In particular, we are interested in natural and anthropogenic emissions of gaseous sulfur species (which form sulfate in the aerosol), soot or black carbon emissions, emissions of particulate organic carbon and of gas phase non-methane hydrocarbon species which produce aerosols during photooxidation, nitrogen oxides (which form nitrate in the aerosol), and ammonium. For some of these species we already have relatively well-developed emissions inventories [e.g., global emissions of sulfur have been developed by Spiro et al. (1992) and global emissions of NO_x have been developed by Penner et al. (1991)]. Also, we updated both anthropogenic NO_x and SO_2 emissions through the International Global Atmospheric Chemistry Global Emissions Inventory Activity (Penner et al. 1994). The global emissions inventories for black carbon and organic aerosols we have developed include 1) detailed emissions inventories

for the amount of particulate matter (e.g., black carbon, organic carbon, and minor species) produced from wood and dung fuel, charcoal burning and charcoal production, agricultural fires, and savannahs and forests burning, 2) an organic matter inventory produced from the diesel and coal sources, and 3) natural organic matter emissions (Liousse et al. 1995). We have also developed an initial representation for aerosol dust (Rau 1994).

Lawrence Livermore National Laboratory Global Aerosol Model Development

At present, several aerosol types (organic aerosols from biomass burning, black carbon, nitrate, sulfate, and dust) are treated by our global aerosol transport and removal model, called GRANTOUR. This Lagrangian model may be run either off-line, using the wind and precipitation fields from a general circulation model, or interactively, in a mode that allows alterations of the wind and precipitation fields consistent with currently calculated species or aerosol concentrations. The model has been run in a coupled mode with several GCMs, including, most recently, the Hamburg GCM. Additionally, a version of the model which treats the gas phase chemistry of tropospheric O_3 has recently been developed. Such capability is important for treating the gas to particle conversion of aerosol components whose gas phase sources react with OH prior to forming aerosol species. The accuracy of those new inventories is being tested by comparing the model simulations with observations. Sensitivity analyses have also been performed by varying the aerosol precipitation scavenging rates.

Aerosol-Cloud Interactions

We have worked on parameterizing the effects of anthropogenic sulfate-containing aerosols on initial cloud drop number concentration (Chuang and Penner 1995). We refined our treatment of the aerosol size distributions by specifically calculating the processes by which aerosol sulfate is produced in the atmosphere, condensation of

H_2SO_4 on pre-existing particles, and the formation of sulfate by in-cloud oxidation of SO_2 . The pre-existing particles with a prescribed size distribution are assumed to mainly consist of organic matter, black carbon, and natural sulfate. The resulting anthropogenic sulfate-containing aerosol size distribution is then used in a detailed microphysical model to provide a parameterization of the cloud drop number concentration for use in the climate model. This parameterization is being introduced into the coupled model of GRANTOUR/National Center for Atmospheric Research Community Climate Model (NCAR CCM1) to test the adequacy and range of the predicted anthropogenic sulfate-containing aerosol cloud forcings and to compare with its direct forcing at different relative humidity. Results which have been multiplied by fraction of surface area are listed in Table 1.

Table 1. Annual average anthropogenic sulfate forcing (Wm^{-2}).

	Indirect	Direct (RH = 80%)	Direct (dry)
S. Hemis	0.144	0.107	0.063
Land	0.047	0.037	0.022
Ocean	0.097	0.070	0.041
N. Hemis	0.449	0.493	0.290
Land	0.199	0.268	0.158
Ocean	0.250	0.225	0.132
Global	0.593	0.600	0.353

Our simulations indicate that current concentrations of anthropogenic sulfate have direct and indirect effects that may be comparable in magnitude and at least locally will tend to mask the warming effects of increased greenhouse gases (Figure 1).

Continuation of Work

The earth's radiation balance is highly sensitive to changes in aerosol characteristics. The direct effects of aerosols on the earth's radiation depend on their single scattering albedo and therefore on their composition. A potentially more important effect of aerosols, however, comes through their impact on cloud microphysical characteristics. Because aerosol concentrations vary, both naturally and because of human activities, it is important to include both effects in GCMs. The goal of the ARM Program is to improve general circulation models and, in particular, the representation of clouds in GCMs; therefore, we plan to couple our aerosol model to the ECHAM model to gain a thorough understanding of aerosol effect on cloud lifetime and its influence on global climate. We will also use the ARM measurements to test and improve the coupled global aerosol, cloud, and general circulation model.

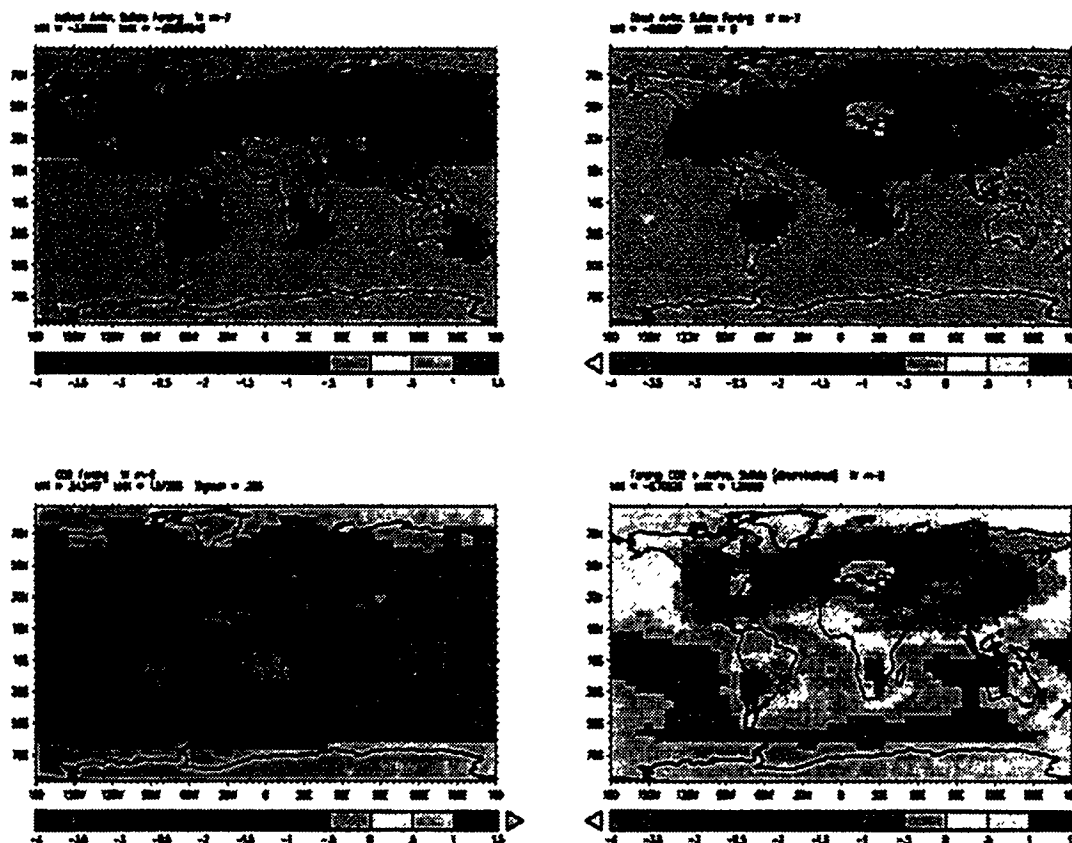


Figure 1. Global distribution of annual average indirect and direct forcing by anthropogenic sulfate-containing aerosols (upper left and right), CO₂ forcing (lower left, from Taylor and Penner 1994) and the total forcing (lower right).

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Performance of Radar Wind Profilers, Radiosondes, and Surface Flux Stations at the Southern Great Plains (SGP) Cloud and Radiation Testbed (CART) Site

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The performance of several routinely operating observational systems at the Southern Great Plains (SGP) Cloud and Radiation Testbed (CART) site has been evaluated. The results of a few specific investigations are shown here for Radar Wind Profilers (RWP), Radio Acoustic Sounding Systems (RASSs), Balloon-Borne Sounding Systems (BBSSs), and Energy Balance Bowen Ratio (EBBR) stations (Coulter and Holdridge 1995; Coulter and Martin 1995; Lesht 1995; Wesely et al. 1995).

Radar Wind Profilers and Radio Acoustic Sounding Systems

A comparison of three months of data on hourly winds and temperatures from the two RWPs collocated at the SGP CART central facility indicated that data from the 50-MHZ

system can be used to determine if the winds measured by the 915-MHZ system are distorted by signal returns from migrating birds. Average differences in the v (north-south) component of the wind measured with the 915-MHZ boundary layer profiler (BLP) versus the 50-MHZ tropospheric profiler (TP) are shown in Figure 1 for the months of June and August. Evidence suggests that the large differences in August were caused by migrating birds passing through the beams of the BLP, particularly with northerly winds at night. Data from the 50-MHZ system, which is unaffected by birds, can be compared to 915-MHZ system data to detect this problem. In the absence of artifacts caused by birds, the agreement between the systems is usually better than 0.5 ms^{-1} .

Table 1 shows statistics generated by comparison of temperatures measured with the BLP and TP RASSs.

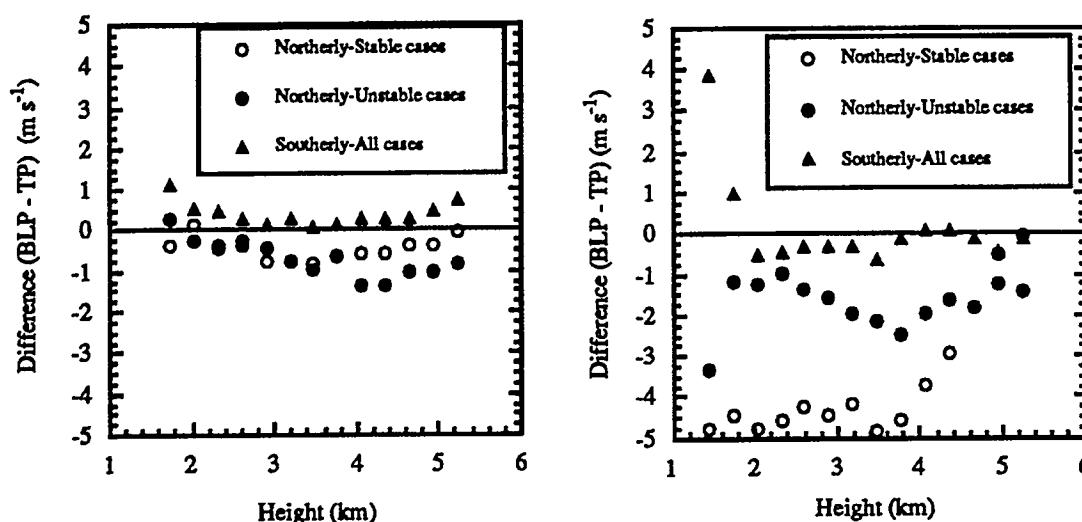


Figure 1. Average differences between the BLP and TP v component of the wind during June and August, according to stability and wind direction.

Table 1. Average and standard deviation of virtual temperature differences ($^{\circ}\text{C}$) between sondes and profilers and between sondes and a linear fit in the zone inaccessible to profilers.

Time (UTC)	Sonde - BLP Average	St. Dev.	Sonde - TP Average	St. Dev.	Sonde - fit Average	St. Dev.
2030	-0.53	0.51	-0.03	0.78	-0.36	0.68
1130	-0.75	0.78	-0.01	0.99	-0.13	1.17
0230	-0.08	0.59	0.33	0.72	-0.03	1.06
0530	-0.41	0.92	0.30	0.75	0.16	1.16
0830	-0.12	0.87	0.56	0.89	-0.25	0.78
1130	-0.85	0.65	0.52	0.73	-0.05	0.63
1430	-0.50	1.07	0.57	0.62	-0.14	1.09
1730	-0.45	0.64	0.56	1.13	-0.10	0.99

Although the virtual temperature profiles from the two systems do not overlap, fitting a polynomial to measurements at the largest three heights where temperatures are reliably detected with the BLP RASS and at the smallest three heights for the TP RASS produces values that agree very well with the radiosonde values in the zone not accessible to either radar system.

Realistic evaluations of statistics on vertical velocity (w) from wind profiler data often require a sample rate larger than that available from normal profiler operations at the CART site; however, the existence of four tilted beams, rather than the minimum of two required for horizontal wind measurements, allows additional measurements of the vertical component of motion with pairs of beams tilted in opposed directions. These tilted beams can be used to provide better statistics on vertical velocity and, potentially, direct measurements of momentum flux. Spectra in convective conditions for w sampled directly at 15-second intervals with only the vertical beam and for w derived from both the vertical and the four tilted beams are shown in Figure 2. These and other tentative results indicate that problems exist with the five-beam approach at high frequencies because increased variance is introduced by spatial and temporal separations in the vertical velocities calculated from opposing tilted beams. Additional noise is apparently introduced by the irregular time spacing of the enhanced time series. On the other hand, the low-frequency resolution of the enhanced time series may be improved, which would aid in the detection and resolution of coherent structures like thermal plumes. Better selection of beam sampling order might be possible to eliminate the irregular

time series. For example, arranging for the vertical low- and high-power beams to operate sequentially and averaging their signals would yield a time series with five different estimates of w that are equally spaced in time.

Balloon-Borne Sounding Systems

A quantitative measure of the quality of the BBSS data is needed because the Atmospheric Radiation Measurement (ARM) Program uses data from the radiosondes in sensitive modeling applications and for purposes of real-time quality control. Finding such a measure is difficult because the basic sensing system, the radiosonde package, is discarded after each use and is not available for repeated calibration checks. Furthermore, methods of evaluating the quality of radiosonde data that depend on point-by-point comparisons with profiles obtained by ground-based remote sensors and sensors aboard aircraft are difficult because they lack simultaneity of sampling in space and time, especially at high altitudes.

From January 1993 through August 1994, the temperature and humidity sensors on radiosondes launched from the SGP CART central facility were checked with a commercially supplied ground-check set before launch. Investigations with the ground-check set showed that the radiosondes are within the manufacturer's specifications for measurement of relative humidity ($\pm 2\%$ for one standard deviation). Because of procedural artifacts, the ground-check data could not be used to similarly verify the specification for

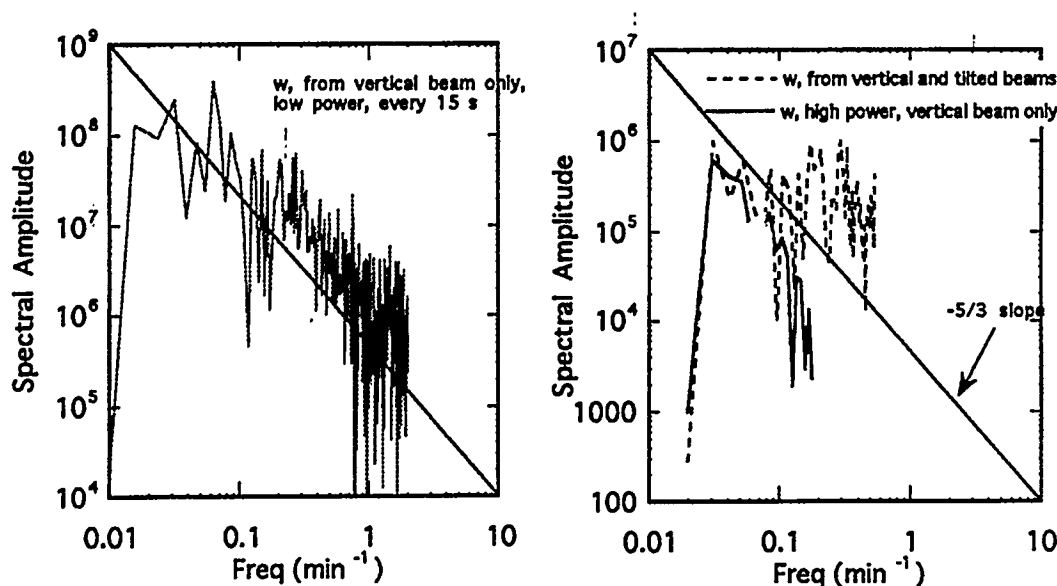


Figure 2. Vertical velocity spectra in convective conditions sampled directly at 15-s intervals (top) on October 13, 1994; and vertical velocities derived from vertical and tilted beams (bottom) on September 28, 1994. Solid straight lines indicate the $-5/3$ slope expected in atmospheric turbulence.

temperature ($\pm 0.2^\circ\text{C}$ for one standard deviation). Examination of the time series of temperature differences showed a clear seasonal pattern, with greater differences occurring during the cold winter months. This pattern suggests that the apparent bias in temperature is an artifact of the ground-check procedure, most likely resulting from a failure of the ground check set to be in thermal equilibrium when the instrument shelter is being heated during the winter. The variability in the differences probably reflects the variable experience of the different operators. Ground checks were suspended in August 1994 because analysis of the accumulated data indicated that the potential for introducing errors into the sounding data as a result of procedural errors outweighed potential improvements over the factory-supplied calibrations.

Figure 3 shows ascending and descending phases of a sounding initiated at 2031 hrs UTC on May 10, 1994. Analysis of such pseudo-replicate data from ascending and descending phases of operational soundings suggests that the temperature measurement is within the $\pm 0.2^\circ\text{C}$ specified for the system, but more analysis is required to confirm the accuracy of the temperature sensor aloft. On average, data collected during the ascending and descending phases of the soundings are quite similar, but the detailed structure during the relatively rapid descent may be obscured by loss of sampling density. Uncertainty about the physical position of

the radiosonde sensors during descent complicates the situation, but the set of ascending and descending soundings obtained at the SGP CART site represents a rich source of data for examination of both operational radiosonde sensor response and short time-scale changes in the state of the atmosphere.

Energy Balance Bowen Ratio Stations

The 350-km by 400-km domain of the SGP CART site is equipped with 10 EBBR stations at grassland sites. These stations operate continuously throughout the year and use an automatic exchange mechanism to switch the positions of the upper and lower sets of temperature and humidity sensors once every 15 min to minimize the effects of offsets and minor drifts in the all-weather sensors. The Bowen ratio (β) of the sensible heat flux (H) to the latent heat flux (λE) is computed with the energy balance equation, which tends to produce "spikes" in the heat flux data when the value of β is near -1. On some days, no spikes occur, but one to five spikes per day are common, particularly near sunrise and sunset. Incorrect estimates of sensible and latent heat fluxes occur mostly for values of β between approximately -0.75 and -1.5. These suspect data points could be replaced with data from a bulk aerodynamic approach. After further

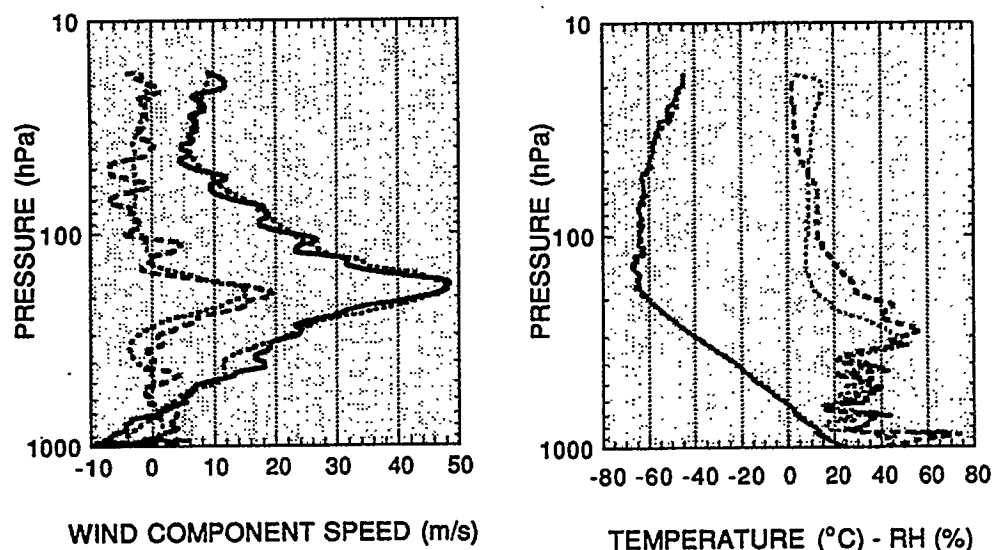


Figure 3. Ascending (solid heavier lines) and descending (lighter dashed lines) phases of sounding on 10 May 1994 at 20:31 UTC. Panel on left shows east and north wind components; panel on right shows temperature and relative humidity.

testing, algorithms for aerodynamic approach might be implemented to provide a routine data product from the SGP CART Site.

A significant source of difficulty in estimating average H and λE values across the CART site can be spatial variability, as illustrated in Figure 4. Here, averages for 0800 to 1600 hrs CST on each day are shown for two sites, one (13, central facility) near the center of the overall CART site and one (22, Cordell) in the southwest corner. Such averages of fluxes for August 1994 varied strongly among the 10 EBBR stations. Because the surfaces were all grass covered and because the momentum flux did not vary greatly from site to site, the likely cause of the variations in H and λE were differences in soil moisture. Obtaining a representative average of the fluxes for the CART site in such a situation requires a more sophisticated approach than simple averaging.

Conclusions

A comparison of three months of data on hourly winds and temperatures from the two RWP collocated at the central facility indicated that data from the 50-MHZ system can be

used to determine if the winds measured by the 915-MHZ system were distorted by signal returns from migrating birds. An investigation of vertical winds obtained from vertically pointing and tilted beams of the 915-MHZ profiler led to suggestions on methods to enhance the time series of vertical velocities. An evaluation of the CART radiosonde system led to the conclusion that it provides data within the specified uncertainties, and its parachute-assisted descending phase provides usable temperature and humidity data below heights corresponding to atmospheric pressures above about 200 mb. An examination of the surface heat fluxes from the EBBR stations showed that the fluxes varied widely from east to west across the CART site during August 1994 and that algorithms are needed to remove data "spikes" in the computed fluxes when the Bowen ratio is near -1.

Acknowledgments

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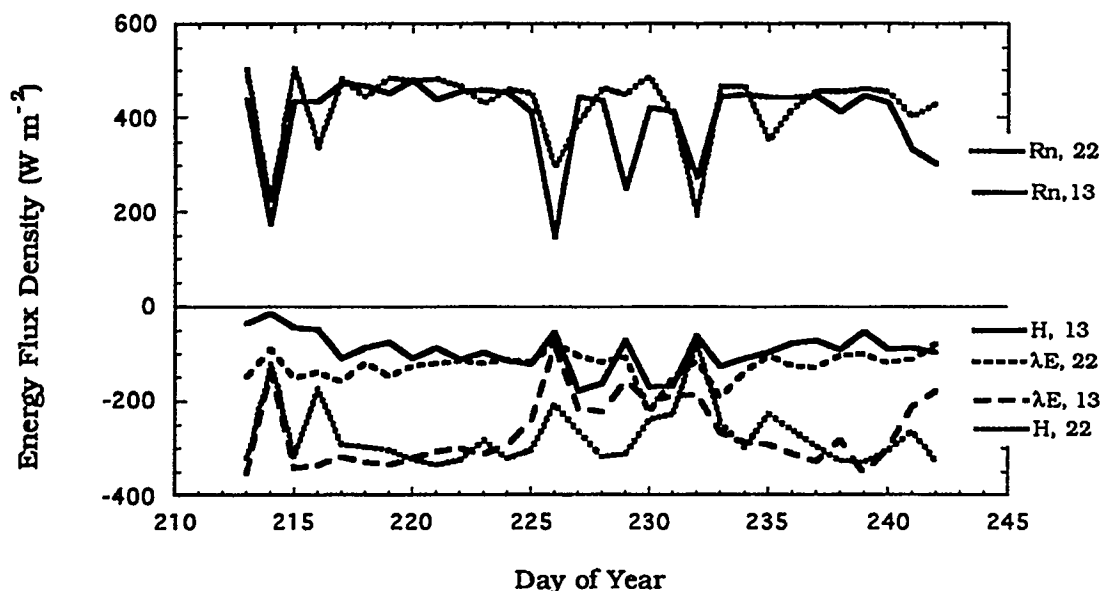


Figure 4. Daily averages of Net Radiation (Rn), H, and λE between 0800 and 1600 hrs CST during August 1994 at two EBBR stations selected for their large differences in H and λE values.

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Using Remotely Sensed Planetary Boundary Layer Variables as Estimates of Areally Averaged Heat Flux

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Introduction

Homogeneity across the Southern Great Plains (SGP) Cloud and Radiation Testbed (CART) site is an issue of importance to all facets of the Atmospheric Radiation Measurement (ARM) Program. The degree to which measurements at the central facility can be used to verify, improve, or develop relationships in radiative flux models that are subsequently used in global circulation models, for example, is tied directly to the representativeness of the local measurements at the central facility for the site as a whole.

The relative variation of surface energy budget terms over a 350-km x 400-km domain such as the SGP CART site can be (and is) extremely large. The Planetary Boundary Layer (PBL) develops as a result of energy inputs from widely varying surfaces. The lower atmosphere effectively integrates the local inputs; measurements of PBL structure can potentially be used for estimates of surface heat flux

over scales on the order of tens of kilometers. This project is focusing on two PBL quantities that are intimately tied to the surface heat flux: 1) the height of the mixed layer, z_1 , that grows during daytime due to sensible heat flux input from the surface; and 2) the convective velocity scale, w^* , normally a scaling parameter defined by the product of the sensible heat flux and z_1 , but in this case defined by coherent structures (thermal plumes in the convectively unstable PBL) that connect the surface layer and the capping inversion that defines z_1 .

Variability of Heat Flux Across the CART Site

Figure 1 illustrates the variability of sensible heat flux, H , at 1230 local solar time (LST), across the SGP CART site over monthly periods, expressed in terms of the square root of the normalized spatial variance, V_s , and the normalized temporal variance, V_t :

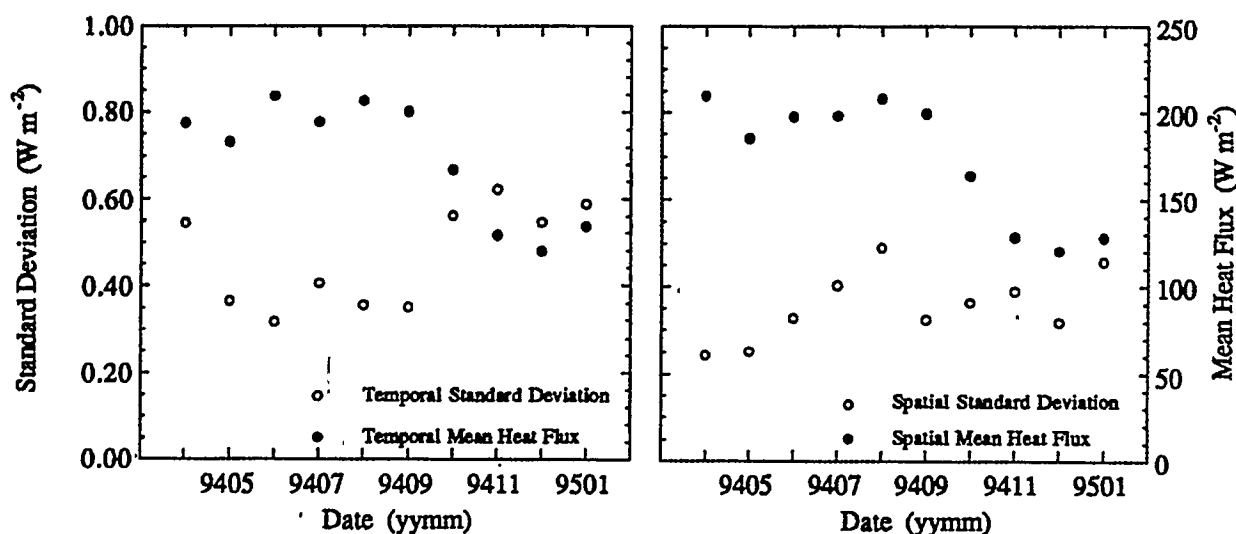


Figure 1. Temporal and spatial standard deviation of sensible heat flux over the SGP CART site. Values are normalized by the mean sensible heat flux, as defined in the text.

$$V_t = \frac{\sum_{i=1}^N \sum_{t=1}^M (H_{i,t} - \bar{H}_t)^2}{\sum_{i=1}^N \sum_{t=1}^M H_{i,t}^2} \quad V_s = \frac{\sum_{t=1}^M \sum_{i=1}^N (H_{i,t} - \bar{H}_s)^2}{\sum_{t=1}^M \sum_{i=1}^N H_{i,t}^2} \quad (1)$$

$$\bar{H}_t = \frac{1}{M} \sum_{t=1}^M H_{i,t} \quad \bar{H}_s = \frac{1}{N} \sum_{i=1}^N H_{i,t}$$

where $H_{i,t}$ is the sensible heat flux measured at site i and time t and where $\bar{H}_t$ and $\bar{H}_s$ are the temporal and spatial means, respectively. Values provided from the Energy Balance Bowen Ratio (EBBR) systems were modified with an aerodynamic calculation of H when the Bowen ratio ($\beta = H/LE$) was between -0.75 and -1.25 (Wesely et al. 1995). Large values (>0.5) in Figure 1 are invariably associated with small mean values, usually during heavily overcast or precipitation conditions. The magnitude of the spatial variance is only slightly smaller than the temporal values. The true spatial variance may well be larger because present locations for heat flux measurements are solely over pasture and grassland. The partitioning of energy among ground, latent, and sensible fluxes will likely be different over cultivated fields such as wheat and alfalfa, which are currently being instrumented.

The primary causes of the spatial variability likely are local values of soil type, soil moisture, and cloud cover. Climatological variations such as NW-SE gradients in rainfall and cloudiness are compounded by local, short-term events such as recent rainfall, weather systems, and local circulations within the CART site. In addition, reductions in net radiation due to transient fair-weather cumulus clouds that dominate over the site can change local short-term (30-min) measurements of heat flux considerably.

Planetary Boundary Layer Estimates of H

Mixed Layer Height

The growth of the mixed layer during daytime is driven primarily by sensible heat flux to the atmosphere from the surface. Boers et al. (1984) developed relationships for the rate of rise of z_i . In the absence of entrainment,

$$\frac{dz_i}{dt} = \frac{H}{\rho c_p \gamma z_i} \quad (2)$$

where γ is the lapse rate through the capping inversion, c_p is the density of air, and c_p is the heat capacity of air at constant pressure. With entrainment,

$$\frac{dz_i}{dt} = \frac{C\sigma_w}{\left(B + \frac{gz_i\delta\theta}{T_0\sigma_w^2} - \frac{A(\delta u)^2}{\sigma_w^2} \right)} \quad (3)$$

where C , B , and A are constants parameterizing entrainment of heat flux, $\delta\theta$ and δu are the temperature and horizontal wind speed jumps across the capping inversion, T_0 is the surface temperature, and σ_w is the standard deviation of vertical velocity. In convective conditions, $B = A = 0$, and Equation (3) becomes

$$\frac{dz_i}{dt} = \frac{CT_0\sigma_w^3}{gz_i\delta\theta} - \frac{CH}{\rho c_p \delta\theta} \quad (4)$$

where σ_w^3 has been equated to the convective velocity scale (see below). We see that both Equations (2) and (4) relate the growth rate directly to the surface heat flux.

Because z_i usually reaches 1 to 3 km in convective conditions, one might expect that surface influences over a scale roughly 10 km or larger contribute to mixed layer growth when advective effects are included. An important point is that an instantaneous measurement of z_i is related to H_t , the time-integrated value of H since mixed layer growth began. This fact is useful because it provides a method for reducing the effects of short-time fluctuations on H (such as cumulus clouds suggested previously) when calculating spatial variability. Figure 2a shows an example of the variation of H , H_t , and z_i during daylight hours, and Figure 2b shows the relationship between z_i at the central facility and H_t over an extended period in July-August 1994. When H_t is averaged over sites neighboring the central facility, the correlation between z_i and H_t is considerably improved (Figure 2b). It is not immediately evident that including neighboring extended facilities should improve the correlation because those facilities are separated by roughly 50 km; the averaging process may account for some of the local variability near the central facility. Additional measurements in upcoming Intensive Operating Periods (IOPs) may be able to address this question directly.

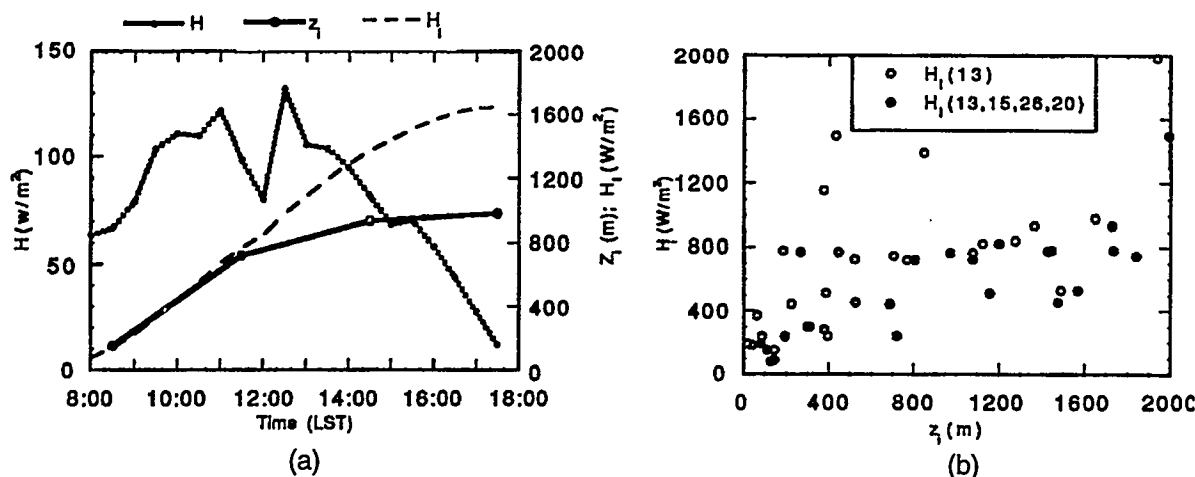


Figure 2. (a) Variations of heat flux, integrated heat flux, and mixed layer depth at the central facility on July 23, 1994; (b) Variation of mixed layer height at central facility (site No. 13) with integrated heat flux at the central facility and integrated heat flux averaged for the central facility and neighboring extended facilities. The time step for integration is 30 min.

Coherent Structures

The primary method by which the PBL transfers energy from the surface through the mixed layer into the capping inversion is through coherent structures such as thermal plumes (Coulter et al. 1993; Stull 1989). Because these structures maintain their identity for 30-60 min and move with the mean wind, they effectively integrate over the surfaces over which they travel. The convective velocity scale, w^* , is a measure of the strength of that transfer,

$$w^* = \frac{g}{T_0} \frac{H}{\rho c_p} z_i \quad (5)$$

where g is acceleration due to gravity. Some researchers (e.g., Boers et al. 1984) have chosen to define a measurement of w^* in terms of the standard deviation, σ_w , of the vertical velocity, w . We choose to estimate w^* by equating it with the vertical velocity associated with the thermal plumes most responsible for energy transfer because of their well-defined lifetime. In particular, we postulate that

$$w^* = \alpha w_T(0.5z_i) \quad (6)$$

where $w_T(0.5z_i)$ is the mean vertical velocity measured within thermal plumes at $z = z_i/2$ and α is a constant to be determined. We investigated this relationship with data from the Boardman ARM Region Flux Experiment (BARFEX) in 1992 (Doran et al. 1992). We used minisodar

data within the lowest 200 m above desert steppe to select well-organized thermal plumes (defined as times with $w > 15$ cm/s for at least 25 s). The vertical velocity profile within the thermal plumes was then modeled as

$$w_T = a + b \sin(z\pi/z_i) \quad (7)$$

and extrapolated (where necessary) to $z_i/2$ to provide an estimate of $w_T(0.5z_i)$. Airsonde launches provided measurements of z_i , and H was calculated from an average of three eddy correlation stations located in the steppe region. Figure 3 shows the relationship between $w_T(0.5z_i)$ and w^* that suggests a value of $\alpha = 1.1$.

During an IOP in October 1994, the 915-MHz wind profiler was operated in a vertical-only mode that allowed estimates of w to be made every 15 s. Although this sample rate is roughly 10 times slower than that used with the minisodar, it was sufficient to sample coherent structures within the mixed layer in light wind conditions. The profiler provided direct estimates of w_T , as well as estimates of z_i , through inspection of the vertical time section of the intensity of the signal return and through analysis of Balloon Borne Sounding System (BBSS) launches from the central facility. Data were limited to 1200 to 1600 LST each day for two weeks. In this case, the thermal plumes were defined by using a wavelet transform technique described by Coulter and Li (1995). Values of $w_T(0.5z_i)$ within the regions selected were limited to those greater than 15 cm/s. The averaging time required by the profiler approximated the time for thermal definition used with the BARFEX data.

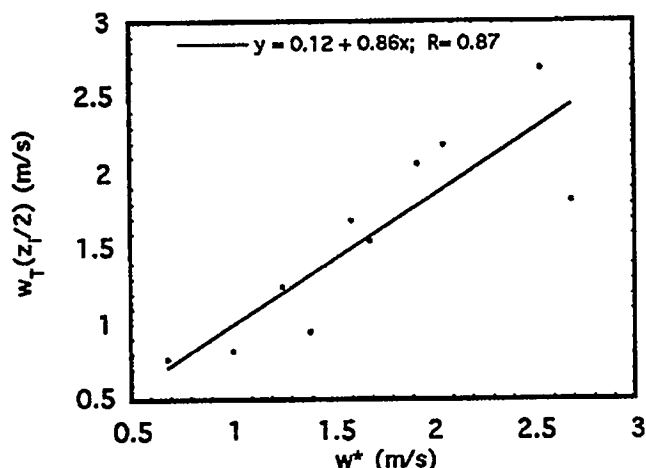


Figure 3. Modeled mid-thermal vertical velocities (from minisodar profiles) as a function of w^* obtained above desert steppe on June 5 and June 14, 1992.

Figure 4 shows the relationship between w^* and w_T at the CART site. In this case a value of $\alpha=1.1$ again seems appropriate. The coefficient required may be greater than 1 because of incomplete sampling of the thermal plumes (sampling the edge rather than the center of the plume) or

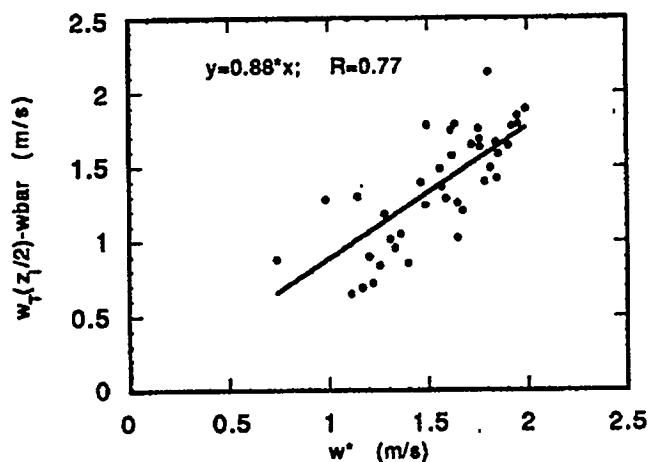


Figure 4. As in Figure 3, except mid-mixed layer vertical velocities are calculated directly from 915-MHz profiler data between Oct. 10 and Oct. 23, 1994 at the central facility of the SGP CART site.

because of non-representative estimates of H at the central facility. If so, these data would suggest that local H is larger at the central facility than at surrounding areas during the IOP. Future IOPs will focus on answering some of these questions and developing methods to use more rapid sampling of the vertical component of motion without compromising mean horizontal wind data.

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Modeling the Summertime Arctic Cloudy Boundary Layer

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Introduction

Global climate models have particular difficulty in simulating the low-level clouds during the Arctic summer. Model problems are exacerbated in the polar regions by the complicated vertical structure of the Arctic boundary layer. The presence of multiple cloud layers, a humidity inversion above cloud top, and vertical fluxes in the cloud that are decoupled from the surface fluxes, identified in Curry et al. (1988), suggest that models containing sophisticated physical parameterizations would be required to accurately model this region. Accurate modeling of the vertical structure of multiple cloud layers in climate models is important for determination of the surface radiative fluxes.

This study focuses on the problem of modeling the layered structure of the Arctic summertime boundary-layer clouds and in particular, the representation of the more complex boundary layer type consisting of a stable foggy surface layer surmounted by a cloud-topped mixed layer. A hierarchical modeling/diagnosis approach is used. A case study from the summertime Arctic Stratus Experiment is examined. A high-resolution, one-dimensional model of turbulence and radiation is tested against the observations and is then used in sensitivity studies to infer the optimal conditions for maintaining two separate layers in the Arctic summertime boundary layer. A three-dimensional mesoscale atmospheric model is then used to simulate the interaction of this cloud deck with the large-scale atmospheric dynamics. An assessment of the improvements needed to the parameterizations of the boundary layer, cloud microphysics, and radiation in the 3-D model is made.

Case Study

The case chosen occurred at 0 Z June 29, 1980, in the Beaufort Sea during the Arctic Stratus Experiment. Measurements of the mean and turbulence structure of the

boundary layer along with cloud microphysical and radiative properties were obtained from the National Center for Atmospheric Research (NCAR) Electra; details of these measurements may be found in Tsay and Jayaweera (1983), Herman and Curry (1984), Curry and Herman (1995), Curry (1986), and Curry et al. (1988).

The synoptic situation on June 28, 1980, consisted of a mature anticyclone centered over the Beaufort Sea. A large region of low-level cloudiness was present in the north-west region of the anticyclone. This cloud deck was approximately 300 km² in horizontal extent, with the surrounding region clear except for altostratus to the north. Weak rising motion of about 0.2 cm s⁻¹ was measured. Figure 1 shows mean vertical profiles of temperature, water vapor, cloud water mixing ratio, and horizontal wind components obtained with the NCAR Electra. The vertical structure of this cloud system, typical of many "multilayered" systems in the Arctic, consisted of a stable surface fog layer about 250-m deep surmounted by an upper cloud deck with a base at 700 m. Mean budgets of heat and moisture demonstrated that the horizontal advection terms were an order of magnitude smaller than the largest terms in the budgets.

1-D Model Results

The complete model equations have been reported in McInnes and Curry (1995a). Only a brief summary is provided here. The basic mean prognostic variables are the horizontal momentum (U, V), the liquid water potential temperature (Θ_l), and the total water mixing ratio (Q_w). Turbulent fluxes of heat, momentum and moisture are obtained from a second-order, level-3 turbulence closure scheme. The turbulence scheme is coupled to a non-precipitating statistically based cloud model. A parameterization is included for the gravitational settling of liquid water droplets. The model used to compute the radiative fluxes was developed by Morcrette (1991) for use by the

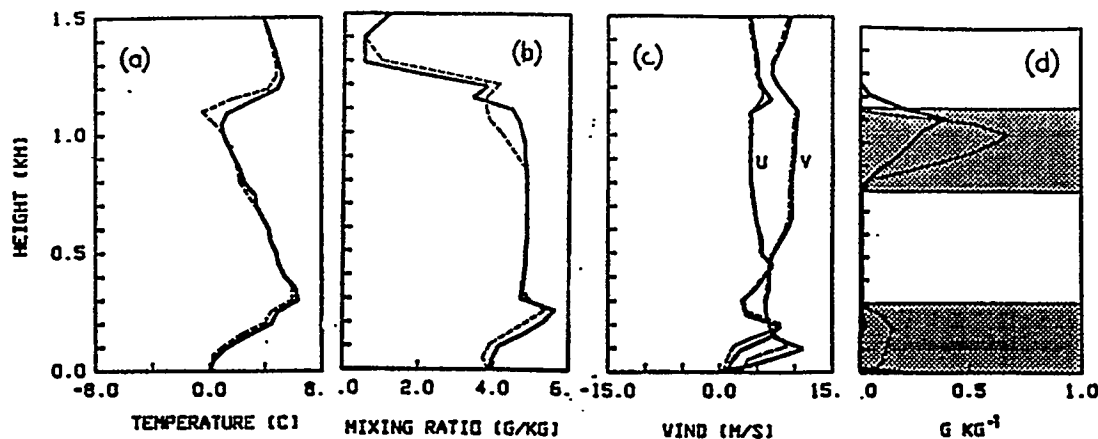


Figure 1. Comparison of observed (solid) and modeled (dashed) mean profiles of a) temperature, b) water vapor mixing ratio, c) winds, and d) liquid water content valid at 0Z June 29, 1980.

European Centre for Medium Range Weather Forecasting (ECMWF) model. The radiative contributions of clouds are modeled following Curry and Ebert (1992) to be a function of the cloud liquid water content and particle effective radius. The cloud water content is determined by the model, while the particle effective radius is specified as an external parameter.

Comparison of the model results with the observations (Figure 1; see also McInnes and Curry 1995a) show that the one-dimensional model described can capture the complex boundary-layer structure which is observed in the Arctic during the summer months. The main deficiency in the model was a slight over-estimation of liquid water content in the cloud layers and under-estimation of the predicted turbulence variables. The following results were obtained by McInnes and Curry (1995a) from sensitivity studies using the 1-D model:

- Radiative effects in the model are responsible for the generation of turbulence in the upper cloud deck and for the condensation of liquid water.
- A uniform large-scale ascent imposed on the profile leads to an elevation and thickening in vertical extent of the two cloud layers. Large-scale descent produces strong turbulent kinetic energy, which eventually mixes out the vertical gradients in the mean profiles, and evaporation eventually destroys the cloud.

- The optimum vertical resolution required by the model to most accurately model the observed features of the boundary layer for the case studied here is found to be 25 m over the lowest 2 km of the atmosphere. However, when the vertical resolution is degraded to 200 m, the model still captures the broad qualitative features of the observed boundary layer.

The 1-D model is also used to examine the formation mechanism for the cloud layering. Two hypotheses have been proposed to explain the layering. Using a simple one-dimensional radiative-diffusive model, Herman and Goody (1976) proposed a hypothesis for the cloud layering in which cloud absorption of solar radiation warms the intermediate depths of the cloud layer, causing evaporation in the interior of the cloud. Using observations from the Arctic Stratus Experiment, Tsay and Jayaweera (1984) proposed that the upper cloud layer is an advective fog, the low layer forming due to surface convection. Using the 1-D model, McInnes and Curry (1995b) propose that the upper cloud layer forms initially by radiative and diffusive cooling of warm, moist air that is advected into the Arctic basin, with the subsequent evolution of a cloud-topped mixed layer generated by the cloud-topped radiative cooling. The cloud-topped mixing does not produce sufficient turbulence kinetic energy to overcome the surface stable layer and penetrate down to the surface. The air in the surface inversion layer is warmer than both the surface and cloud overhead and thus cools radiatively, eventually forming a radiation fog. An example of the modeled evolution of a two-layer cloud deck is shown in Figure 2.

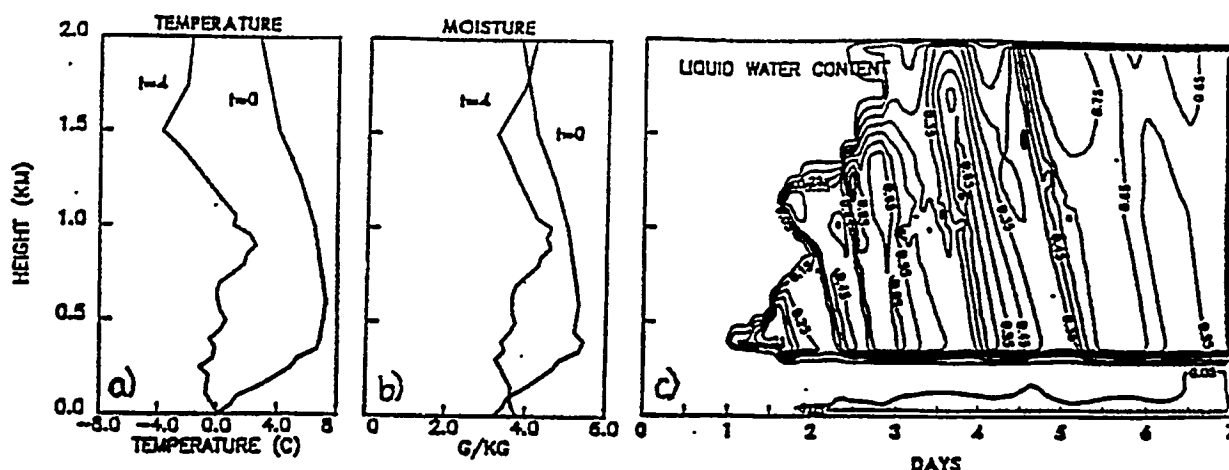


Figure 2. Modeled evolution of layered cloud system: a) temperature profiles at $t = 0$ and at $t = 4$ hours; b) humidity profile at $t = 0$ and $t = 4$ hours; and c) time evolution of cloud liquid water content.

3-D Model Results

The model used for the 3-D simulations is Version 5.1 of the Pennsylvania State University (PSU)/NCAR Mesoscale Model (MM5). For a complete description of the model, see Grell et al. (1994). In this study we use a horizontal resolution of 40 km and a vertical resolution which varies with height from about 70 m just above the surface to 1 km near the top of the troposphere. The model is run using the hydrostatic option. For moist physics, the Dudhia (1989) simple ice phase and Hsie (1984) stratiform warm-phase precipitation formulations are used. Moist vertical diffusion is allowed and an upper radiative boundary condition is employed to allow energy to pass through the atmosphere unreflected while the lateral boundaries are relaxed toward the large-scale analysis. The two-stream radiation scheme of Dudhia (1989) is used for radiative transfer computations, and Blackadar's high-resolution planetary boundary layer scheme (Zhang and Anthes 1982) is used to vertically mix momentum, moisture variables, and temperature in the boundary layer. The surface temperature and albedo were specified at 271.4 K and 0.55, respectively, in accordance with the observations described by Herman and Curry (1984) for this case and remain constant over the integration period. The model is initialized using ECMWF analyses for June 27, 1980, at 12 Z and is run for a period of 60 hours.

Model results are examined after 36 hours integration, corresponding to the time of the observations shown in Figure 1. The modeled surface pressure field and liquid water content at 960 mb are shown in Figure 3. The spatial coverage of cloud predicted by the model is much larger than that observed in the satellite visible image.

Comparison of the modeled liquid water content (Figure 4a) with the observations (Figure 1) indicates that the cloud modeled by the 3-D model consists of a single-layered cloud with base at the surface. The cloud microphysics scheme produces unrealistically large amounts of cloud water. This overestimation of cloud water has a large impact on the net radiative fluxes at the surface (Figure 5). The shortwave flux is greatly attenuated by these optically thick clouds with incident shortwave radiation at the surface of less than 10% of the top-of-the-atmosphere values.

It appears that the conversion of cloud water to rain occurs much too slowly allowing for the buildup of cloud water in the atmospheric boundary layer for the baseline run (Figure 3b; Figure 4b). In the baseline simulation, we used an autoconversion threshold of 0.5 g kg^{-1} . Observations described by Curry (1986) suggest that drizzle begins at much lower values of liquid water content in the Arctic stratus clouds. Therefore, in Figure 4b we show the time evolution of cloud water content for an autoconversion threshold of 0.25 g kg^{-1} . This effectively reduced the total amount of cloud water by allowing for rain at much lower cloud water mixing ratios. The autoconversion threshold is certainly not a universal constant. For example, one would expect different rates of autoconversion depending on the drop size distribution and the size of the largest drops. These parameters are dependent on the concentration, type, and size of available condensation nuclei which vastly differ over climatic regimes (e.g., continental versus maritime).

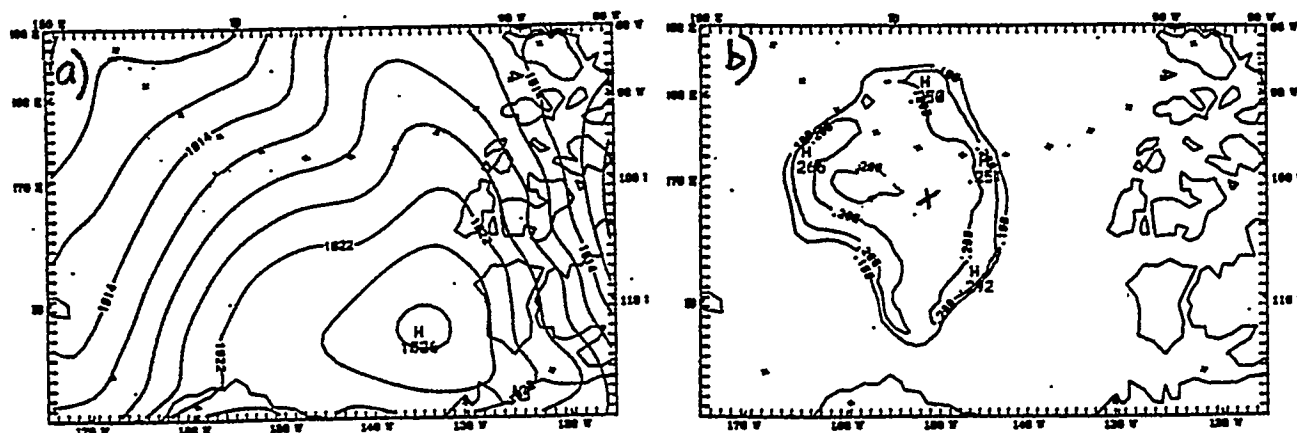


Figure 3. Model simulation of a) surface pressure field and b) cloud liquid water mixing ratio at 960 mb for the same time as the observations in Figure 1 "X" denotes the location of the aircraft measurements described in Figure 1.

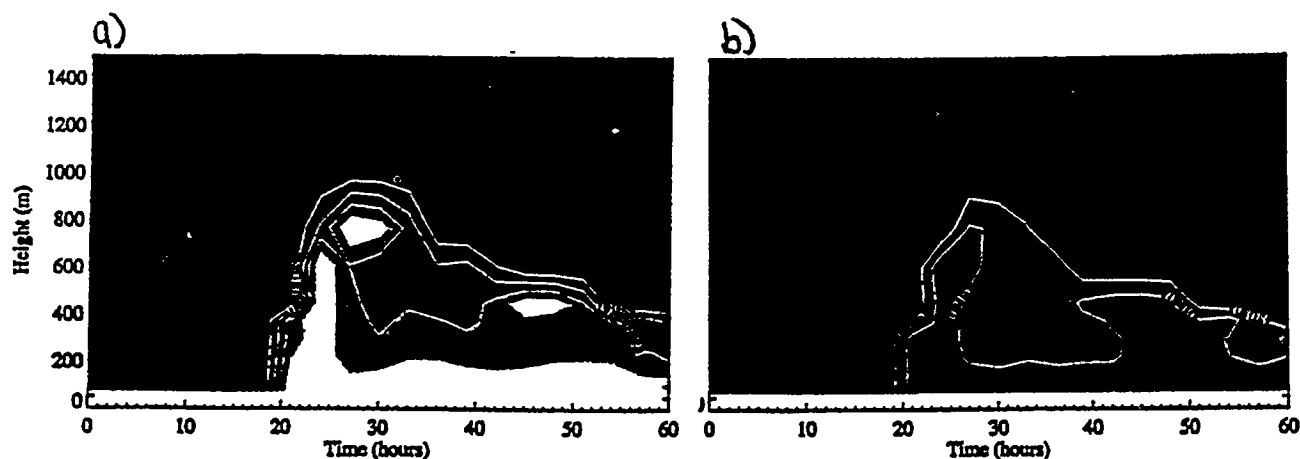


Figure 4. Modeled time evolution of cloud liquid water mixing ratio (g kg^{-1}) using the 3-D model, at the location corresponding to the observations in Figure 1: a) autoconversion threshold of 0.5 g kg^{-1} ; b) autoconversion threshold of 0.25 g kg^{-1} .

The importance of an accurate depiction of cloud microphysical processes is seen in Figure 5. After the onset of cloud, the curves for the incident shortwave radiative diverge by as much as $30\text{--}40 \text{ W m}^{-2}$.

Conclusions

These results imply that

- The layered cloud formation requires a fairly sophisticated boundary layer parameterization such as second-order closure.
- Modeled cloud water content is very sensitive to the autoconversion threshold in bulk microphysical parameterizations, and drizzle must be included to control the modeled liquid water content.
- Surface shortwave radiation fluxes are very sensitive to the modeled cloud microphysical characteristics.

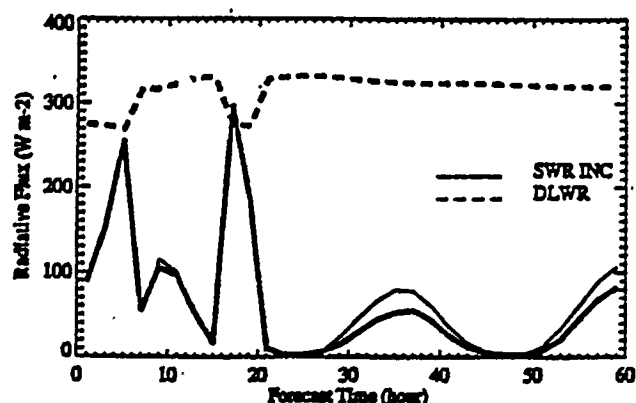


Figure 5. Time series of modeled surface radiation fluxes. DLWR: downwelling longwave radiation; SWR INC: incoming shortwave radiation. For SWR INC, the bold line corresponds to the water content in Figure 4a and the thin solid line corresponds to the water content in Figure 4b.

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Observed and Simulated Temperature Dependence of the Liquid Water Path of Low Clouds

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Disagreements among climate models as to the sign and magnitude of cloud feedback to be expected in a warming climate arise largely because of their differing treatments of cloud optics feedbacks. Models with prescribed/diagnosed cloud optical properties tend to give positive/negative cloud feedback, while those with prognostic cloud water budgets can give feedback of either sign depending on the details of the parameterization. Available observations provide conflicting evidence about the temperature dependence of cloud optical thickness in the current climate, at least for low-level clouds. Simple thermodynamic arguments (Betts and Harshvardhan 1987) and in situ observations (Feigelson 1978) suggest that liquid water content should increase with temperature. But satellite retrievals from the International Satellite Cloud Climatology Project (ISCCP) indicate that the optical thickness of low clouds increases with temperature only at cold temperatures and primarily over land; at warm temperatures, optical thickness decreases with temperature (Tselioudis et al. 1992).

The most recent version of the Goddard Institute for Space Studies (GISS) general circulation model (GCM) includes a prognostic cloud water budget parameterization for stratiform clouds (Del Genio et al. 1996). Unlike the previous version of the GCM, which prescribes cloud optical properties, the prognostic cloud water version successfully reproduces the ISCCP-observed transition from increasing to decreasing low cloud optical thickness with temperature as temperature increases. Assuming that this behavior is real and not an artifact of subpixel cloud cover variations, there are three possible explanations: 1) Microphysical sinks of cloud water such as precipitation or entrainment become more efficient in warmer environments, 2) The physical thickness of clouds systematically decreases with temperature, and 3) Droplet effective radii are much larger in warmer clouds. Sensitivity tests designed to suppress variations in each of these parameters in the GCM suggest that optical thickness decreases with temperature are most sensitive to cloud physical thickness variations, but observational confirmation is lacking.

Data being acquired at the Atmospheric Radiation Measurement (ARM) Southern Great Plains (SGP) Cloud and Radiation Testbed (CART) site can be used to examine the factors determining the temperature dependence of cloud optical thickness. Although cloud droplet effective radius variations may also be important for the satellite-observed optical thickness behavior, we focus initially on cloud liquid water and physical thickness variations, which can be derived from existing routine ARM measurements. For the analysis, we require nearly coincident data from four ARM instruments: 1) Microwave Water Radiometer (MWR) retrievals of vertically integrated cloud liquid water path; 2) ceilometer estimates of cloud base height; 3) Geostationary Operational Environmental Satellite (GOES) infrared brightness temperatures, which indicate cloud top height; and 4) radiosonde temperature profiles, which allow us to translate infrared (IR) brightness temperature into cloud top height and also indicate the mean cloud temperature. We restrict the analysis to low-level clouds with no IR-detectable clouds overhead by selecting only data for which the nearest (in time) colocated satellite pixel indicates a cloud top pressure >680 mb (the ISCCP definition of low clouds), and we ignore differences between IR brightness temperature and actual cloud top temperature, which should be small for clouds whose liquid water signature is detectable in the microwave. We also restrict analysis to cloud top temperatures >273 K, to eliminate biased impressions caused by the insensitivity of the MWR to cloud ice.

The MWR data set is routinely reported as 5-minute average liquid water paths (although higher resolution is available with decreased signal/noise ratio). Typical horizontal wind speeds at the SGP site in the lower troposphere are 5-10 m/s, implying an effective spatial resolution of 1.5-3.0 km, slightly better than ISCCP. Averaging the data on 10- or 15-minute time intervals does not qualitatively change the observed behavior. Ceilometer data, which are reported every 30 seconds, are averaged over the coincident 5-minute intervals to obtain the average cloud base height. Satellite data are available only at half-hour intervals; the nearest

observation to each MWR time is used. The biggest limitation is the radiosonde, which is often available only once a day and must be used for all observations during a 24-hour period when this is the case.

Another limitation is the uncertainty involved in microwave liquid water retrievals. Histograms of MWR liquid water paths indicate a small but fairly uniform population of points at paths > 0.1 cm, which corresponds to optical thicknesses > 150 for a $10\text{ }\mu\text{m}$ droplet effective radius. These are probably erroneous and are eliminated from the analysis. The microwave also has difficulty with small liquid water paths because of varying water vapor contents, cloud temperatures, etc., that are not adequately taken into account in the retrieval algorithm; negative liquid water paths are possible. The frequency distribution of liquid water paths at the SGP site for observation times at which the ceilometer indicates clear skies is centered near zero but with a standard deviation of 0.004 cm; we thus eliminate points below this threshold as well. Neither cutoff qualitatively affects the results we present here.

Unfortunately, difficulties in the ARM Programs ability to retrieve data from the ARM archive, combined with the satellite data being taken offline since mid-1994 for revision, have prevented us from obtaining coincident data from all four instruments for even a single full month thus far. To get started, we have therefore focused on June 1993, for which all instruments but the ceilometer are available, and have explored several approximations to substitute for an actual cloud base observation. The simplest assumption is that cloud base does not vary systematically with temperature. We thus define a proxy cloud thickness as the pressure difference between the surface and cloud top, and define the mean cloud temperature accordingly. Figure 1 (left) shows the resulting variations of liquid water path with temperature, both for individual points (top) and binned into 1 K temperature intervals (bottom). Liquid water path clearly exhibits a tendency to decrease with increasing cloud temperature, consistent with the ISCCP optical thickness behavior in midlatitude summer. To see whether this is caused by liquid water content or cloud physical thickness variations, we normalize the liquid water path by the proxy pressure thickness and plot it versus temperature (Figure 1, right). This quantity increases with temperature at a rate of about $0.05/\text{K}$, somewhat larger than but qualitatively consistent with the theoretical rate of increase of cloud liquid water content for a lifted parcel. This suggests that water content changes are close to adiabatic and the observed satellite optical thickness behavior can be explained by a systematic decrease of cloud physical

thickness with temperature. But other explanations are possible, depending on cloud base behavior.

To try to estimate cloud base variations in the absence of the ceilometer data, we use radiosonde relative humidity profiles to identify cloud base. Sensitivity tests for other months in which the ceilometer data are available suggest that a relative humidity threshold of 94% results in the highest percentage of reasonably accurate identifications of cloud base. With this approximate definition of cloud base, we examine several different hypotheses for the cause of liquid water path variations: 1) cloud thickness variations dominated by cloud top changes, 2) cloud thickness variations caused by cloud base variations, and 3) constant cloud thickness, but systematic changes in mean cloud pressure (and temperature, implying changes in liquid water content). Preliminary results are shown in Figure 2. Liquid water path decreases with increasing cloud top pressure, but only for top pressures > 800 mb. On the other hand, liquid water path increases with cloud base pressure to some extent, but only for base pressures < 875 mb. One possible explanation is that different physics operate for low clouds within and above the planetary boundary layer. If anything, there is a slight tendency for liquid water paths to *decrease* with increasing mean cloud pressure and pressure thickness, just the opposite of expectations.

Further examination reveals, though, that our results are compromised by the necessity of using the radiosonde to estimate cloud base. Data from the 5-month period September 1994 - January 1995, when ceilometer data exist, show a clear decrease of liquid water path with increasing cloud base height. But when cloud base is estimated from the radiosonde, there is no obvious dependence on cloud base. This may be due both to inaccuracies in radiosonde relative humidity estimates and to the sparse sampling of radiosonde data, which precludes coincident liquid water path and cloud base estimates for most of the data set. Nonetheless, the preliminary results are encouraging enough to conclude that the ISCCP result is probably realistic and that we will be able to differentiate the roles of liquid water, cloud thickness, and cloud height variations once a statistically significant sample of data from all four instruments is made available. The next phase of the research will address the more difficult question of whether the observed temperature dependence is indicative of the feedback to be expected in a climate change. For this purpose, we will need to investigate the large-scale conditions that give rise to the observed correlations (e.g., static stability, synoptic scale pressure variations, aerosol variations). Ultimately, we will have to document the role of cloud droplet size variations as well.

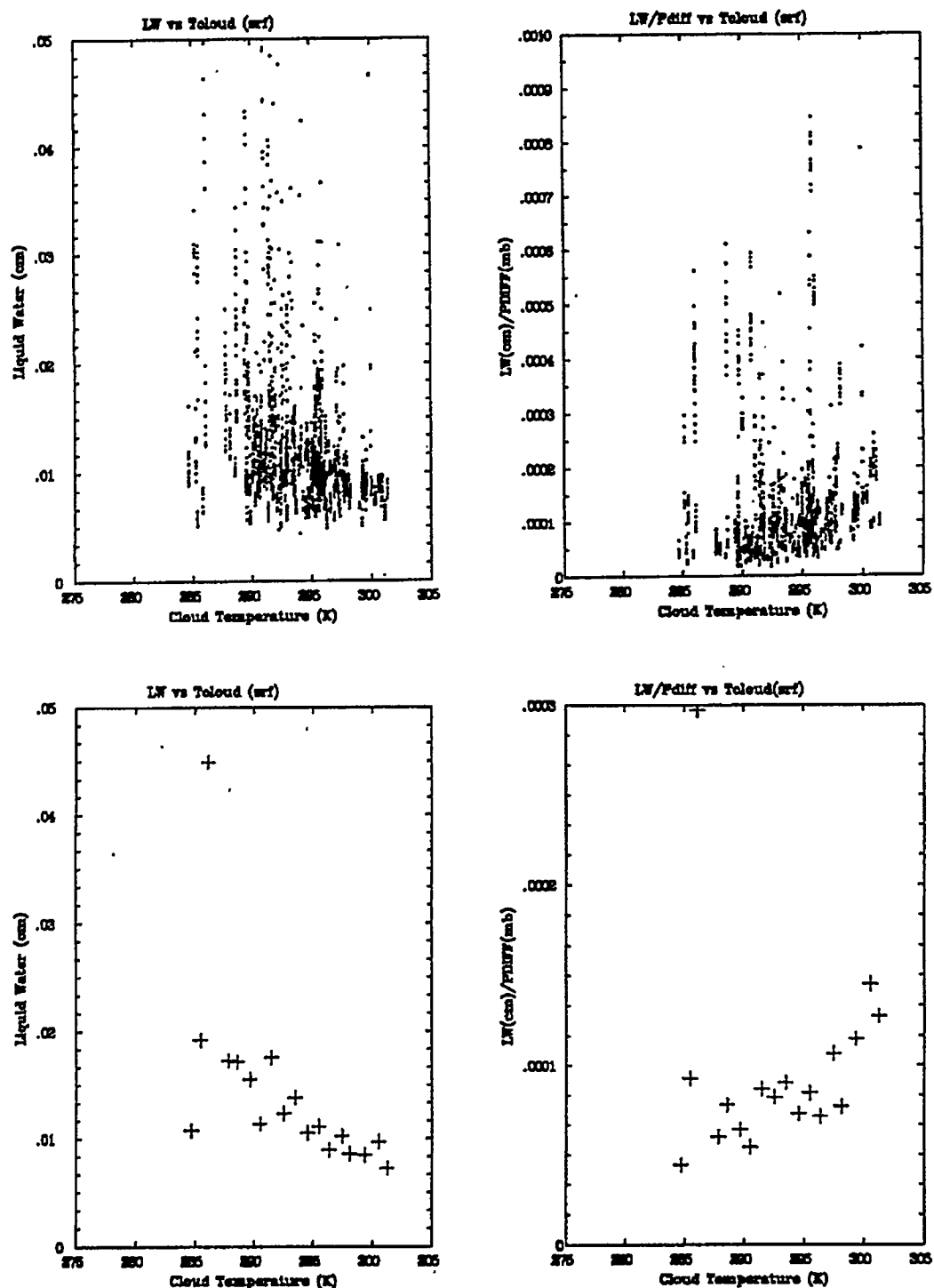


Figure 1. June 1993 MWR liquid water path variations as a function of cloud temperature (left), plotted as individual observations (top) and binned into 1 K temperature intervals (bottom), and analogous plots of liquid water path divided by proxy cloud pressure thickness (right).

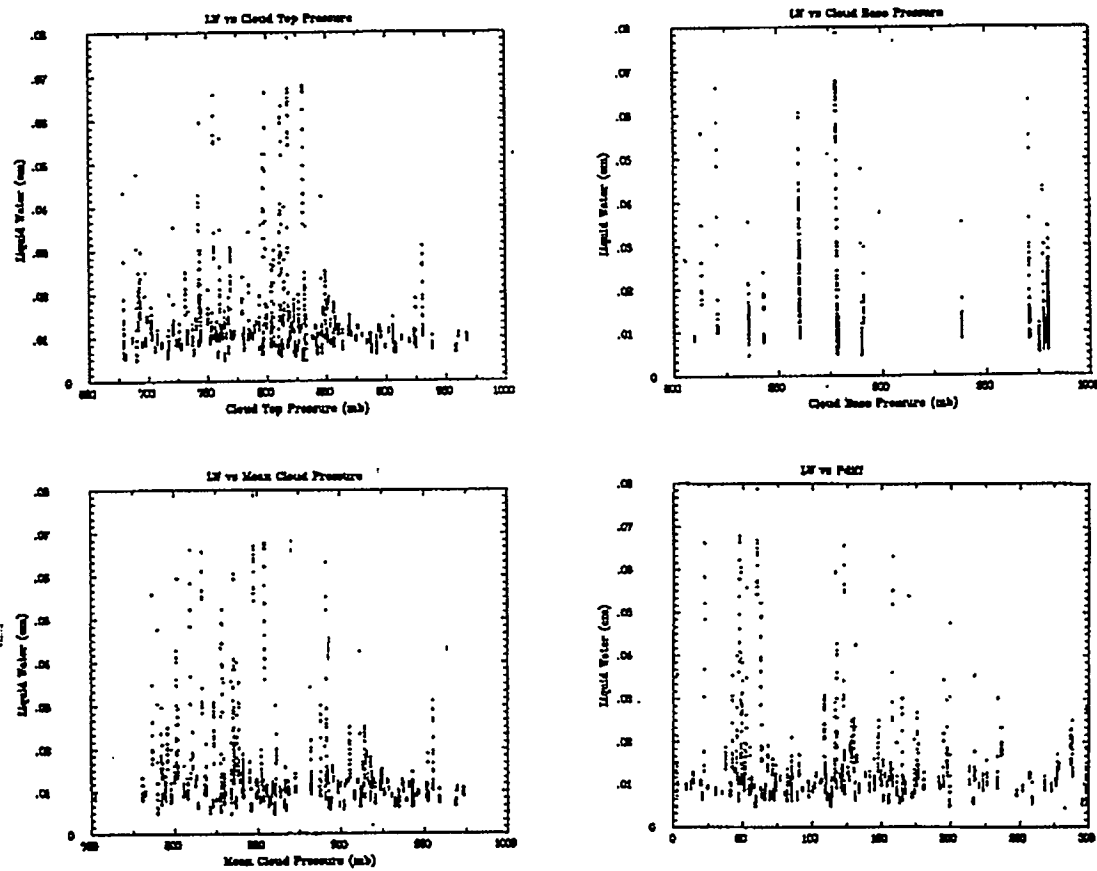


Figure 2. June 1993 MWR liquid water path vs. cloud top pressure (upper left), cloud base pressure (upper right), mean cloud pressure (lower left), and cloud pressure thickness (lower right). Cloud base is identified from radiosonde relative humidity profiles.

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Heterogeneous Surface Fluxes and Their Effects on the Southern Great Plains (SGP) Cloud and Radiation Testbed (CART) Site

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Introduction

The treatment of subgrid-scale variations of surface properties and the resultant spatial variations of sensible and latent heat fluxes have received increasing attention in recent years. Mesoscale numerical simulations of highly idealized conditions, in which strong flux contrasts exist between adjacent surfaces, have shown that under some circumstances the secondary circulations induced by land-use differences can significantly affect the properties of the planetary boundary layer (PBL) and the region of the atmosphere above the PBL. At the Southern Great Plains (SGP) Cloud and Radiation Testbed (CART) site, the fluxes from different land-surface types are not expected to differ as dramatically as those found in idealized simulations. Although the corresponding effects on the atmosphere should thus be less dramatic, they are still potentially important. From an Atmospheric Radiation Measurement (ARM) perspective, in tests of single column models (SCMs) it would be useful to understand the effects of the lower boundary conditions on model performance. We describe here our initial efforts to characterize the variable surface fluxes over the CART site and to assess their effects on the PBL that are important for the performance of SCMs.

Fluxes Over the CART Site

The SiB2 model (Sellers et al. 1986) was chosen to help produce a flux map of the CART site. The SiB2 model makes use of soil, plant, and meteorological data to calculate the sensible and latent heat fluxes over vegetated or bare surfaces. Estimates of leaf area index for vegetation are obtained from satellite-derived Normalized Difference Vegetation Index values, while meteorological fields are computed using data from the Oklahoma Mesonet, National

Weather Service stations, the Kansas State University Mesonet, and the CART central and extended sites. The multiquadric basis function interpolation scheme of Nuss and Titley (1994) is used to obtain values for precipitation, temperature, humidity, incoming solar radiation, and wind speeds over the CART domain, and these variables are then mapped onto an array of 50 x 58 grid cells, each 6.25 km on a side. A new set of fields is computed every 30 minutes for the time periods of interest.

Initial soil moisture values are computed using the approach of Liston et al. (1993). The SiB2 model is then run for each grid cell in the array to obtain the sensible and latent heat fluxes at each grid point. If there are several principal vegetation types found in a given grid cell, SiB2 can be run for each vegetation type, and the net fluxes are then computed from a weighted average of the contributions from each type. An example of the fluxes computed by this methodology is shown in Figure 1. The gridded flux values are then used to obtain a time series of spatially averaged fluxes over the domain, at resolutions ranging from $(6.25 \text{ km})^2$ on up to domain-wide averages.

Such flux calculations are subject to several sources of uncertainty. The largest appears to be the difficulty in specifying the soil moisture profiles accurately for the whole site. At this time, soil moisture measurements are quite limited. Although this situation will improve over the next two years with the addition of soil moisture probes at the extended sites, soil moisture descriptions are expected to remain problematic. A second difficulty is determining the accuracy of the interpolated fields, particularly for variables such as solar radiation, which can change rapidly and irregularly under partly cloudy conditions. A third problem is the specification of the plant characteristics on a fine

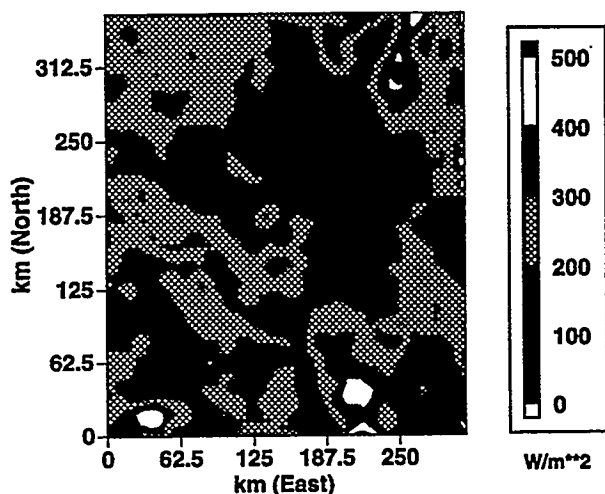


Figure 1. SGP CART sensible heat fluxes derived from the SiB2 model for July 28, 1994.

enough scale, and with sufficient differentiation among the various plant types, to allow SiB2 methods to compute fluxes accurately.

Some degree of tuning is possible by comparing modeled flux values with fluxes measured at the extended sites and making some adjustments in the plant parameters or soil moisture values to obtain improved agreement. This procedure must be used with care, however, to avoid forcing agreement between modeled and measured flux values for the wrong reason. To date, flux measurements have not been available from wheat fields, which constitute a major crop in the CART site, but this situation is also expected to improve substantially by late 1995.

Mesoscale Modeling

Using data from the 1992 Boardman experiment (Doran et al. 1995) and applying the RAMS (Regional Atmospheric Modeling System [Pielke et al. 1992]) mesoscale model as an analysis tool, it has been shown that spatially varying surface heat fluxes can result in varying mixed-layer depths over the domain (Doran and Zhong 1995). A similar result is anticipated for the CART region. In addition, differences in column-integrated moisture, temperature, and the probability of cloud formation are also expected to be found. To

examine these features, the RAMS model is being used to simulate conditions over the SGP CART site for a number of selected cases.

Initial tests of the RAMS model for this domain do show effects in the PBL arising from spatially varying heat fluxes caused by differences in soil moisture. However, the parameterizations used in Version 3a of RAMS show a sensitivity of sensible and latent heat fluxes to soil moisture that appears to be greater than that actually found from CART data. As a result, the soil moisture in RAMS must currently be treated as a tuning parameter that can be adjusted to produce flux fields similar to those obtained from the SiB2 modeling exercise described above. Figure 2 shows the variations in mixed-layer depth computed for a flux distribution similar but not identical to that given in Figure 1. The variations across the domain are significant and are not currently accounted for in SCMs.

Acknowledgment

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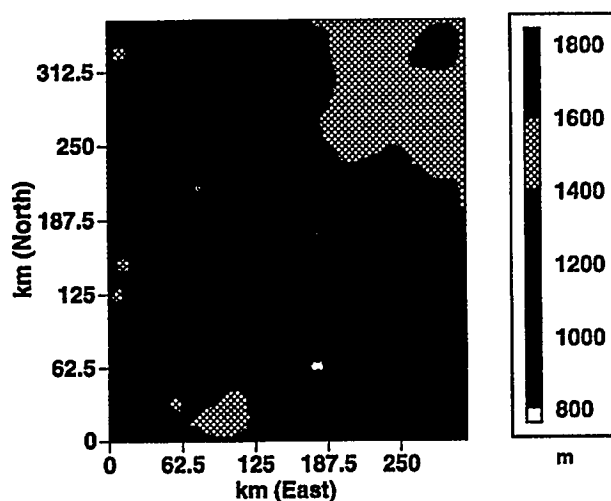


Figure 2. Mixing heights derived from the RAMS model for fluxes similar to those in Figure 1.

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Data Assimilation of a Ten-Day Period During June 1993 Over the Southern Great Plains Site Using a Nested Mesoscale Model

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Introduction

A goal of the Atmospheric Radiation Measurement (ARM) Program has been to obtain a complete representation of physical processes on the scale of a general circulation model (GCM) grid box in order to better parameterize radiative processes in these models. Since an observational network of practical size cannot be used alone to characterize the Cloud and Radiation Testbed (CART) site's 3D structure and time development, data assimilation using the enhanced observations together with a mesoscale model is used to give a full 4D analysis at high resolution.

The National Center for Atmospheric Research (NCAR)/Penn State Mesoscale Model (MM5) has been applied over a ten-day continuous period in a triple-nested mode with grid sizes of 60, 20 and 6.67 in. The outer domain covers the United States' 48 contiguous states; the innermost is a 480-km square centered on Lamont, Oklahoma. A simulation has been run with data assimilation using the Mesoscale Analysis and Prediction System (MAPS) 60-km analyses from the Forecast Systems Laboratory (FSL) of the National Ocean and Atmospheric Administration (NOAA). The nested domains take boundary conditions from and feed back continually to their parent meshes (i.e., they are two-way interactive). As reported last year, this provided a simulation of the basic features of mesoscale events over the CART site during the period 16-26 June 1993 when an Intensive Observation Period (IOP) was under way.

The MME Model

The MME Model has the following characteristics:

- nonhydrostatic, compressible, with terrain-following coordinates and polar-stereographic map projection

- prognostic equations for wind components, vertical velocity, pressure perturbation, temperature, water vapor, ground temperature, and microphysical water and ice content variables
- B-grid staggering, upper radiative boundary condition, relaxation lateral boundary conditions, second-order centered spatial and leapfrog temporal differencing with short step for sound-wave terms
- microphysics with cloud, rain, snow/graupel, and ice processes on all domains' resolved scales
- Grell single-cloud mass flux cumulus parameterization scheme on 60- and 20-km domains
- Blackadar high-resolution planetary boundary layer with four stability regimes, five layers in lowest kilometer. Implicit vertical diffusion above boundary layer
- atmospheric longwave and shortwave radiation scheme interacting with model clouds and land surface
- land surface energy budget calculation to predict ground temperature.

Overview of Results

Comparisons of model data with surface observational data have been made. This has pointed to some improvements in model physics that were required to adequately represent cloud cover and, hence, diurnal heating.

One area of scientific interest is a nocturnal jet on 24 June 1993 that was simulated. The model captured well the low-level wind maximum observed with NCAR's Integrated Sounding System (ISS) 915-MHZ profiler.

Further work has involved acquiring surface and upper-air standard observations and profiler data from NOAA's Demonstration Network and ARM's CART site, and preparing these data for input to the model. A second simulation will assimilate these data by the "station-nudging" technique on the inner two domains. Comparisons with observations will then be made to evaluate the impact of the added data.

Cloud Cover and Ice Sedimentation

From comparison of the model temperature 40 m above ground (short-dash line in Figure 1) at the center of the 6.67-km domain versus 3-hourly analyses based on surface observations at that point (long-dash line), it can be seen that the diurnal heating was severely underestimated, particularly on 16, 19, and 22 June.

This was traced to excessive cirrus cloud cover in the model. Clouds originating over the Rockies nearly every afternoon were advected over Oklahoma, reducing the surface diurnal heating there the next day through the model's cloud/radiation interaction.

The solution was to add sedimentation of ice particles (following Heymsfield and Donner 1990), thus allowing the cirrus clouds to dissipate more realistically. The temperature trace is improved (solid line) with this modification in a rerun of the ten-day simulation.

Also noted was the significant reduction in integrated cloud water over the domain. This is an improvement in areal coverage that is vital to making realistic simulations over

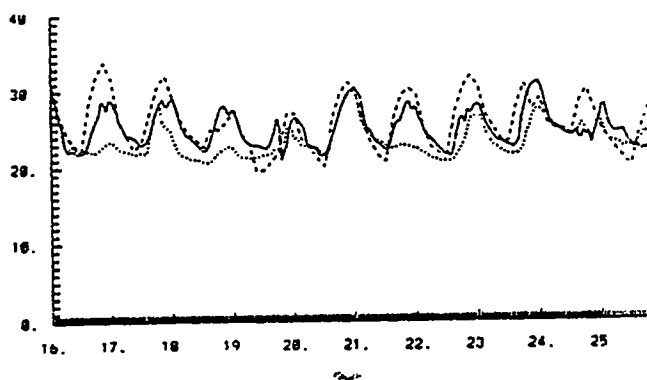


Figure 1. Temperature trace (degrees C versus June date), surface observation (long dash), model without (short dash) and with (solid) sedimentation.

cloud-cover forecasts in an independent wintertime study with MM5 where model output was compared in detail with satellite pictures (K. Manning and C. Davis, personal communication).

The Nocturnal Jet

A nocturnal jet is a low-level wind maximum that may reach $20\text{--}30\text{ m s}^{-1}$ within a kilometer of the ground. It is typically found over the SGP, making the CART site an ideal location for its study. It is likely to be important in the large-scale moisture transport of the region.

Its formation is believed to result from the collapse of the daytime boundary layer which decouples the ageostrophic wind from the frictional layer that was responsible for it. Upon decoupling, the ageostrophic component turns clockwise through Coriolis forces.

In situations favorable for the nocturnal jet, the geostrophic wind has a southerly component, and so the ageostrophic wind has an easterly component during the day that rotates to southerly at night, making the wind supergeostrophic, and hence the jet is formed. The role of the gradual westward terrain slope of the region in aiding the ageostrophic upslope flow is also probably important.

For the model to reproduce this sequence accurately requires 1) an adequate representation of the synoptic-scale flow, 2) proper boundary-layer formation during the day, 3) proper collapse of the boundary layer at night, and 4) a good representation of the terrain.

The ISS 915-MHZ profiler (Figure 2) showed a clear nocturnal jet development after 00Z 24 June. Winds reached 25 m s^{-1} in the lowest kilometer and turned slowly clockwise with time.

A time sequence of the model wind at about 1 km above ground also shows the nocturnal jet reaching a maximum of 25 m s^{-1} and turning clockwise as it accelerates and becomes more parallel to the isobars by 06Z (Figure 3) than earlier at 00Z. The level of the jet is slightly higher than observed, but the feature's time development is reproduced well.

Boundary Layer Development in MM5

The soundings taken at the center of the domain near Lamont, Oklahoma, every 6 hours from 00Z 24 June till 00Z 25 June, also show the jet's development and boundary

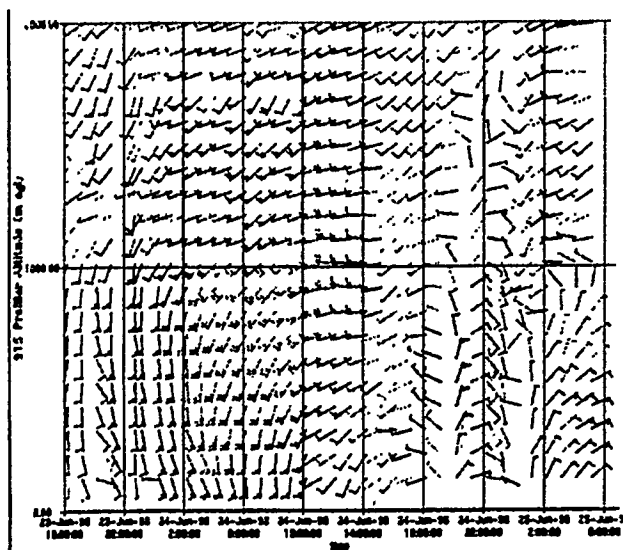


Figure 2. 915-MHZ ISS profiler height versus time plot for 0 to 2 km altitude between 18Z 23rd and 06Z 25th June 1993. Wind barbs in m/s.

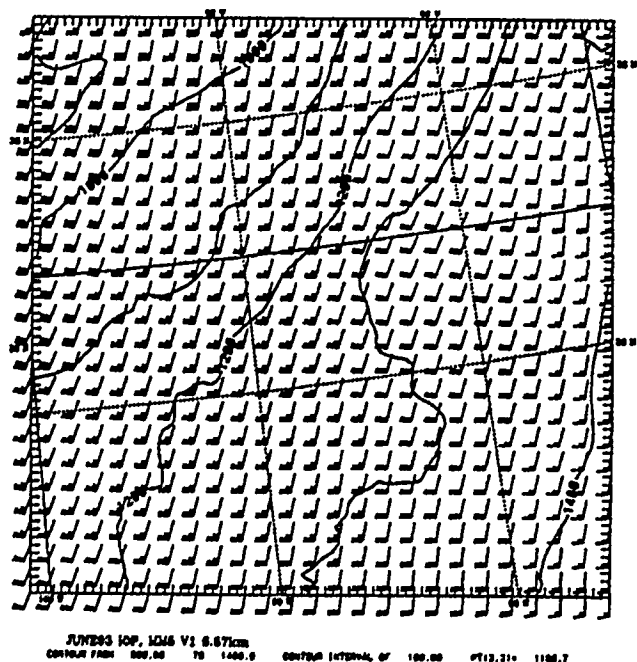


Figure 3. Wind at about 1 km above ground for model 6.67-km domain at 06Z 24th June 1993. Wind barbs in m/s. Pressure contoured with 1 hPa interval.

layer's collapse in the model, followed by the influence of convection at the cold front that moved into the region.

The daytime boundary layer is seen at 00Z 24 June (Figure 4a), but is gone by 06Z (Figure 4b) because of surface night-time cooling, and the jet becomes evident. By 12Z there is an inversion and the jet continues to rotate. The convection causes a near moist-adiabatic structure by 18Z (Figure 4c), followed by a drying by 00Z 25 June as a result of convective downdrafts. An accompanying paper (see Dudhia and Parsons in this volume) shows the rainfall associated with this front and the radar summary.

The realistic model soundings are encouraging particularly since this period represents the ninth of the ten days assimilated. It is clear that MM5 is capable of maintaining a reasonable simulated atmospheric state without exhibiting a "climate drift" despite being only controlled by information beyond the boundaries of the 20-km domain, 600 km from the northern Oklahoma ARM site.

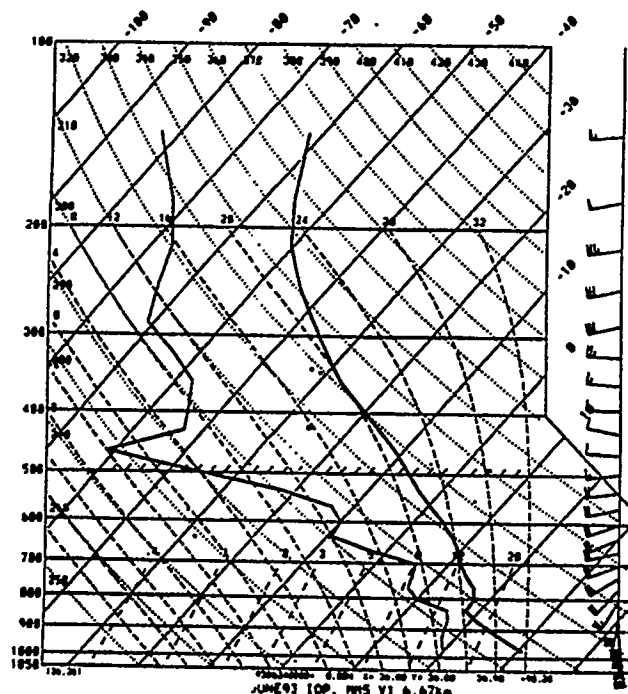


Figure 4a. Sounding in center of 6.67-km domain at 00Z 24 June 1993.

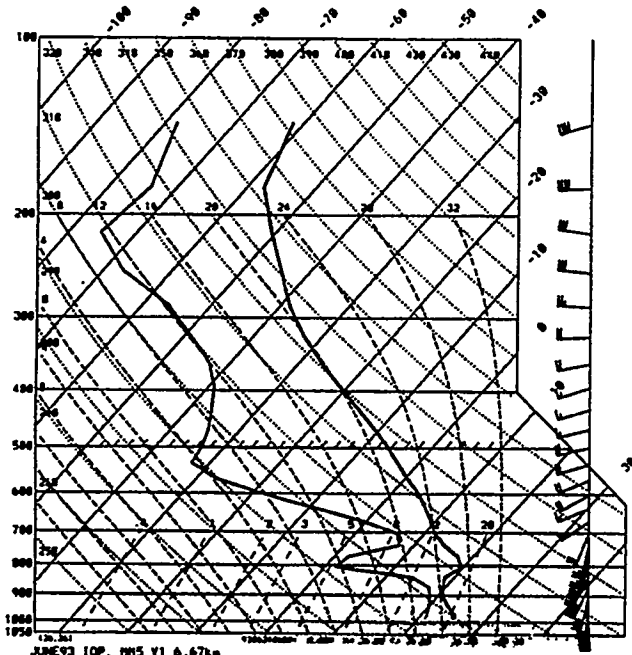


Figure 4b. As Figure 4a but for 06Z.

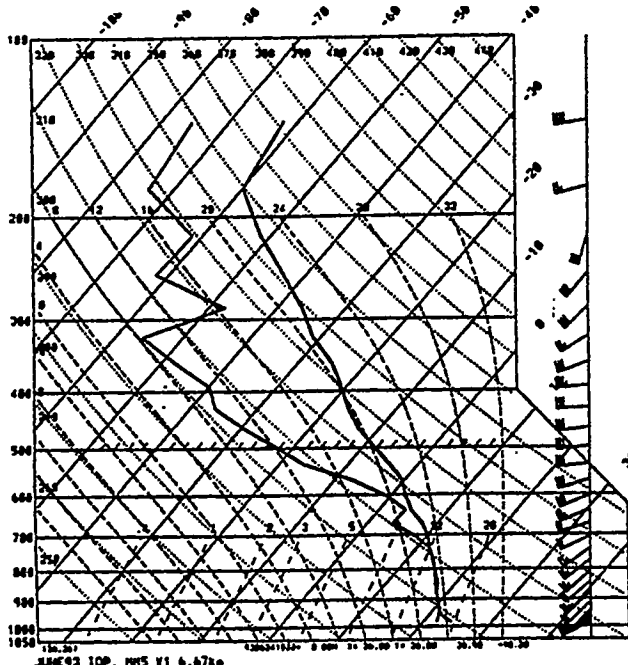


Figure 4c. As Figure 4a but for 18Z.

Further Work: Full Data Assimilation Run

With the recent availability of processed profiler data from the two 915-MHZ profilers at the CART site, there is now a complete dataset ready for assimilation into MM5 in a full Data Assimilation mode. The inner two meshes contain much data that are directly assimilatable into MM5, including enhanced NWS CLASS soundings taken 6-hourly at Topeka, Kansas; Dodge City, Kansas; and Norman, Oklahoma, around the CART site during this IOP. The balloon data is modified to allow for an estimated balloon drift and time lag, which are important considerations in the fine grid.

The NOAA Demonstration Network of 405-MHZ wind profilers (Figure 5) provided data that will be input to MM5. These data were quality controlled during this past year, and some effort has gone into determining an objective algorithm for this large quantity of data. The applied algorithm checks the data against neighboring levels and nearby stations (within 400 km). It provides hourly averaged wind profiles through the troposphere. Previous studies including Kuo and Guo (1989) and Dudhia (1993) have shown the value of a broad coverage of wind data in allowing retrieval of the thermal field.

Additionally, dense standard surface observations are used to provide an analysis towards which the model will be nudged.

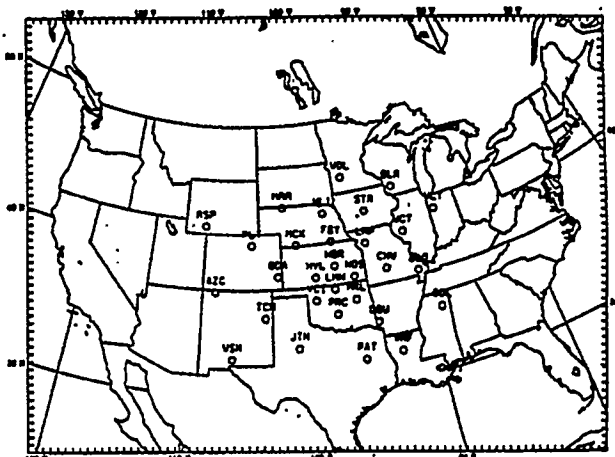


Figure 5. NOAA Demonstration Network of 405-MHZ profilers.

Supplemental balloons were launched from three locations in the CART site during this IOP that can be used as independent verification of the model's assimilation effectiveness since these data are not being assimilated.

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Comparison of Mean Properties of Simulated Convection in a Cloud-Resolving Model with Those Produced by Cumulus Parameterization

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Introduction

An Intensive Observation Period (IOP) of the Atmospheric Radiation Measurement (ARM) Program took place at the Southern Great Plains (SGP) Cloud and Radiation Testbed (CART) site from June 16-26, 1993.

The National Center for Atmospheric Research (NCAR)/Penn State Mesoscale Model (MM5) has been used to simulate this period on a 60-km domain with 20- and 6.67-km nests centered on Lamont, Oklahoma. Simulations are being run with data assimilation by the nudging technique (Kuo and Guo 1989; Stauffer and Seaman 1990) to incorporate upper-air and surface data from a variety of platforms. The model maintains dynamical consistency between the fields, while the data correct for model biases that may occur during long-term simulations and provide boundary conditions. For the work reported here the Mesoscale Atmospheric Prediction System (MAPS) of the National Ocean and Atmospheric Administration (NOAA) 3-hourly analyses were used to drive the 60-km domain while the inner domains were unforced. A continuous 10-day period was simulated.

Overview

One goal of the ARM Program is to improve general circulation models (GCMs) by obtaining detailed meteorological information in limited areas of under 200 km square and comparing GCM parameterizations with the mean radiative and convective properties in such areas. Typical GCM grid boxes are 100-200 km square, but there is in reality much structure at smaller scales that is

represented by their parameterizations. Meteorological observations alone cannot represent this structure, so we use a full-physics mesoscale model forced by large-scale tendencies to give as complete a picture of the sub-100-km scale structures as possible.

Here we will concentrate on one day (June 24) when a cold front entered the domain. The convection associated with this front is simulated at several resolutions and with various cumulus schemes on the climate-scale model resolution to evaluate whether these schemes can represent the model behavior at high resolution where the clouds are resolved.

The MM5 Model

The model features and options used in this study are as follows. Equations are for nonhydrostatic, compressible motion, in terrain-following coordinates with a polar-stereographic map projection. Prognostic equations exist for wind components, vertical velocity, pressure perturbation, temperature, water vapor, ground temperature, and microphysical water and ice content variables. It has an upper radiative boundary condition, relaxation lateral boundary conditions, and interactive two-way nesting.

The model includes microphysics with cloud, rain, snow/graupel, and ice processes on all domains' resolved scales. The Grell cumulus parameterization scheme is adopted only on the 20-km and coarser domains. The Blackadar high-resolution planetary boundary layer and a surface energy budget calculation are used. There is also an atmospheric longwave and shortwave radiation scheme interacting with model clouds and land surface.

Domain and Simulations

Figure 1 shows the areas covered by the three MM5 domains. The 60-km domain coincides with the MAPS domain. The inner 6.67-km domain is centered on a profiler hexagon of the demonstration network around Lamont, Oklahoma, and covers a 480-km square.

Concentrating on 24 June, a series of test simulations was run to compare convection at several grid sizes. A simulation in which the 6.67-km domain was switched off at 00Z/24th was run 24 hours with the Grell scheme with 20-km grid size, and similarly one was run at just 60-km resolution. A further pair of simulations used 60-km initial conditions at 00Z/24th degraded to 180-km grid size, using 1) the Grell scheme, and 2) the CCM2 convective adjustment scheme.

However, 6.67 km is probably not sufficient to resolve updrafts well, so another experiment was run from 12Z/24th for 12 hours at 2.22 km nested in the center of the 6.67-km domain. This serves as a test of the explicit 6.67-km simulation, and the highly resolved cloud structures in a 160-km square will be compared to the representations on coarser meshes. Here we will present the mean effects of convection on the 160-km scale.

The Cold Front of June 24, 1993

Figure 2a shows a cold front across Oklahoma/Kansas around 18Z 24 June 1993. This front had moved into

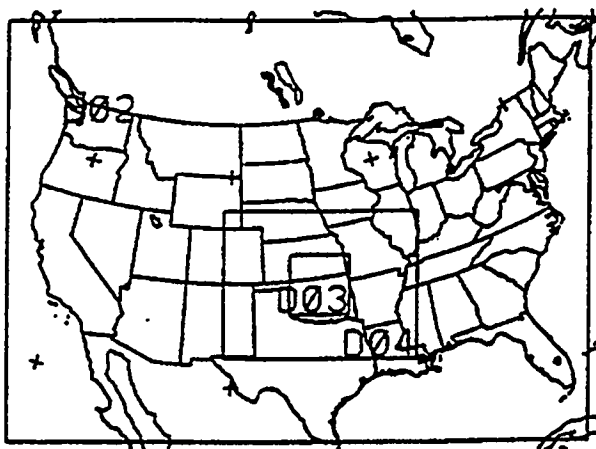


Figure 1. Domains for the simulations (60 km, 20 km, 6.67 km).

Oklahoma over the previous 6-12 hours and became stationary over the CART site. The radar summary (Figure 2b) shows that convergence ahead of the front was capable of triggering convection over northern Oklahoma and southern Kansas. The modeled surface winds and temperature at 18Z on June 24 are shown on Figure 3.

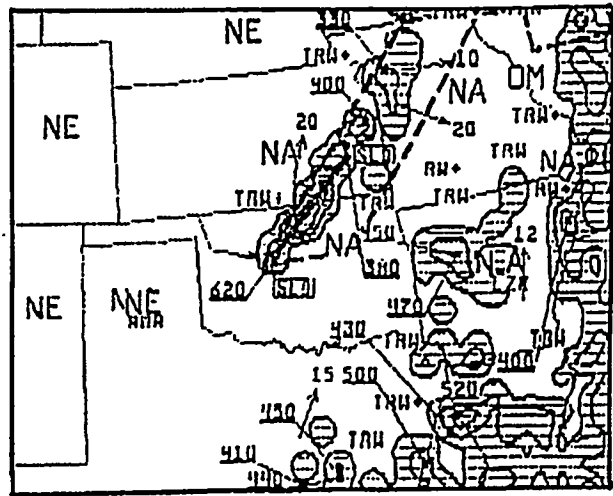


Figure 2a. Radar summary for 1935Z 24 June 1993.

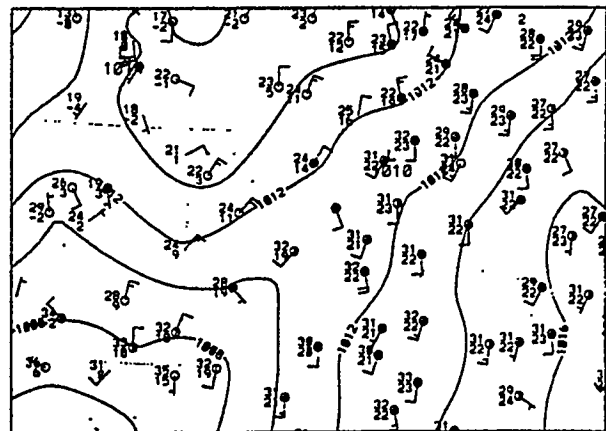


Figure 2b. Surface pressure analysis, wind and temperature observations for 18Z 24 June 1993. Contour interval 2 hPa, wind barbs in m/s.

Figures 5a and 5b show the high-level apparent heat source and lower apparent moisture sink during the convective stage with the signs reversing in a dissipative stage around 20Z, when evaporation exceeds condensation in the domain

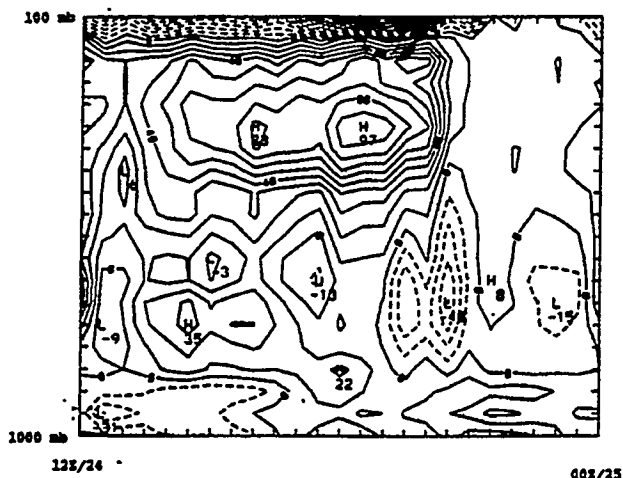


Figure 5a. Pressure versus time plot of the apparent heat source, Q_1 , between 12Z 24 June and 00Z 25 June 1993, for the 2.22-km domain. Contour interval 10^{-5} K/s.

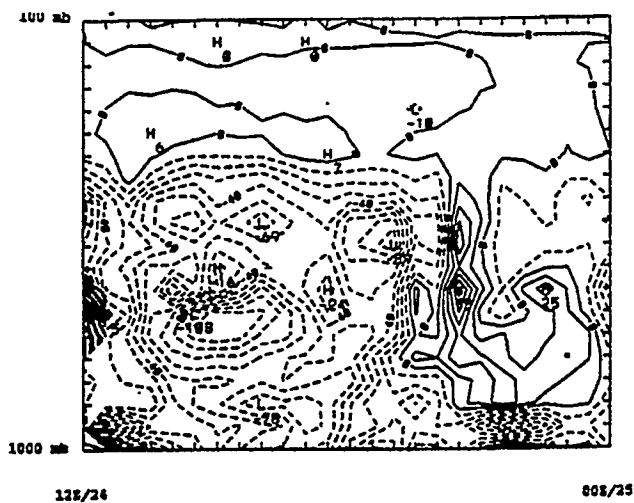


Figure 5b. Pressure versus time plot of the apparent latent heat source, (equation) Q_2 , between 12Z 24 June and 00Z 25 June 1993, for the 2.22-km domain. Contour interval 10^{-5} K/s.

average. Maximum values correspond to 10 K/d in sensible and latent heating due to the convective processes in the area.

Mean Mass Fluxes of the Simulations

Figure 6 shows how the mean mass flux, taken over the area of the 2.22-km domain, decreases as the grid size gets larger. This is consistent with the reduced rainfall rates at low resolutions. Figure 6a shows that the mean mass flux over the 12-hour period, $M = (\text{equation})$, reaches nearly $0.2 \text{ kg m}^{-2} \text{ s}^{-1}$ for the 2.22-km domain average, but at 60 km (Figure 6b), it is only about $0.04 \text{ kg m}^{-2} \text{ s}^{-1}$ averaged over the same area.

This emphasizes that to evaluate convective parameterizations, one must use cloud-resolving models because it is only at high resolution that some convection is triggered, while the current cumulus schemes underestimate vertical mass flux and rainfall as a result of failing to trigger.

In this case, there was a stable layer ahead of the front that probably inhibited the parameterized convection, whereas the resolved lifting at fine resolution was sufficient to overcome this inhibition.

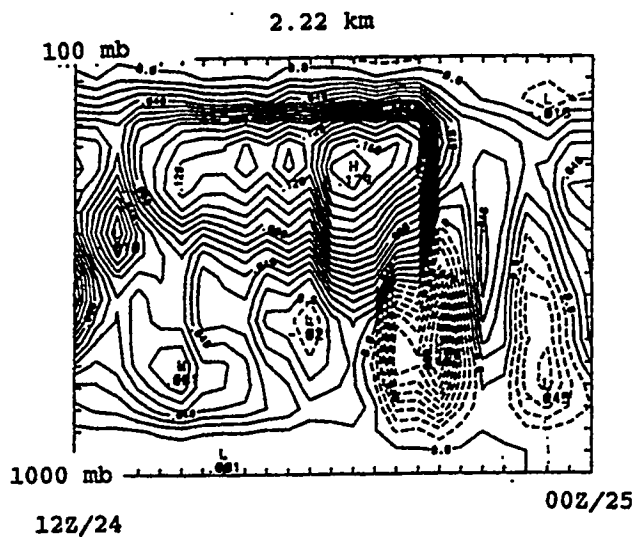


Figure 6a. Pressure versus time plot of mean mass flux, M , in 2.22-km domain, between 12Z 24 June and 00Z 25 June 1993. Contour interval $0.01 \text{ kg m}^{-2} \text{ s}^{-1}$.

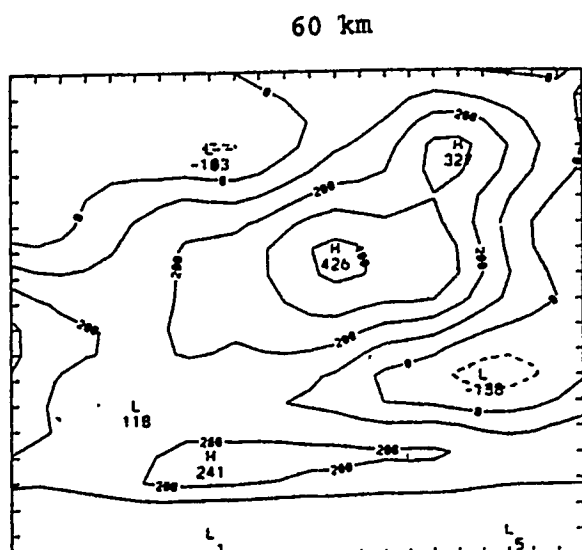


Figure 6b. Pressure versus time plot of mean mass flux, M -, in 60-km domain, between 12Z 24 June and 00Z 25 June 1993. Contour interval $0.01 \text{ kg m}^{-2} \text{ s}^{-1}$.

Further Work

Further work in this area will include the following:

1. Enhance high-resolution simulations with assimilation of more local data (e.g., profiler network). This could allow a more accurate simulation of the rainfall pattern associated with the front.
2. Evaluate cloud fractional coverage versus that predicted in GCMs' radiation schemes. Determine mean radiative properties of resolved cloud fields.
3. Evaluate mean boundary-layer fluxes in high-resolution models and compare with GCMs.
4. Improve cumulus parameterization such that it can reproduce a cloud-resolving model's rainfall/mass flux, possibly by modifying the scheme's initiation conditions or by trying a variety of schemes.

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Integrated Cumulus Ensemble and Turbulence (ICET): An Integrated Parameterization System for General Circulation Models (GCMs)

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Project Statement

Successful simulations of the global circulation and climate require accurate representation of the properties of shallow and deep convective clouds, stable-layer clouds, and the interactions between various cloud types, the boundary layer, and the radiative fluxes. Each of these phenomena plays an important role in the global energy balance, and each must be parameterized in a global climate model. These processes are highly interactive.

One major problem limiting the accuracy of parameterizations of clouds and other processes in general circulation models (GCMs) is that most of the parameterization packages are not linked with a common physical basis. Further, these schemes have not, in general, been rigorously verified against observations adequate to the task of resolving subgrid-scale effects.

To address these problems, we are designing a new Integrated Cumulus Ensemble and Turbulence (ICET) parameterization scheme, installing it in a climate model (CCM2), and evaluating the performance of the new scheme using data from the Atmospheric Radiation Measurement (ARM) Program Cloud and Radiation Testbed (CART) sites.

Objectives

1. Development of an Integrated Cumulus Ensemble and Turbulence model for global models: ICET
2. Analysis of data from the ARM Southern Great Plains (SGP) site to provide driving terms and validation for the ICET scheme.
3. Concomitant analysis of data from the Tropical Western Pacific (TWP) site to provide driving terms and validation in a tropical regime.

ICET Design

Overview

The cumulus parameterization consists of a cloud ensemble model that represents the lifecycles of cumulus clouds, coupled to a level 2.5 turbulence scheme. The cumulus cloud ensemble model is initialized with information from the turbulence scheme when conditions at any vertical level are sufficiently unstable to produce buoyant, saturated vertical drafts. These cumulus ensembles simulate flues of quantities resulting from eddies that penetrate vertical distances greater than one model level.

In ICET, the closure is specified in terms of the cloud ensemble mass flux at cloud base. The assumption here is that the cloud ensemble will maintain an equilibrium of the stability of the layers immediately above cloud base. Increasing the cloud base mass flux warms and stabilizes this air, inhibiting further convection, while grid-scale uplift, radiative cooling, and other large-scale processes tend to cool and destabilize air in this layer. Coupling the cumulus parameterization directly to the turbulence scheme provides a mechanism for linking the host climate model's boundary layer parameterization, shallow and deep convection, and diffusion into a single, physically consistent package. Finally, the ICET package is fully interactive with the stable cloud parameterization and provides important information to the radiative transfer parameterization. A schematic of ICET is shown in Figure 1. CCM2 is available in single column or fully three-dimensional time dependent forms. Implementation and testing of the new ICET parameterization package in both configurations of CCM2 is one goal of the work here. The Pennsylvania State University Department of Meteorology has been chosen as a β -test site for the CCM2 single column model. To this end, CCM2 is currently installed on the Department's Sun workstation network.

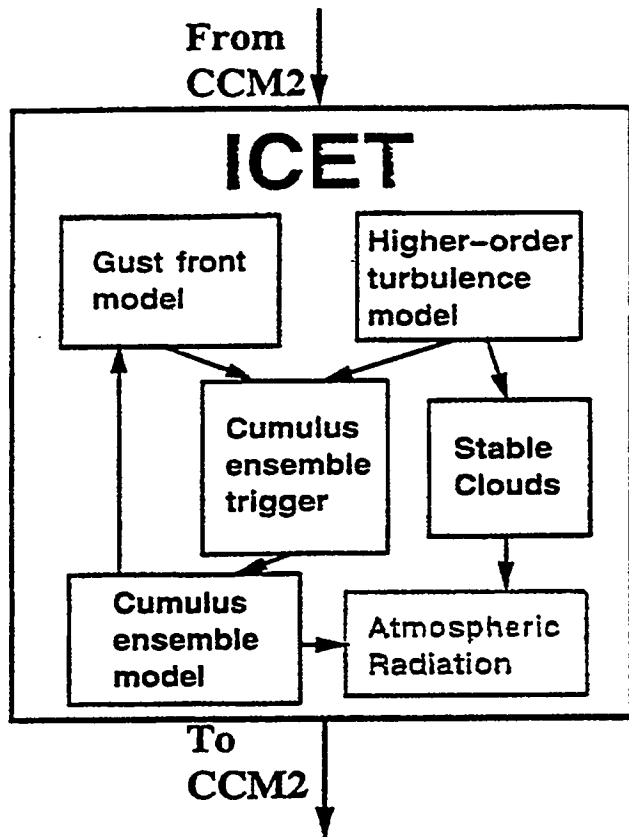


Figure 1. Schematic of the interactions of the ICET parameterization package. Gridscale variables (such as winds, temperature, moisture and surface fluxes) are fed in from the host model (CCM2). Arrows within the ICET box indicate forcings and feedback relationships contained within ICET. The resultant grid-scale forcing effects due to subgrid-scale processes are fed back from ICET to CCM2.

Gust Front Model

Convection is extremely sensitive to variations in boundary layer structure on scales smaller than those resolved by GCMs. From a cumulus parameterization standpoint, the two most important resolution problems are

1. Convection is typically fed by the undisturbed, unstable air that exists outside the disturbed (cooled by downdrafts) regions of the boundary layer. The GCM cannot resolve the disturbed and undisturbed regions and so it mixes them to single values of temperature and moisture in the grid column and, hence, tends to stabilize convectively active regions prematurely.
2. Once deep convection becomes organized, it provides forced uplift of unstable air to depths determined in large part by the depth of the downdraft gust fronts. This has a major effect on the parameterization trigger functions.

To address these two problems, we have designed an embedded gust front model that partitions the GCM boundary layer into disturbed and undisturbed regions. The spreading of cold downdraft air and wake recovery due to enhanced surface fluxes are accounted for in this simple model. This model also provides estimates (distance and vertical velocity) of the lifting of boundary layer air by the moving gust fronts. A schematic of the gust front model is shown in Figure 2.

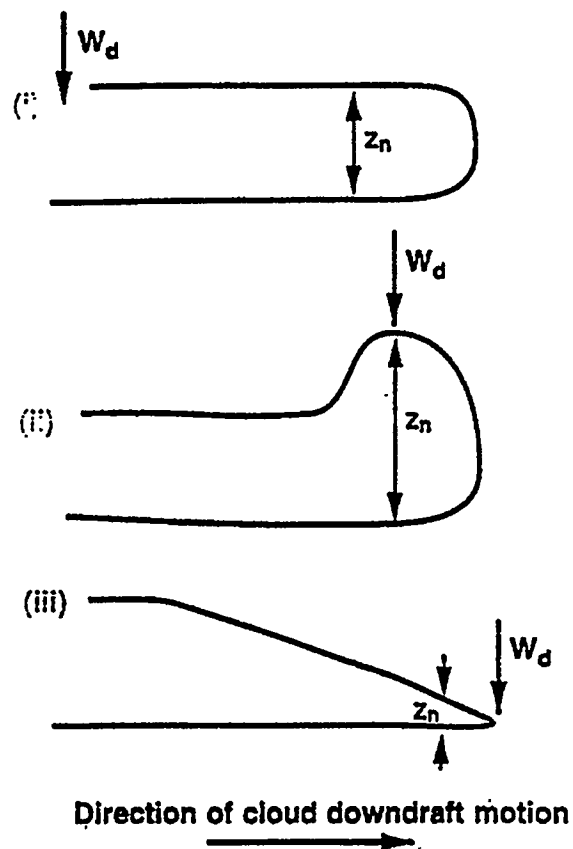


Figure 2. Schematic of gust front structure dependence on storm (downdraft) speed: (i) gust front speed (i.e., the density current) > twice the storm speed; (ii) gust front speed = twice storm speed; (iii) gust front speed < twice storm speed.

ICET Model Testing

Since the ICET parameterization is to be used in a global model, it must be validated over a variety of atmospheric regimes. The ARM testbed sites are ideal sources of data for this validation process. Currently, data from the ARM SGP site are being analysed to provide forcing and validation fields for the ICET modeling system (see the companion poster of Flatley and Mace).

Immediately after the data become available, similar analyses are planned on the data from the ARM TWP site. The inclusion of an extra tropical site in a regime with less frequent deep convective activity would provide an additional very useful control for validation of the cumulus

ensemble in the tropics. Data from these two ARM sites are being (or will be) used to run the ICETsystem within the single column version of CCM2 and, later, within the three-dimensional CCM2 global model.

Future Work

Testing of the parameterized gust front is almost completed. The next phase of this research is to couple this gust front model to the ensemble cloud model. Once we are satisfied that this gust front model provides a sensible forcing for the ensemble cloud model, the full system will be assembled by coupling to the higher order turbulence scheme and feeding the relevant inputs to both the stable cloud and atmospheric radiation schemes.

Drizzle Production in Stratocumulus

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Introduction

Although stratocumulus clouds are not prodigious producers of precipitation, the small amounts of drizzle they do produce have an important impact on both cloud macrophysical properties (e.g., spatial coverage, depth and liquid water content) and microphysical properties (e.g., droplet size distributions, effective radii). The radiative effects of stratocumulus are intimately connected to both these macro- and microphysical properties, and it is thus essential that we understand the mechanisms of droplet growth which generate precipitation sized droplets.

Drizzle production is closely related to cloud condensation nucleus (CCN) number and size (see, for example, Albrecht 1989), as well as to cloud dynamics and the ability of clouds to support droplets within their bounds and allow for repeated collision-coalescence cycles (Mason 1952). In order to address both the microphysical and dynamical aspects of drizzle formation (and their close coupling), we have adapted a large eddy simulation (LES) model to include explicit (size-resolving) microphysical treatment of the CCN and droplet spectra (Feingold et al. 1994; Stevens et al. 1995). By directly calculating processes such as droplet growth by condensation and stochastic collection, evaporation, and sedimentation in the LES framework, we are in a position to elucidate the drizzle formation process.

Here we examine only one aspect of drizzle formation, namely, the relationship between it and a measure of the energy of the cloud (in this case root-mean-square velocity). We follow the classic work of Bowen (1950) and Mason (1952), who calculated trajectories of a large collector droplet rising in an updraft and collecting small droplets. Bowen showed that the size of the drop, as it exits the cloud, depends on the updraft velocity and cloud liquid water content. Mason argued that the turbulent motions in stratiform clouds will allow drizzle drops to form by

increasing their dwell-time in the cloud. Nicholls (1987) followed this line of thought by studying stochastic drop coalescence within a stochastic turbulence model. This work differs from the aforementioned works in a number of respects. Firstly, we explicitly model droplet growth processes and the mechanisms through which some statistically fortunate droplets grow large enough to initiate gravitational collection. Secondly, given the fact that these droplets do exist, we examine how important the turbulent motion within the cloud is in determining the amount of precipitation produced. We allow full coupling between microphysics and dynamics in a multi-dimensional framework.

Model Description

To expedite processing of a large number of numerical experiments, we run the model in two-dimensions. We have used a single sounding, taken from the research vessel *Malcolm Baldrige*, June 16, 1992, 07:05 GMT, which produced a solid cloud deck and drizzle at the ground.

We examine the sensitivity of drop growth to cloud vertical velocity by artificially modifying the drop terminal velocities. The approach is justified by the fact that the factor of importance is the velocity of the drops relative to the cloud velocities. The net vertical motion of a droplet is given by the sum of the local updraft velocity w and the drop terminal velocity V_T . Thus, the simulation of a more vigorous cloud with, say, $w \pm \epsilon$, is equivalent (in terms of w_{net}) to the simulation of that cloud with the same w but terminal velocity $V_T \mp \epsilon$. Based on this principle, we can simply add or subtract an amount ϵ to each drop terminal velocity to simulate less or more vigorous clouds. There is, however, a practical problem in the implementation of this approach—when subtracting ϵ from V_T , we find that for $V_T \leq \epsilon$ the applied fall velocity becomes negative, and droplets fall upwards, eventually reaching the dry

above-cloud air where they rapidly evaporate. Thus, we have assumed that all drops with V_T smaller than ϵ move with the cloud motions. (For symmetry, we have applied this rule for all cases.)

We note that application of corrections to fall velocities has the added benefit of providing insight into the extent to which bulk microphysical parameterizations would misrepresent the drizzle process with an erroneous choice of drop terminal velocity within the bounds $V_T - \epsilon$; $V_T + \epsilon$.

Results of the Bowen Model Without Dynamical Feedback

We ran our base case assuming that all droplets with $r < 0 \mu\text{m}$ ($\epsilon = 50 \text{ cm s}^{-1}$) move with the air motions, while drops larger than $60 \mu\text{m}$ fall at their correct V_T . This value of ϵ is somewhat arbitrary, but not restrictive in the sense that results vary only quantitatively for different choices of ϵ . We then repeated this run but subtracted or added an ϵ of 50 cm s^{-1} from V_T ($V_T - \epsilon$, or $V_T + \epsilon$) for $r > 60 \mu\text{m}$. These cases represent a more (or less) vigorous cloud with an enhanced (or diminished) ability to maintain droplets in the cloud. To isolate the effect of the drop terminal velocity relative to cloud updraft velocity, we imposed on the $V_T - \epsilon$ and $V_T + \epsilon$ runs the identical cloud vertical velocities as produced for the V_T run. The differences in drizzle production between these cases are summarized by Figure 1. Drizzle rate (R) for the $V_T - \epsilon$ run is more than double that for the V_T run, while for the $V_T + \epsilon$ run, it is less than half of that in the V_T run. Analysis of radar reflectivity shows that the precipitation reaching the ground in the $V_T + \epsilon$ run is in the form of larger drops, while in the case of the $V_T - \epsilon$ run, it is in the form of small drops. When drops have reduced fall velocities (equivalent to more vigorous clouds), re-circulation in the cloud produces more drizzle, comprising larger drops. The converse is true for the case where drops have enhanced fall velocities (equivalent to less vigorous clouds).

Results of Bowen Model with Dynamical Feedback

We repeated the experiments $V_T - \epsilon$ and $V_T + \epsilon$ in Section 3a but this time allowed feedback to the dynamics (i.e., w was prognosed correctly). Figure 2 shows the same trends for R observed in Figure 1, although the degree of enhancement in R for the $V_T - \epsilon$ run is reduced. Therefore the dominant effect is the fall velocity of the drops relative to cloud vertical velocity, and feedbacks are not sufficient to erase this effect. In Figure 3, we show profiles of the

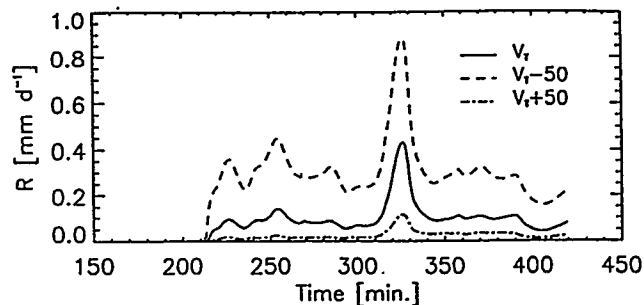


Figure 1. Temporal evolution of R [mm d^{-1}] for the V_T , $V_T - \epsilon$, $V_T + \epsilon$ cases: fixed dynamics.

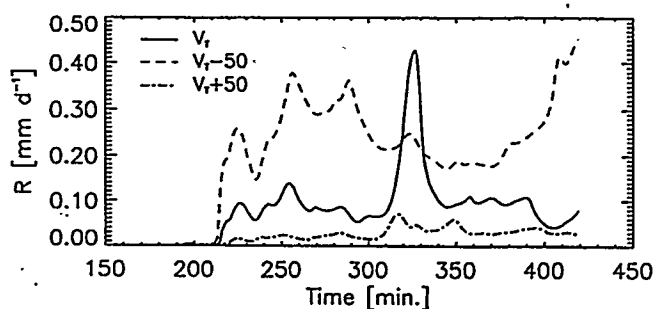


Figure 2. Temporal evolution of R [mm d^{-1}] for the V_T , $V_T - \epsilon$, $V_T + \epsilon$ cases: with feedback to dynamics.

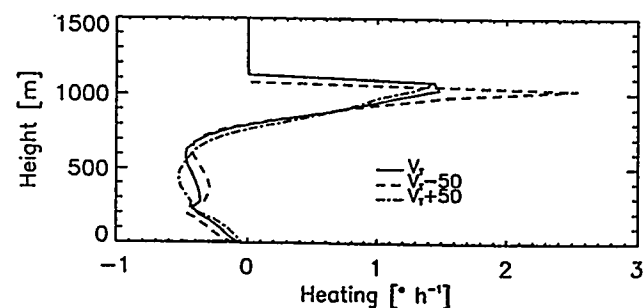


Figure 3. Profiles of heating associated with the drizzle flux for the V_T , $V_T - \epsilon$ and $V_T + \epsilon$ cases. The $V_T - \epsilon$ run shows stabilization relative to the V_T run, while the $V_T + \epsilon$ run shows relative destabilization.

heating associated with the drizzle (calculated from the divergence of the drizzle flux). The profiles show that in the case of the $V_T + \epsilon$ run, most of the cooling associated with evaporation was confined to the region below cloud base (200-500 m), while the $V_T - \epsilon$ run shows cooling spread over the depth of the sub-cloud layer. Relative to the V_T run, the $V_T + \epsilon$ run produces stronger destabilization of the

boundary layer and tends to produce deeper circulations. In contrast the $V_T - \epsilon$ run produces a relative stabilization of the boundary layer; cooling in the region just below base is weaker than in the V_T run, while cooling near the ground is stronger. The resulting circulations tend to be weaker, and confined to two levels. We note that it is the *distribution* of this cooling with height that is critical to generating the stronger deeper circulations.

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Diagnostics from a 1-D Atmospheric Column

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Various diagnostics were computed from an array of radiosondes during an intensive field operation arranged by the Department of Energy's Atmospheric Radiation Measurement (ARM) Program. The diagnostics were derived using objective analysis techniques applied to the radiosonde network data centered on the Cloud and Radiation Testbed (CART) site at Lamont, Oklahoma (Mace 1994).

In the current work, the apparent heat source (Q1) and apparent moisture sink (Q2) were computed (Yanai et al. 1973) and compared to the kinematic vertical velocity for both real data (i.e., objective analysis by applied radiosonde data) and the mesoscale analysis and prediction system (MAPS). Three different case studies, of various weather regimes, were examined from November 1994.

An example in which shallow convection occurred is shown by case 3 (06Z 05 Nov - 06Z 06 Nov 94). Upward vertical motion (Figure 1) associated with a cold front touched off precipitation between 11Z and 17Z (9 mm fell). The apparent heating (Q1), seen in Figure 2, reached a maximum between 10Z and 12Z (the incipient stage of the precipitation) and the apparent drying (Q2), seen in Figure 3, reached a maximum between 13Z and 16Z (during the main precipitation event).

The next phase of this research will be to extend the single column to the lowest 1 km of the atmosphere and to generate a precipitation estimate using the Q1 method (to compare with observations). Through comparison of the budget-derived precipitation estimate and the observed precipitation, the level of accuracy inherent in the data can be evaluated for processes relevant to atmospheric convection. The ultimate goal of this work will be the construction of data sets of cases by which objective analysis errors are below a critical threshold (Mace 1994) and are deemed useful for further diagnostics. These constructed data sets will then be exploited by validating the Integrated Cumulus Ensemble Turbulence (ICET) parameterization scheme currently being developed at Penn State.

Acknowledgments

I would like to thank my advisor, Dr. Jenni Evans for her guidance and suggestions. This research was supported under Department of Energy (DOE) Grant Number DE-FG02-94ER61773.

(a) J. L. Evans, W. Frank, and G. Young, personal communication.

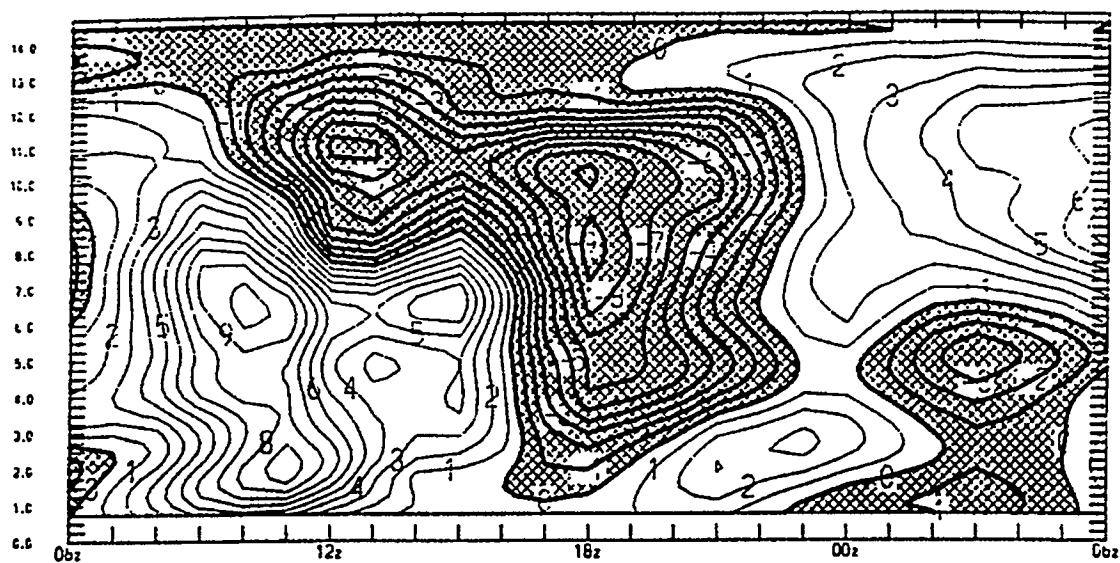


Figure 1. The kinematic vertical velocity at the CART site from 06Z 05 Nov - 06Z 06 Nov 1994.

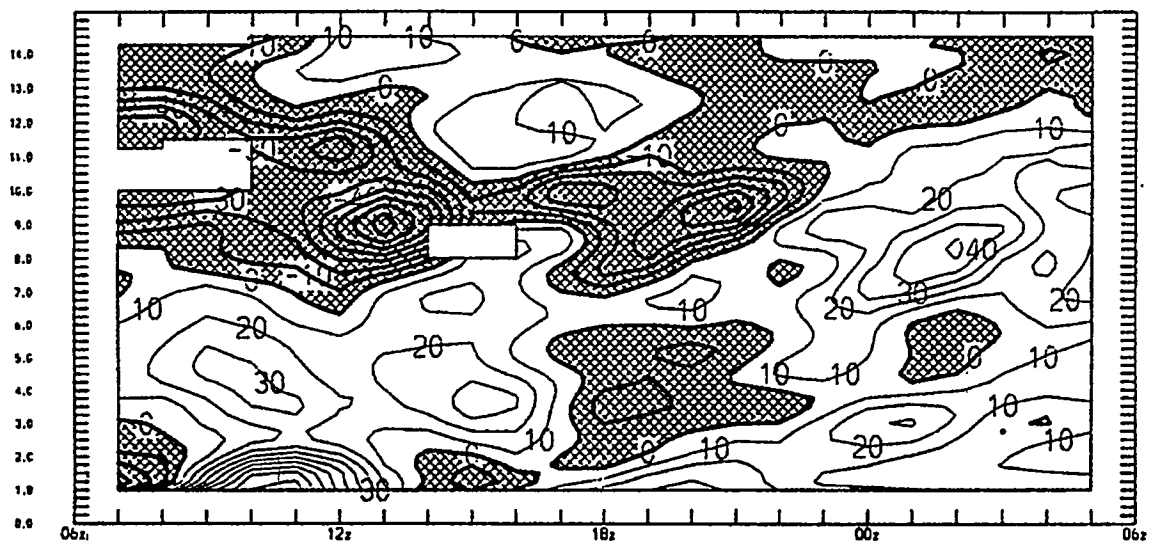


Figure 2. The apparent heat source (Q1) at the CART site from 06Z 05 Nov - 06Z 06 Nov 1994.

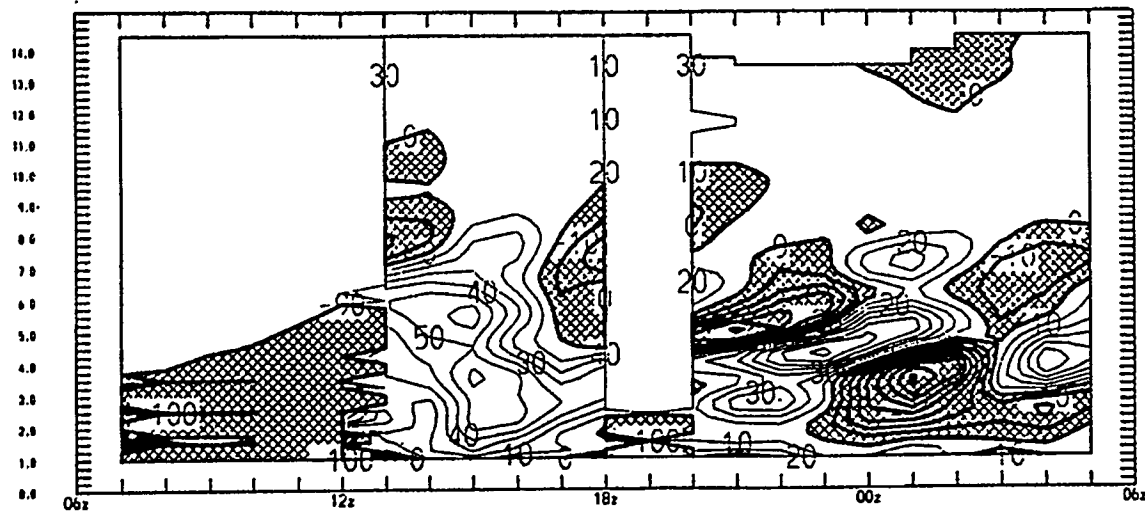


Figure 3. The apparent moisture sink (Q_2) at the CART site from 06Z 05 Nov - 06Z 06 Nov 1994.

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Island Based Radar and Microwave Radiometer Measurements of Stratus Cloud Parameters During the Atlantic Stratocumulus Transition Experiment (ASTEX)

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Introduction

During the Atlantic Stratocumulus Transition Experiment (ASTEX) in June 1992, simultaneous measurements were made with a vertically pointing cloud sensing radar and a microwave radiometer. The radar measurements are used to estimate stratus cloud drizzle and turbulence parameters. In addition, with the microwave radiometer measurements of integrated cloud liquid water and radar measurements of reflectivity, we estimated the profiles of cloud liquid water and effective radius.

We used radar data for computation of vertical profiles of various drizzle parameters such as droplet concentration, modal radius, and spread. A sample of these results is shown in Figure 1. In addition, in non-drizzle clouds, with the radar and radiometer we can estimate the vertical profiles of stratus cloud parameters such as liquid water concentration and effective radius. This is accomplished by assuming a droplet distribution with droplet number

concentration and width constant with height. A sample of these results is shown in Figure 2.

Turbulence measurements are also available in low reflectivity regions of the cloud where droplet mean fall velocity is negligible. We used the Doppler shift to estimate the vertical velocity variance and skewness throughout the cloud. Figure 3 shows an example of the variance and skewness in a stratus cloud. The variance is small compared to the clear air convective boundary layer, and the negative skewness at cloud top indicates that the convection is driven from the top by radiative cooling.

The Doppler shift at several range gates in the cloud can be used to compute the temporal and spatial vertical covariance (Figure 4). Using Taylor's hypotheses, we estimate the vertical integral scale to be about 70 m and the horizontal integral scale about 100 m. In a clear boundary layer driven by surface heating, the horizontal scale would be about 300 m (Lenschow and Stankov 1986). We also estimate the turbulent dissipation rate to be about $0.35 \text{ cm}^2/\text{s}^3$. Our

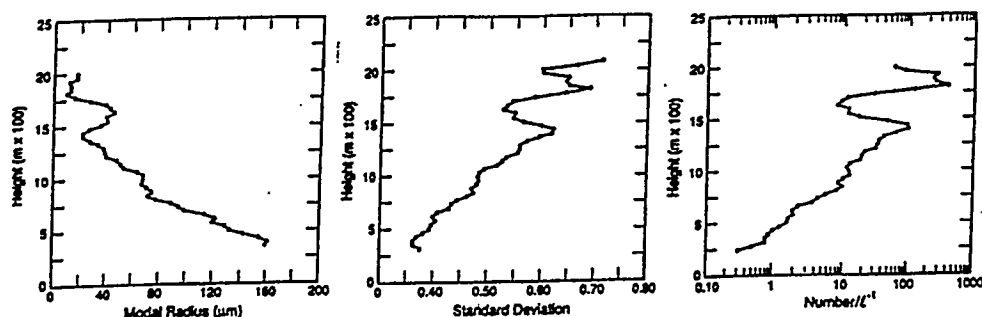


Figure 1. Drizzle parameters. The data for these measurements were taken on June 6, 1992, at 6 am. The first three Doppler moments from the vertically pointing radar were used in a log-normal three parameter cloud droplet model to determine the vertical profiles of modal radius, the standard deviation, and the number of droplets.

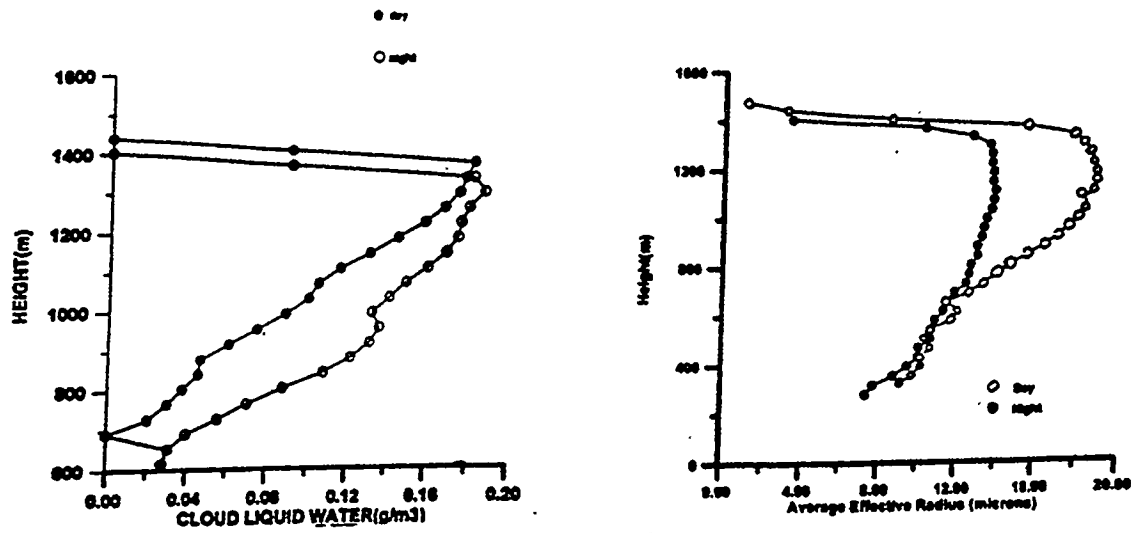


Figure 2. Cloud parameters. These measurements were taken on June 17, 1992, and separated into a daytime and nighttime average. Profiles of cloud liquid water and effective radius were calculated from the radar backscattered power and the radiometer integrated liquid water measurements.

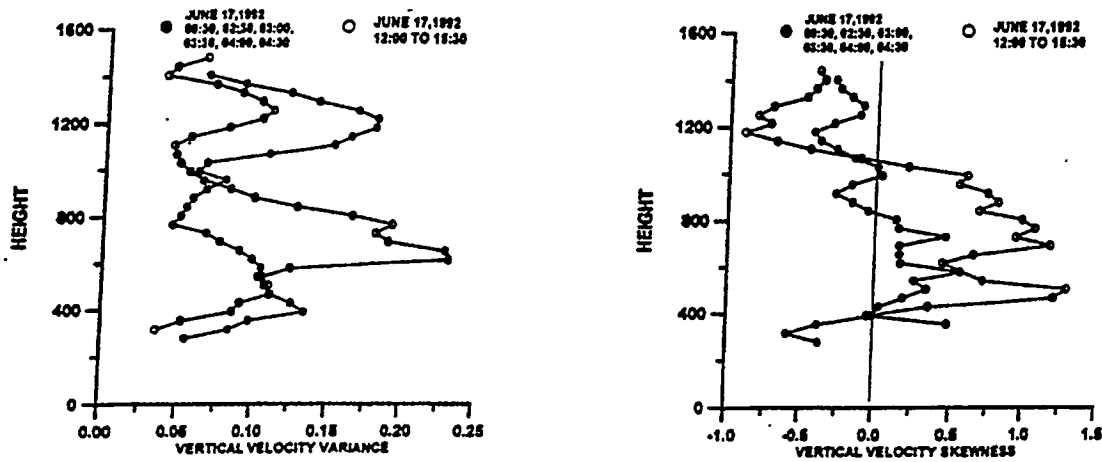


Figure 3. Cloud turbulence parameters. The radar can be used to determine various turbulence quantities in the non-drizzling clouds. Profiles of vertical velocity variance and skewness were computed for June 17, 1992. The start times of the measurements are shown in the figure. The averaging time interval was 23 minutes.

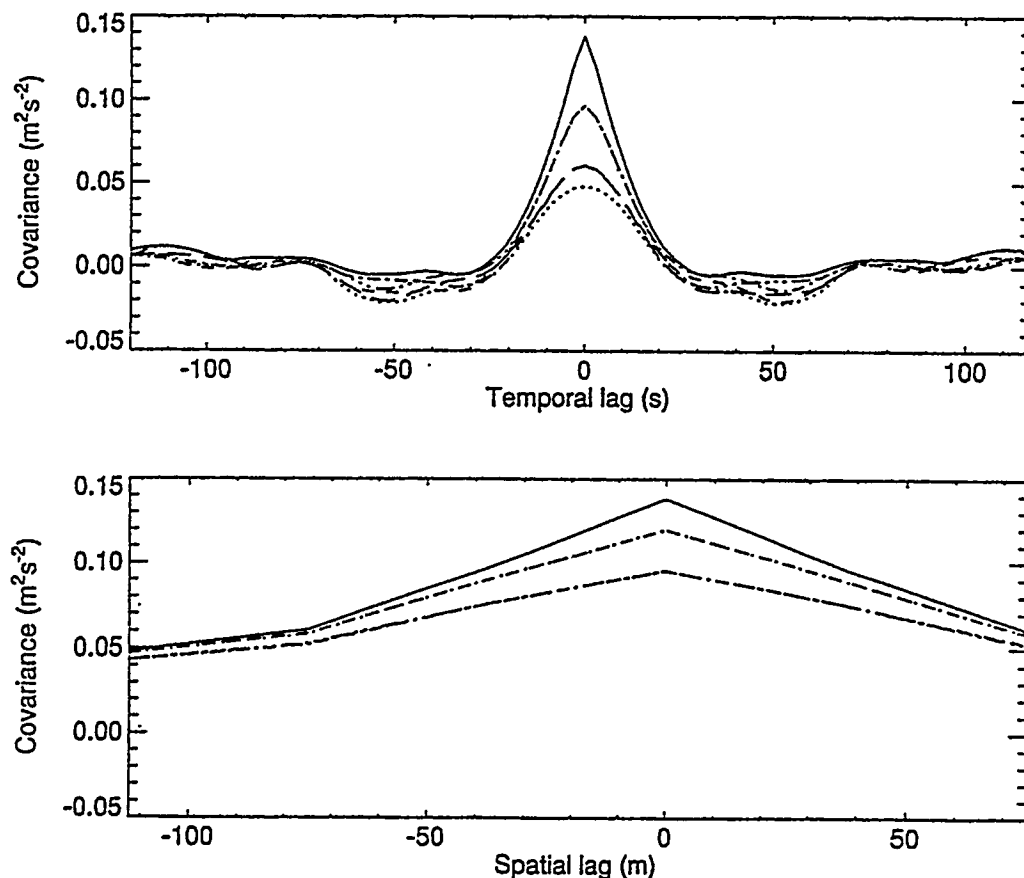


Figure 4. Doppler radar vertical velocity vertical and temporal covariance measurements. The top panel shows the temporal lag and the bottom panel the spatial lag. Sampling is done every 3 seconds and every 37.5 meters.

results show that the stratus turbulence scales are about a factor of 3 smaller than the clear air boundary layer. This result supports our skewness measurements which show that the cloud turbulence is not surface driven, but most likely from radiative cooling at cloud top.

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A Study of Monte Carlo Radiative Transfer Through Fractal Clouds

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Abstract

An understanding of radiation transport (RT) through clouds is fundamental to studies of the earth's radiation budget and climate dynamics. The transmission through horizontally homogeneous clouds has been studied thoroughly using accurate, discrete ordinates radiative transfer models. However, the applicability of these results to general problems of global radiation budget is limited by the plane parallel assumption and the fact that real cloud fields show variability, both vertically and horizontally, on all size scales. To understand how radiation interacts with realistic clouds, we have used a Monte Carlo radiative transfer model to compute the details of the photon-cloud interaction on synthetic cloud fields. Synthetic cloud fields, generated by a cascade model, reproduce the scaling behavior, as well as the cloud variability observed and estimated from cloud satellite data.

Introduction

The goal of this experiment is to simulate radiation through clouds using a scaling field to reproduce the variability of the cloud optical thickness at small-scale length. The scaling fields are generated using cascade processes characterized by a stable distribution specified by two parameters: α and C_1 (see Figure 1). A Monte Carlo model was used to compute the interaction (scattering and absorption) between photons and cloud droplets. The radiance and flux fields were obtained from two realizations of the cloud field, both having the same values α and C_1 .

This experiment is meant to investigate two important questions: 1) To what extent are the scaling and statistical properties of the original scaling fields of the cloud optical depth preserved in the resultant radiation fields? and 2) How well can they be derived from the observed radiance and

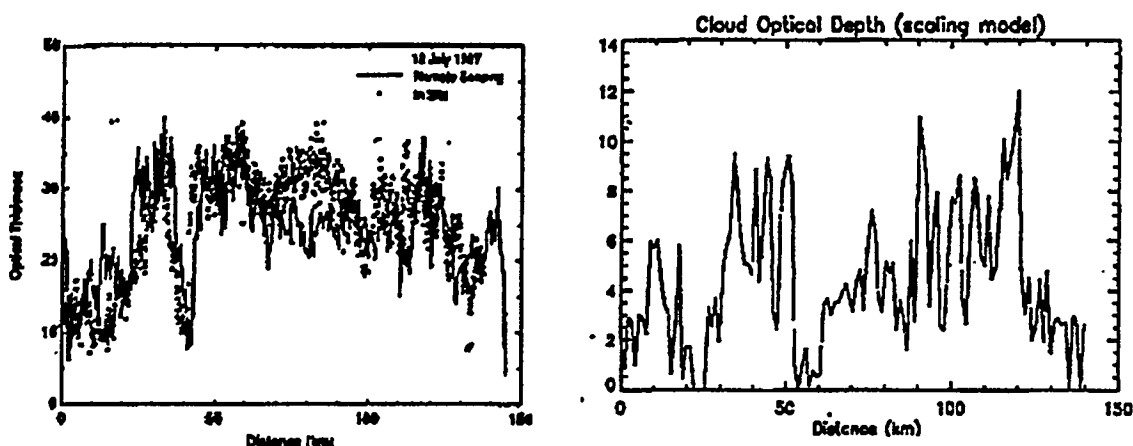


Figure 1. Variability of cloud optical depth derived from remote sensing and in situ measurements (left) (taken from Nakajima and King 1991) and from the scaling model (right) with values $\alpha = 1.3$ and $\beta = 0.1$ (values corresponding to those observed for cloud satellite radiance).

flux fields? These issues are fundamental to the question of how large-scale statistical properties of cloud radiance fields are to be used to describe smaller scale macro- or micro-physical parameters.

Traditional methods to derive cloud properties from satellite imagery have relied on the so called "independent pixel" approximation. In this approach it is assumed cloud properties are constant and horizontally homogenous within each pixel. This assumption permits the use of plane parallel radiative transfer theory to derive the optical depth and other cloud properties within each picture element. This technique has been used within the International Satellite Cloud Climatology Project (ISCCP) to derive global cloud properties at ~km-scale resolution.

DISORT Calculations

As a baseline with which to compare the Monte Carlo runs, we have used the independent pixel approximation together with a discrete ordinate radiative transfer (DISORT) model to compute the radiant intensity received by a satellite sensor. In our simulations the nadir viewing angle and solar zenith angle are both set to zero.

For both the Monte Carlo and DISORT calculations, we use an ocean surface and include effects of marine aerosols with a surface visibility of 23 km. Both codes use the LOWTRAN 7 band models to compute molecular absorption. The surface albedo and aerosol distribution are assumed to be horizontally homogeneous over the entire scene. The clouds are made up of pure water droplets with a size distribution characterized by an effective radius of 8 μm . The radiative properties of the cloud droplets are computed from Mie scattering theory and are parameterized using the Henyey-Greenstein phase function. The DISORT calculation uses 20 radiation streams and 34 vertical computational layers.

In Figure 2 we show DISORT's predictions for the upwelling radiant flux at the top of the atmosphere (TOA) for cloud optical depths in the range 0 to 64. The Channel 1 band is characterized by conservative photon scattering, whereas the Channel 2 band displays significant water vapor absorption.

The Monte Carlo Model

The Monte Carlo method is essentially a direct simulation of the physical processes involved in radiative transfer where the path of a photon is described by probability functions. These functions describe the distance a photon travels before an interaction; the result of the interaction (absorption or scattering); and, if scattered, the scattering direction. The

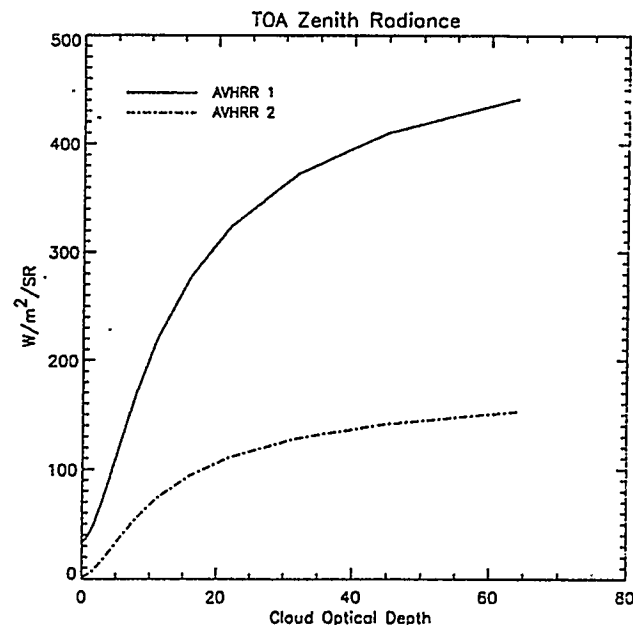


Figure 2. The top of atmosphere (TOA) radiance from DISORT (plane parallel) for two visible channels, 630 nm and 931 nm, representing the pass bands of the AVHRR Channel 1 and 2 detectors. Compared with Channel 1, the Channel 2 pass band is characterized by much greater water vapor absorption and nearly vanishing reflectance from the ocean surface.

Monte Carlo model used in these simulations is designed around a 3-dimensional spatially dynamic structure. This structure provides for large, low variability cells to reduce computations and nested higher resolution cells to allow the mixing of various spatial scales. Each cell can contain any combination and concentration of gases, aerosols, and cloud droplets.

Results

Manifestation of scaling can be observed using standard spectral analysis. For instance, power spectrum of a scaling fields is given by

$$E(k) \sim k^{-\beta} \quad (1)$$

where $k=|k|$ is the wave number magnitude and the scaling exponent β associated with $E(k)$. For 2-D fields, the power spectrum is obtained by first integrating over the azimuth angle the squared modulus of the Fourier transform of the field and second by taking the statistical ensemble average (usually approximated by a sum over independent statistical samples when available).

Comparisons between the scaling behavior of $E(k)$ for cloud optical thickness, and the radiance from both DISORT and Monte Carlo are presented in Figure 3. The power spectra of the radiance fields decrease with power laws similar to the input optical depth field. This indicates that the scaling is fairly well preserved. It also suggests that, within a multiplicative constant, the correlations are preserved by DISORT and Monte Carlo, and each have a similar distribution of the second order moments (associated with cloud optical thickness and both radiance channels) as a function of the wave number. The very similar spectral slope of DISORT's radiance and the original optical depth field are due to a low average value of the cloud optical thickness. For small values of optical depth, the radiance from DISORT is an almost linear transformation of the original cloud optical thickness.

Although the spectral slopes of the Monte Carlo radiances are close to that of the cloud optical thickness (a difference smaller than 10%), the curves are more variable (perhaps due to photon count statistics). Also, whereas the DISORT results indicate no difference in the $E(k)$ for the two advanced very high resolution radiometer (AVHRR) channels and the optical depth, the Monte Carlo results show slightly less spectral power at high spatial frequencies for Channel 2 and more power at high frequencies for Channel 1. This result is surprising. We had anticipated

that the horizontal photon transport allowed in Monte Carlo would cause lower power at high spatial frequency for both the Channel 1 and Channel 2 radiances, with perhaps less high frequency power in Channel 1 because the conservative scattering should augment the horizontal diffusion of photons. Instead, we find reduced high frequencies in the Channel 2 radiance. We have repeated this experiment on 10 different cloud optical depth samples and have consistently found steeper spectral slopes for Channel 2 vs. Channel 1. A possible explanation is that water vapor absorption is accentuated in regions of increased optical depth variability. Monte Carlo's 3-D photon diffusion allows deeper penetration of photons into regions where they can be absorbed by water vapor. Thus Channel 2 power spectra tend to be depleted at high frequencies because the regions with increased variability reflect back fewer photons.

These results, although preliminary, are suggestive of interesting 3-D effects which may have an impact on cloud retrieval algorithms that depend on the wavelength dependence of cloud reflectance. For example, the retrieval of cloud optical depth and cloud drop effective radius depends on simultaneous observations at visible and near infrared (IR) wavelengths. The enhanced absorption in the near IR brought about by increased cloud variability may distort these retrievals. In future experiments we will clarify this issue by repeating the same study over a larger number of samples and for other spectral bands. Analysis of the scaling properties in the physical (real) space will also be investigated.

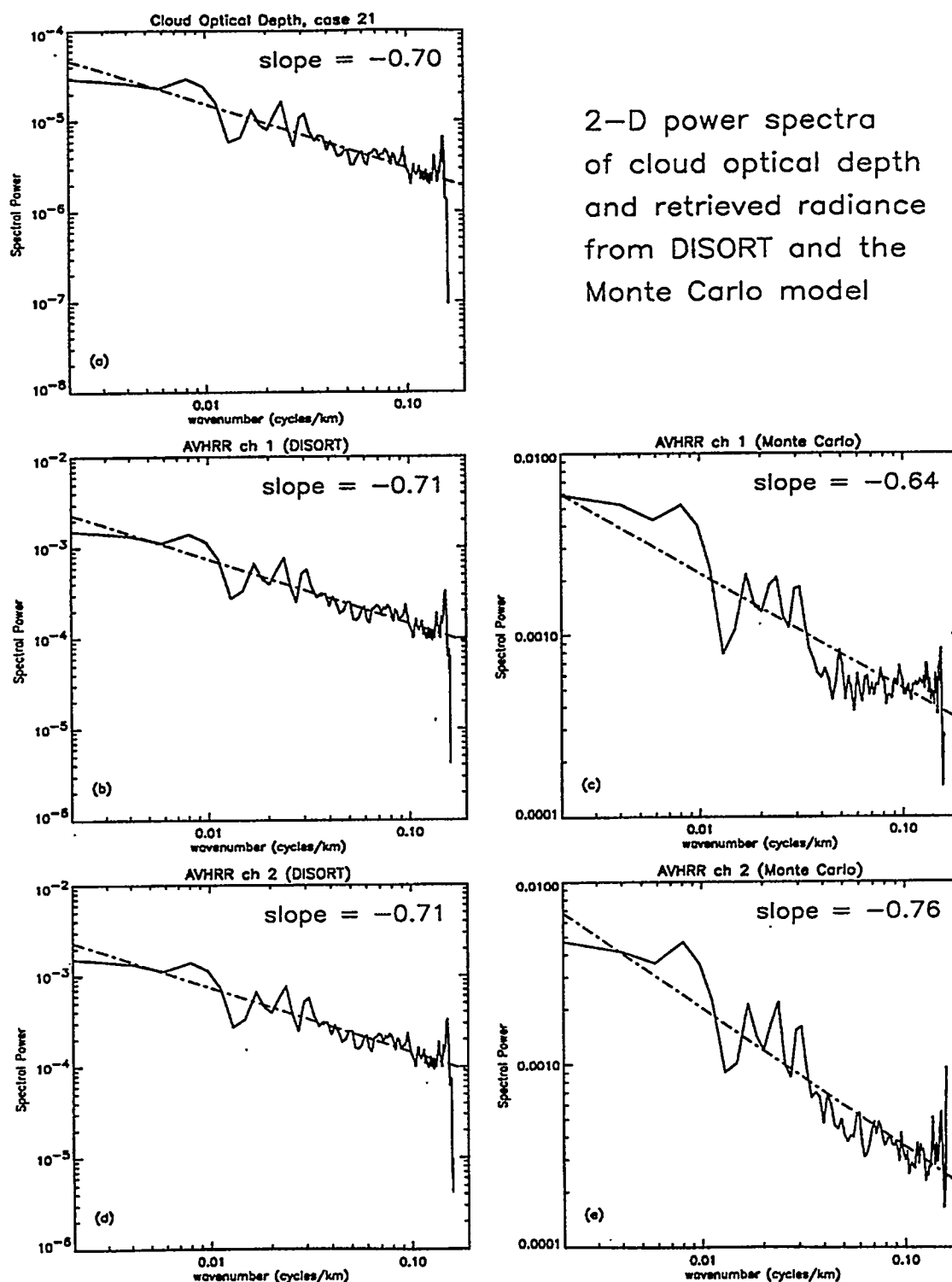


Figure 3. 2-D power spectra of a) the cloud optical depth realization, b) retrieved AVHRR Channel 1 radiance from DISORT, c) retrieved AVHRR Channel 1 radiance from Monte Carlo, d) retrieved AVHRR Channel 2 radiance from DISORT, and e) retrieved AVHRR Channel 2 radiance from Monte Carlo.

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Simulation of the Low-Level Jet by General Circulation Models

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To what degree is the low-level jet (LLJ) climatology and its impact on clouds and precipitation being captured by current general circulation models (GCMs)? My hypothesis is that current GCMs do not simulate the relationship between clouds and the LLJ, so that a parameterization will be required. Here I evaluate the need for a parameterization.

To assess the simulation of the LLJ, I use the LLJ criteria of Bonner (1968). I applied Bonner's criteria to the winds simulated by two GCMs, namely the European Center for Medium-Range Weather Forecasts (ECMWF) model and the National Center for Atmospheric Research Community Climate Model (CCM2). Both models were run at T42 resolution, with the ECMWF model driven by sea surface temperatures observed during the Atmospheric Model Intercomparison Project period 1980-1989 and the CCM2 driven by climatological mean sea surface temperatures. The ECMWF simulation was performed by the Program for Climate Model Diagnoses and Intercomparison (PCMDI) at Lawrence Livermore National Laboratory. My analysis focuses on the ECMWF simulation for 1980-1982 and the CCM2 simulation for 3 years. Figure 1 shows the spatial distribution of simulated LLJ frequency for Bonner criterion 1, using 6-hour samples. The LLJ frequency distribution simulated by the ECMWF model is remarkably similar to that estimated from observations (Bonner 1968), with the maximum frequency of about 30% occurring over Oklahoma, and frequencies of less than 5% over most of the eastern and western United States. The LLJ frequency distribution simulated by the CCM2 also exhibits a (somewhat weaker) maxima over Oklahoma, with a minima on the Gulf coast as observed (Bonner 1968). Helfand and Schubert (1994) found a very similar distribution of the LLJ frequency simulated by the Geostationary Operational Environmental Satellite-1 model. The nocturnal maximum of the LLJ frequency and intensity is also simulated quite well by all three GCMs. Interestingly, all three GCMs also simulate a frequent LLJ off the coast of California.

It appears that at least some GCMs can simulate the LLJ, so the LLJ itself need not be parameterized. However, the influence of the LLJ on clouds and precipitation may still need to be parameterized. To evaluate the need for parameterizing this influence, I have examined the diurnal

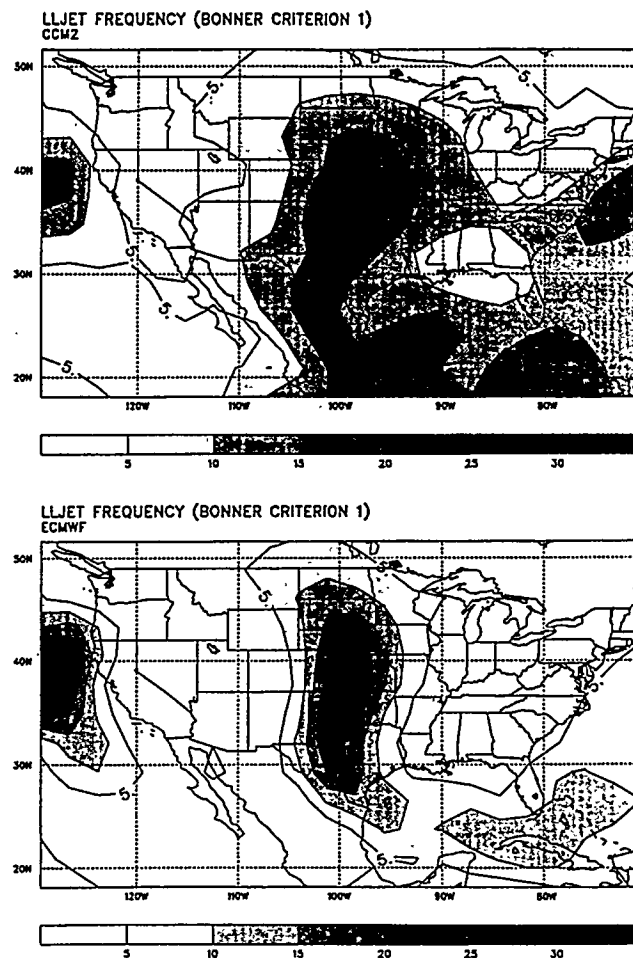


Figure 1. Frequency of LLJ (Bonner criterion 1) simulated by CCM2 (above) and ECMWF (below) models for a period of three years.

cycle of the convective precipitation rate simulated by the ECMWF and CCM models. Convective precipitation is strongly related to the occurrence of thunderstorms (Slingo 1987). Hering and Borden (1962) and Pitchford and London (1962) have shown that associated with the LLJ is a nocturnal maximum in convective activity over the central

United States (Wallace 1975). Thus, one means of evaluating the simulation of the influence of the LLJ on clouds and precipitation is to examine the diurnal cycle of the simulated convective precipitation, shown in Figure 2 for

the ECMWF model and for CCM2. Neither GCM simulates a significant nocturnal maximum in convective precipitation in the central United States. Thus, it appears that the influence of the LLJ on convective clouds and precipitation will have to be parameterized.

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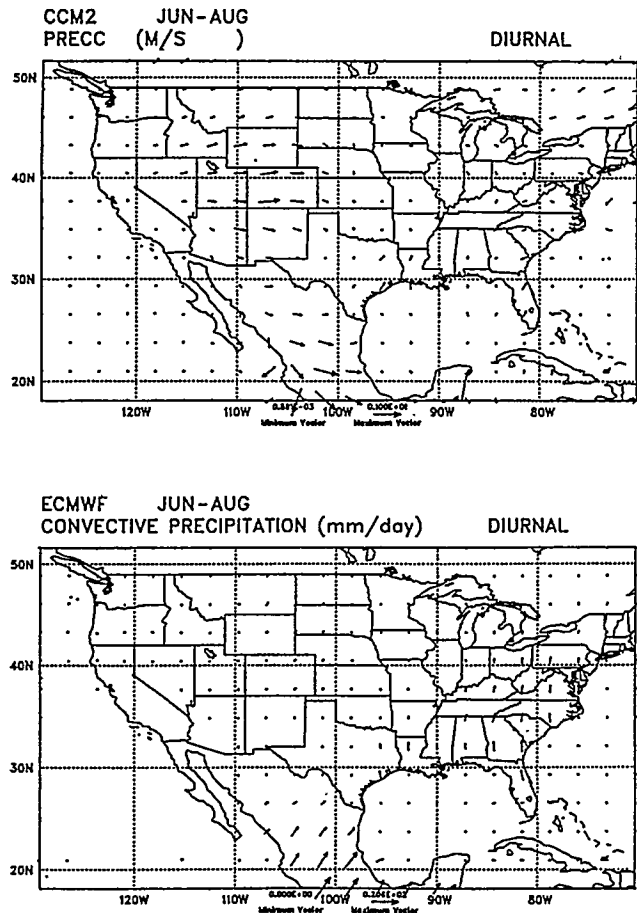


Figure 2. Amplitude and phase of diurnal cycle of convective precipitation simulated by CCM2 (above) and ECMWF (below) models for three years of June-July-August. Phase is local midnight for vectors pointing southward, 6 a.m. for vectors pointing westward, etc.

Prediction of Cloud Droplet Number in a General Circulation Model

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We have applied the Colorado State University (CSU) Regional Atmospheric Modeling System (RAMS) bulk cloud microphysics parameterization to the treatment of stratiform clouds in the National Center for Atmospheric Research Community Climate Model (CCM2). The CSU RAMS cloud microphysics parameterization predicts mass concentrations of cloud water, cloud ice, rain and snow, and number concentration of ice. To predict droplet number and its dependence on aerosols, we have introduced the droplet number conservation equation

$$\frac{\partial N_c}{\partial t} = -\nabla \cdot (N_c V) + N_{vc} - E_{cv} - A_{cr} - C_{cr} - C_{ci} - C_{cs} - F_{ci}$$

where E_{cv} represents droplet evaporation; A_{cr} is the droplet sink due to droplet coalescence; C_{cr} , C_{ci} , and C_{cs} , are sinks due to collection by rain, ice and snow; and F_{ci} is the sink due to droplet freezing to form ice crystals. Assuming all cloud droplets are formed as air flows into clouds, the droplet nucleation rate can be expressed as $N_{vc} = -\nabla \cdot N(V)$ where N equals the number nucleated for flow from clear air into cloud, and zero for flow out. For stratiform clouds, we neglect droplet formation on the sides of the cloud, but account for both resolved and turbulent transport of air into both the top and base of the cloud:

$$N_{vc} = \frac{1}{\Delta_z} \left[\int_0^{\infty} w_b N_n(w_b) p(w_b) dw_b - \int_{-\infty}^0 w_t N_n(w_t) p(w_t) dw_t \right]$$

where Δ_z is the model layer thickness, w is the vertical velocity, $p(w)$ is the subgrid probability distribution of w at cloud base (subscript b) and cloud top (subscript t). The number nucleated N_n is parameterized in terms of vertical velocity, radiative cooling, aerosol number concentration, size distribution, and composition (Ghan et al. 1993, 1995). Supersaturation forcing by cloud-top radiative cooling is accounted for by expressing the number nucleated in terms of

$$w^* \equiv w - \frac{c_p}{1 - c_p} \frac{Q_r}{T(\epsilon L)} \frac{Q_r}{g}$$

which follows from consideration of the supersaturation balance (Ghan et al. 1993).

The probability distribution of vertical velocity is parameterized in terms of the grid cell mean and subgrid variance of vertical velocity, assuming a normal distribution. The subgrid variance of vertical velocity is related to the turbulent kinetic energy predicted in the Pacific Northwest National Laboratory version of CCM2. Turbulent kinetic energy is predicted by the application of the Yamada and Mellor (1979) second-order turbulence closure (level 2.5) scheme. The integrations over the probability distributions are performed semi-numerically, taking advantage of asymptotic behavior to perform part of the integration analytically.

To test the prediction of cloud droplet number, we have performed two simulations in which the aerosol number concentration, size distribution, and composition are prescribed. In both simulations the aerosol is composed of ammonium sulfate with a log-normal size distribution with a number mode radius of $0.05 \mu\text{m}$ and a geometric standard deviation of 2. The aerosol number concentration is prescribed at 100 cm^{-3} in one simulation and 1000 cm^{-3} in the other, representative of marine and continental aerosol concentrations, respectively. The simulations were performed at T42 resolution for a period of thirty days, beginning September 1.

The two simulations yield substantially different droplet number concentrations. Figure 1 shows the frequency distribution of the simulated droplet number concentration (averaged over all times cloud water is present) for the two simulations. Note that a small fraction (5%) of the clouds have droplet number concentrations exceeding the aerosol number concentration; this physically unrealistic result is due to differences between the treatment of vertical diffusion of droplets from the cloud and the treatment of air-flow into the cloud. The median droplet number concentrations in the two simulations are 24 and 127 cm^{-3} , for aerosol number concentrations of 100 and 1000 cm^{-3} , respectively. Thus, a ten-fold increase in aerosol number resulted in a five-fold increase in droplet number.

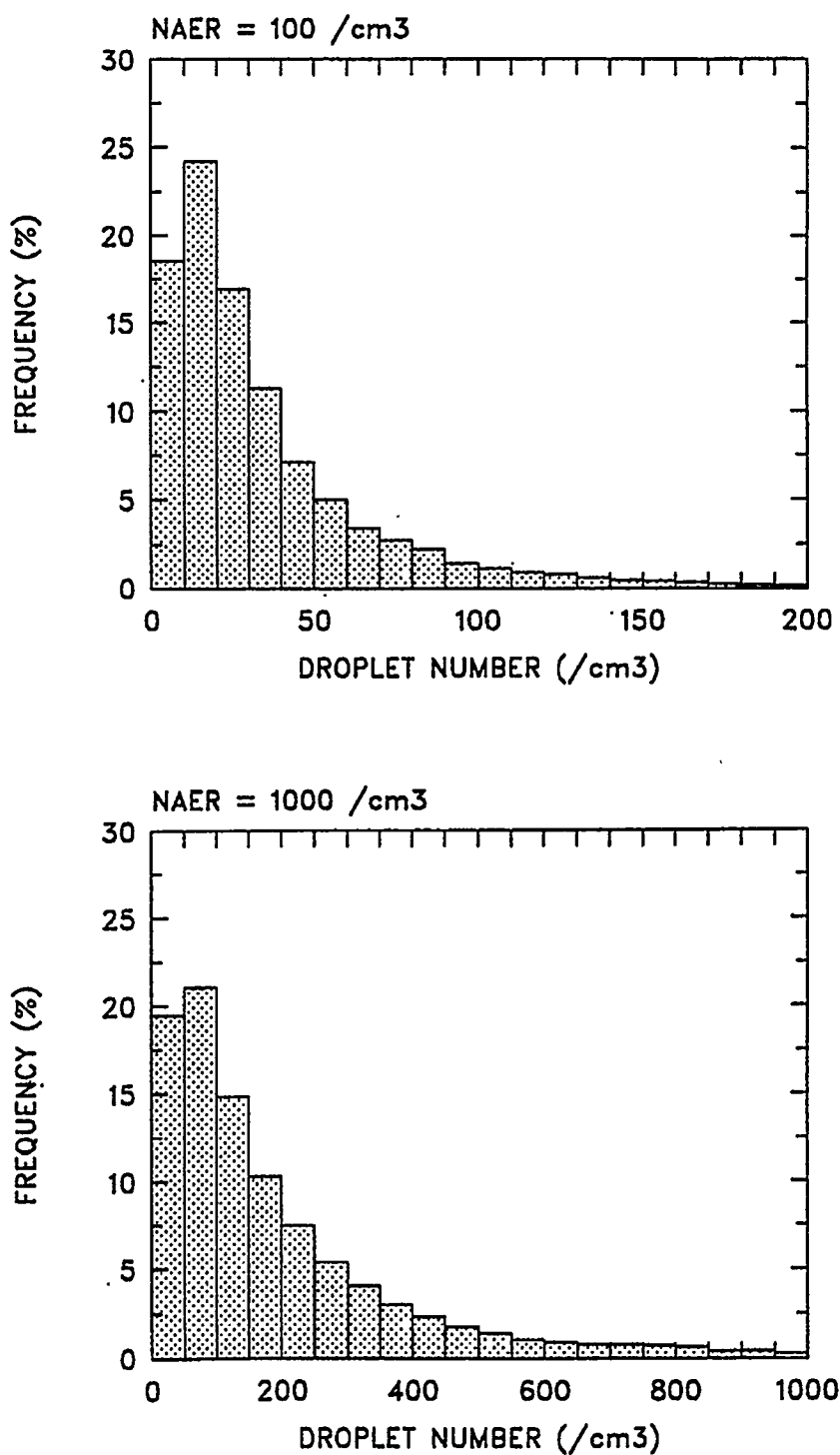


Figure 1. Frequency distribution of 30-day mean droplet number concentration simulated by Pacific Northwest National Laboratory version of CCM2 with aerosol number concentration prescribed at 100 cm^{-3} (above) and 1000 cm^{-3} (below).

The increased droplet number concentration resulted in an increase in the planetary albedo through two mechanisms. In addition to decreasing the effective radius of cloud droplets, increasing the droplet number concentration also permits higher cloud liquid water concentrations before autoconversion is initiated. The mean liquid water path of clouds is doubled by the increase in droplet number concentration. The mean cloud optical depth is more than doubled, partly due to the increase in liquid water path and partly due to the reduction in effective radius. The increase in the mean planetary albedo is 1.3%, corresponding to a 4.4 W m^{-2} increase in reflected solar radiation. We expect our estimate of the indirect radiative impact of aerosols to increase as we correct biases in the simulated low clouds and in the simulated partitioning between the liquid and ice phases of cloud water.

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Evaluation of a Stratiform Cloud Parameterization for General Circulation Models

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One of the requirements of a cloud parameterization is that it represent the full lifecycle of clouds: their formation, persistence, and decay. For clouds forming under strongly advective conditions, the domain traversed by a cloud during its lifecycle can be thousands of kilometers, much larger than that of a single grid cell of a General Circulation Model (GCM). Evaluation of a cloud parameterization under such conditions can only be achieved by a model spanning a domain much larger than that of a Single Column Model (SCM); otherwise the lateral boundary conditions for the cloud variables will control the simulation of the cloud. Thus, even if the Atmospheric Radiation Measurement (ARM) Program could measure the cloud variables on the lateral sides of the Cloud and Radiation Testbed (CART) site, the evaluation of cloud parameterizations would, under strongly advective conditions, be compromised by the lateral boundary conditions for the cloud variables.

To evaluate the relative importance of horizontal advection of cloud versus cloud formation within the grid cell of an SCM, we have performed a series of simulations with our SCM driven by a fixed vertical velocity (1 cm s^{-1} , characteristic of 250-km resolution models; Sasamori 1975) and various rates of horizontal advection. Horizontal advection is treated assuming no clouds are transported into the SCM grid cell (for lack of observations of clouds), but clouds formed within the grid cell can be transported out. The simple upstream scheme is used to treat advection, which for constant flow across a grid cell of width Δx reduces to

$$(\mathbf{V} \cdot \nabla \mathbf{r})_i = v_{i+1/2} \frac{r_{i+1} - r_i}{\Delta x} \quad (1)$$

where $v_{i+1/2}$ is the velocity on the upwind side of the grid cell, r_i is the cloud variable simulated by the SCM, and r_{i+1} is the cloud variable upwind from the grid cell, assumed to be zero (or r_v in the case of advection of total water r_w) for lack of cloud observations. The grid size is fixed at 200 km, but the horizontal wind speed varies from zero to 15 m s^{-1} . The case with zero wind speed corresponds to the assumption that clouds outside the SCM domain are the same as clouds within. Figure 1 shows the column cloud water as a function of time for different values of horizontal wind speed. The difference between the treatment assuming homogeneous conditions (zero wind) and the treatment assuming no clouds are transported into the domain increases monotonically with increasing wind speed (roughly 20%, 35%, and 50% for wind speeds of 5, 20, and 15 m s^{-1} , respectively). This indicates that the results can be quite sensitive to the assumed cloud distribution along the lateral boundaries. We therefore conclude that an SCM is not a viable testbed for evaluating cloud parameterizations, except under conditions of weak wind speeds, which in our simulation case are less than 5 m s^{-1} .

As an alternative to the use of SCMs as a testbed for evaluating cloud parameterizations, we are using a multi-dimensional Regional Circulation Model (RCM). The model generates cloud variables both within and outside the CART measurement domain, but only the clouds simulated within the CART domain are compared with observations. It is important to note that the same cloud parameterization used within the CART domain is also used outside the domain so that, in contrast to simulations by an SCM driven by boundary conditions from a Four-Dimensional Data

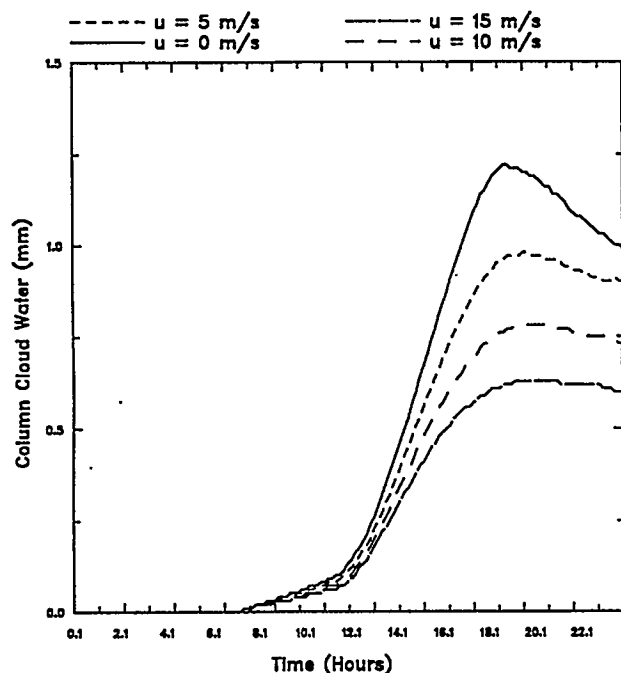


Figure 1. Simulations of column cloud water by the single column model with different horizontal wind speeds. The model was initialized with relative humidity that decreases from 90% to 60% from the surface to 6 km, and is driven by a fixed vertical velocity of 1 cm s^{-1} .

Analysis (FDDA) system, there will be no ambiguity about whether cloud simulation errors are due to the cloud parameterization inside or outside the CART domain. The simulation period is at least a few days to allow the clouds to form with little dependence on the poorly known initial specific humidity. Four-dimensional data assimilation is used to constrain the simulation toward the observed winds throughout the simulation. The domain of the RCM is chosen to be large enough that the lateral boundary conditions for the cloud variables do not significantly

influence the clouds simulated within the CART domain. Experience with the First International Satellite Cloud Climatology Project Regional Experiment (FIRE) Cirrus experiments (Westphal and Toon 1990) suggests a domain of several thousand kilometers is required. The horizontal resolution of the model is consistent with that of GCMs (i.e., 100-200 km, so that cloud parameterizations can be evaluated at the appropriate resolution).

As a proof of principle, we offer the following comparison between simulated and observed outgoing shortwave radiation for a model run with simple nudging. The MM5 prototype MMZIGGY, running with our cloud parameterization (Ghan and Easter 1992), was used to simulate the development of a frontal system on the lee side of the Rocky Mountains from November 6 through November 8, 1987. Calculations were performed on a 40×45 gridded domain centered at 34.5°N and 99.0°W , with a grid spacing of 60 km. Two model integrations were completed. For the first, boundary conditions obtained from the European Centre for Medium-Range Weather Forecasts-gridded analysis were used, and no data assimilation was performed. For the second, the same boundary conditions were employed, and in the interior of the domain, the model temperature and horizontal winds were nudged toward values obtained from the gridded analysis, using nudging coefficients of $2.5 \times 10^{-4} \text{ s}^{-1}$. In Figure 2a Geostationary Operational Environmental Satellite (GOES) shortwave image is compared to the output from both model runs 44 hours into the model simulations, at 20 GMT, November 7. Figure 2a shows a GOES visible image from 19:31 GMT; Figure 2b shows the model-calculated outgoing shortwave radiation from the nudged run; and Figure 2c shows the model-calculated outgoing shortwave radiation from the run with no nudging. The developing cold front can be easily discerned in the nudged simulation as it crosses the CART site in Oklahoma. It is clear that the simulation of the large-scale features of the system has been improved with the inclusion of wind and temperature nudging.

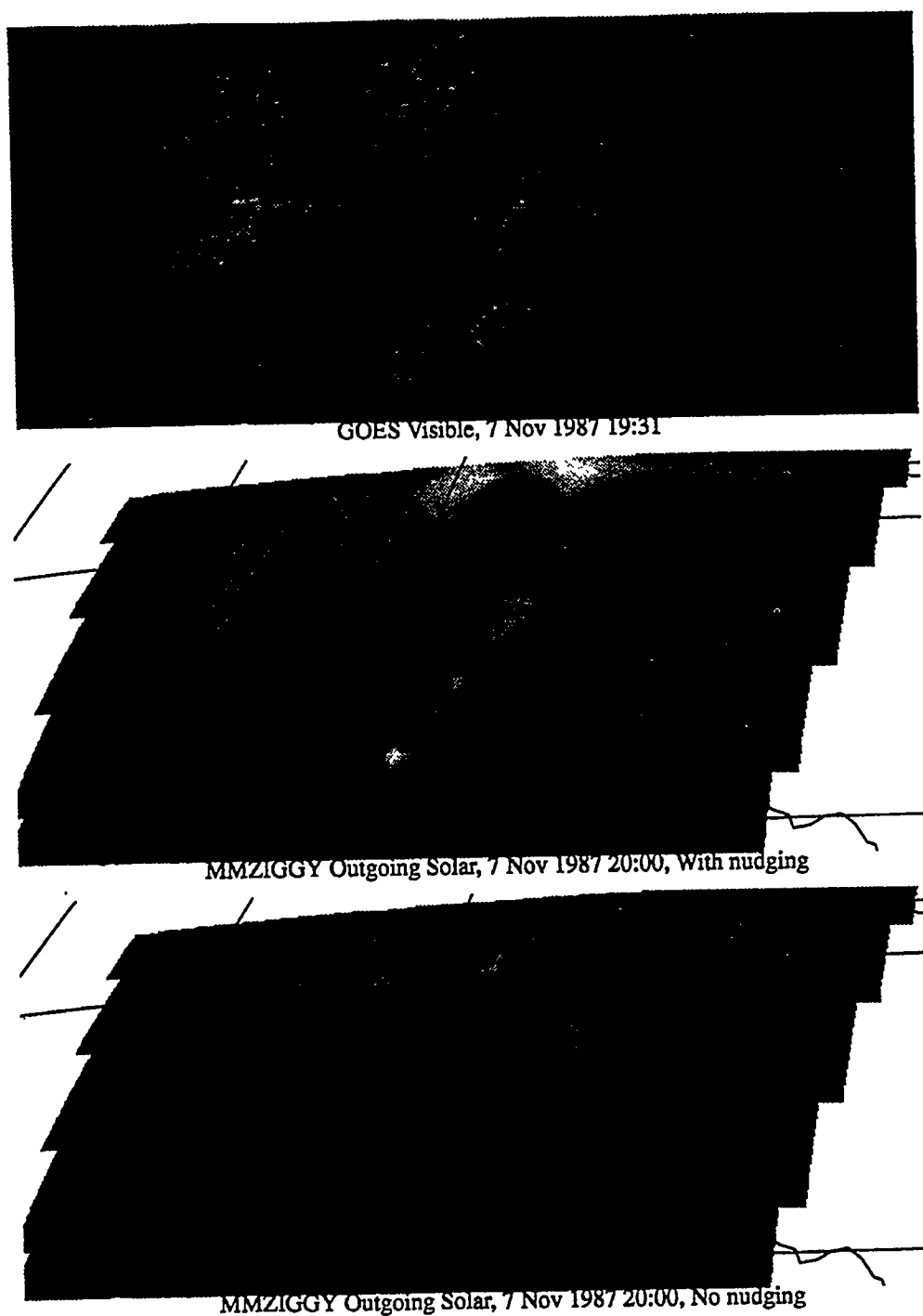


Figure 2. Simulated and observed outgoing solar radiation for a domain centered over the ARM Southern Great Plains CART site: (a) GOES visible image at 19:31 GMT, November 7, 1987; (b) Outgoing solar radiation simulated at 20 GMT, November 7, 1987, after initialization at 00 GMT, November 6, 1987, and 44 hours of simulation with data assimilation; and (c) As in (b), but without data assimilation.

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Implementation of Raman Lidar for Profiling of Atmospheric Water Vapor and Aerosols at the Southern Great Plains (SGP) Cloud and Radiation Testbed (CART) Site

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Introduction

There are clearly identified scientific requirements for continuous profiling of atmospheric water vapor at the Southern Great Plains (SGP) Cloud and Radiation Testbed (CART) site. Research conducted at several laboratories, including our work in a previous Atmospheric Radiation Measurement (ARM) Instrument Development Project, has demonstrated the suitability of Raman lidar for providing measurements that are an excellent match to those requirements. We are in the final stages of building a ruggedized Raman lidar system that will reside permanently at the CART site and that is computer-automated to reduce the requirements for operator interaction. In addition to the design goal of profiling water vapor through most of the troposphere during nighttime and through the boundary layer during daytime, the lidar will provide quantitative characterizations of aerosols and clouds, including depolarization measurements for particle phase studies.

Raman lidar systems detect selected species by monitoring the wavelength-shifted molecular return produced by Raman scattering from the chosen molecule or molecules, as illustrated in Figure 1. For water-vapor measurements, the nitrogen Raman signal is observed simultaneously with the water-vapor Raman signal; proper ratioing of the signals yields the water-vapor mixing ratio. Similarly, by simultaneously recording the backscatter signal at the laser wavelength (which contains contributions from both Rayleigh and aerosol scattering), the ratio of the backscatter signal to the nitrogen Raman signal yields a quantitative measurement of the aerosol scattering ratio, and a variety of aerosol and cloud parameters can be derived from this measurement. In aerosol-free regions of the atmosphere, temperature profiles can be derived from the density measurements obtained from the nitrogen Raman signal. Finally, by adding polarizing optics and an additional direct-backscatter channel,

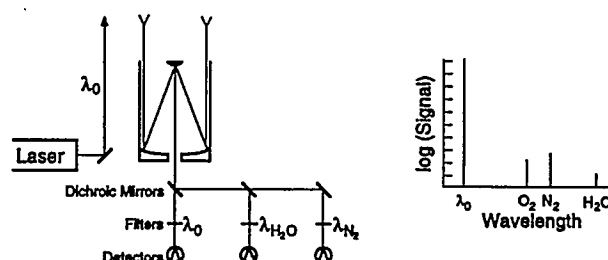


Figure 1. (a) Schematic diagram of idealized Raman lidar system. (b) Signals produced by the primary species in the atmosphere.

depolarization measurements can provide information about the phase (water droplet or ice particle) of clouds detected by the lidar system.

The Raman lidar system we are building to reside at the SGP CART site is housed in a seatainer, a metal shipping container that measures approximately 8 ft x 8 ft x 20 ft. The system is fully self-contained, requiring only an external supply of three-phase 208-V power. Optical access is provided by a weather-tight window in the roof of the seatainer; the window is covered by a motorized hatch during bad weather. A great deal of attention has been paid to the climate-control system to ensure reliable operation in the non-laboratory environment of the CART site. The overall layout of the system is shown in Figure 2.

Two laser systems are good candidates for a Raman lidar system of this type, Nd:YAG lasers and excimer lasers. Both have had considerable success in several systems. Although an excimer-based system has recently been developed at Sandia National Laboratory, we elected to use a frequency-tripled Nd:YAG laser in the CART system. Nd:YAG lasers have four distinct advantages over excimer lasers: better beam quality (needed for narrow-field-of-view

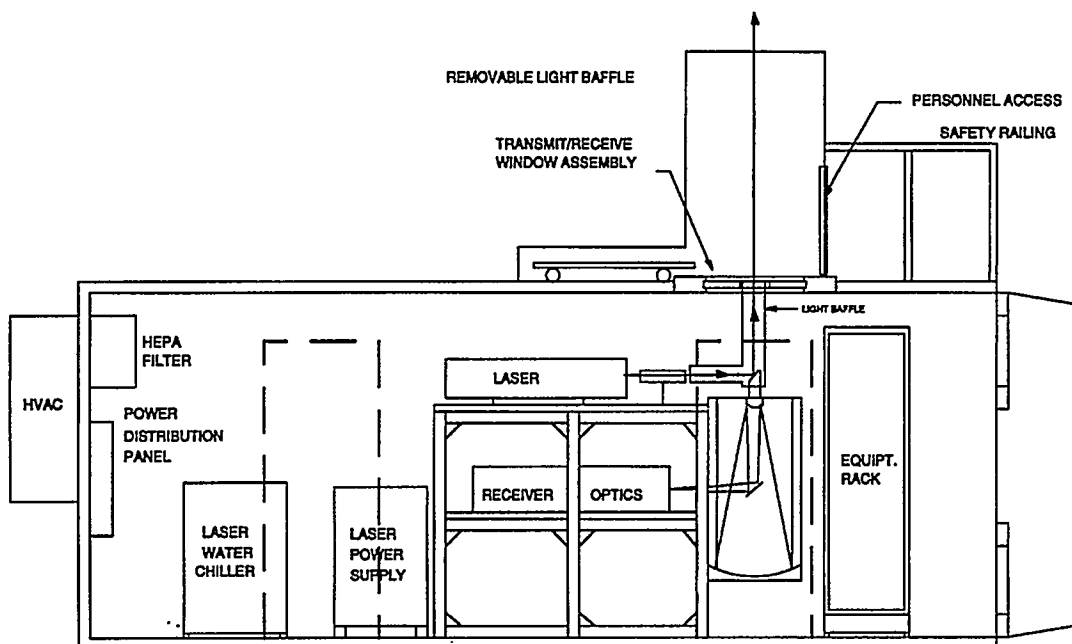


Figure 2. Layout of the seatainer that houses the CART Raman lidar.

operation), more readily polarized (needed for depolarization measurements), no toxic gases required, and much lower cost for consumables and maintenance. Overall laser reliability is difficult to judge, although Nd:YAG lasers have the significant advantage of not requiring corrosive gases and having no moving parts. Excimer lasers do produce higher ultraviolet output power, but system performance modeling indicates that the higher power does not offset the other disadvantages.

The need to have a system that operates in both daytime and nighttime and that can make accurate measurements both near the ground and at long ranges, places somewhat conflicting constraints on the system design. We have implemented a dual-field-of-view design to provide the best compromise among these requirements (see Figure 3). The first beamsplitter in the path marked "from telescope" in Figure 1 directs 5% of the light collected by the telescope through a relatively large aperture, which defines the "wide" field-of-view necessary to establish a transmitter-receiver overlap function which can record signals at short range. Dichroic beamsplitters direct the three wavelengths of interest through narrowband interference filters to photomultiplier tubes. This wide field-of-view produces a large solar background during the daytime, but this large background is offset by the large close-range signal. The remainder of the light collected by the telescope passes through a smaller aperture, which defines the "narrow" field of view, followed again by

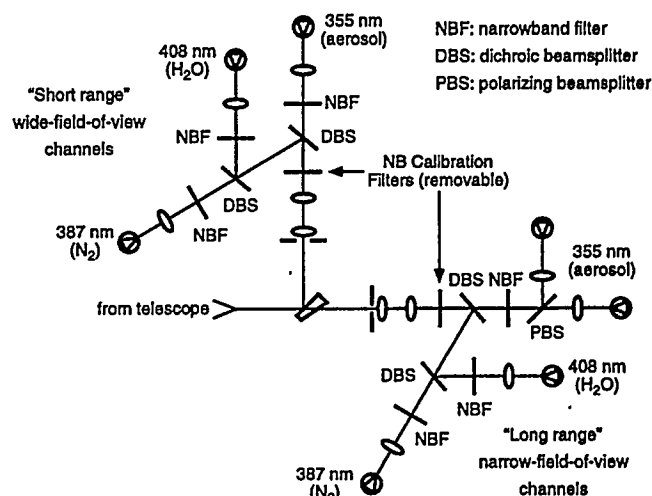


Figure 3. Optical layout for dual-field-of-view Raman lidar receiver.

dichroic beamsplitters and narrowband interference filters, with an added polarizer and a second 355-nm channel for making depolarization measurements. The narrow field-of-view produces far less solar background, and provides optimum long-range signal. Our modeling indicates that this design will provide excellent daytime performance without sacrificing any nighttime capability.

The lidar electronics have also been designed to optimize the performance of the dual-field-of-view design. Analog-to-digital conversion is being used for the short-range, wide-field-of-view channels, where large signals make photon counting very difficult. Photon counting is used for the narrow-field-of-view channels to provide optimum long-range performance. Detailed system specifications are listed in Table 1. To provide higher sensitivity for longer-range measurements that do not require such high resolution, the profiles can be averaged in time and/or space during post-processing. The system thus provides complete flexibility in trade-offs among sensitivity, spatial resolution, and temporal resolution. These trade-offs can be explored at any time and with any range dependence desired.

The entire system is operated by a LabView-based program implemented on a conventional Windows-based PC. The program optimizes the doubling and tripling crystals in the laser, centers the laser beam in the receiver field-of-view, interchanges interference filters for a calibration procedure, acquires the data, processes it in real time, and produces real-time displays of the data. The measurements will be automatically transferred to the central CART computer system over ethernet, where they will be integrated into the automated data processing environment of the ARM data system.

The Raman lidar system described here provides an excellent match to the scientific requirements for continuous profiling of atmospheric water vapor at the SGP CART site. Because the system is under final construction even as this is being written, we cannot present any measurements at this time. However, our previous experience with Raman lidar indicates that this system can provide measurements of the quality desired. Much of the challenge in building the system comes from the desire to make the transition from a research-style system that is partially automated to a field-hardened, CART-ready system that has a greater degree of computer automation and requires significantly less operator attention.

Table 1. System specifications.

Transmitter	
Laser	Nd:YAG third harmonic
Wavelength	355 nm
Energy/pulse	400 mJ
Repetition rate	30 Hz
Divergence	0.1 mr
Receiver	
Configuration	Dall-Kirkham
Diameter	61 cm
Channel bandpass	0.3 nm (narrow fov)
	1.2 nm (wide fov)
Filter transmission	30-40%
Field of view	Dual, adjustable
	(typically 0.2 mr, 2 mr)
Ranges	2 (narrow, wide fov)
Channels	3
Species	Rayleigh/aerosol (355 nm)
	Water vapor (408 nm)
	Nitrogen (387 nm)
Electronics	
Short range	Analog to Digital Conv.
Long range	Photon counting
Averaging period	1 minute
Range resolution	15 m A/D
	37.5 m photon counting

Variability of Radiatively Forced Diurnal Cycle of Intense Convection in the Tropical West Pacific

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Introduction

Strong differences occur in daytime versus nighttime (DVN) net radiative cooling in clear versus cloudy areas of the tropical atmosphere. Daytime average cooling is approximately $-0.7^{\circ}\text{C}/\text{day}$, whereas nighttime net tropospheric cooling rates are about $-1.5^{\circ}\text{C}/\text{day}$, an approximately two-to-one difference (Gray and Jacobson 1977). The comparatively strong nocturnal cooling in clear areas gives rise to a diurnally varying vertical circulation and horizontal convergence cycle. Various manifestations of this cyclic process include the observed early morning heavy rainfall maxima over the tropical oceans. The radiatively driven DVN circulation appears to strongly modulate the resulting diurnal cycle of intense convection which creates the highest, coldest cloudiness over maritime tropical areas and is likely a fundamental mechanism governing both small and large scale dynamics over much of the tropical environment.

Data and Methods

Data include three-hour (i.e., eight images/day) satellite infrared (IR) and visible data for a total of 36 partly current months, extending from the mid 1980s through 1992. The satellite data are the International Satellite Cloud Climatology Project B-1 data, which have a 10-km radiance sampling interval. Periods of study emphasize the summer and winter months with, at this point, comparatively less analysis for the transition seasons. Depending on our purpose, data are expressed either as emittance values (Watts/m^2) or as Equivalent Blackbody Temperatures (EBTs).

Results

The map and analysis in Figure 1 show the distribution of EBT-based pixel temperatures colder than -55°C , expressed as the percent of total pixel counts in each of 160 five-degree (lat/lon) "bins." As the domain spans seven time zones, local time for the (00Z) data in Figure 1 ranges from 0600 in the west to 1200 in the eastern most areas.

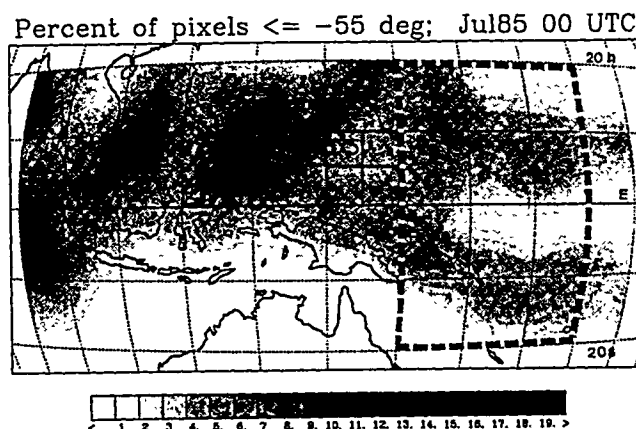


Figure 1. Map showing analysis domain for GMS OLR data. Basic spatial units are five by five latitude-longitude areas, four of which are shown outlined by solid lines in the area north of Papua New Guinea; the large dot on the equator at 140°E shows the approximate satellite sub-point and the large area outlined between 150°E to 180° , 20°N to 20°S is the subject of analyses in Figures 2 and 3. Shading shows areas of intense convection at 00Z during July 1985.

Retaining the data as "pixel counts" in the resolved B-1 mode for fairly narrow temperature bands offers greater accuracy for studying the diurnal cycle than does creating various time or space averaged values.

Observations include a comparatively sharp drop-off of pixel counts at temperatures below -45°C . This effect is roughly consistent with the prior observation of -45°C as the coldest IR temperature normally observed for residual thin stratus clouds. Another sharp drop of pixel counts occurs for temperatures colder than -65°C . Prior work suggests that the mixing ratios associated with maximum sea surface temperature values observed in the West Pacific warm pool correspond to latent heat release and buoyancy of undiluted convective parcels up to approximately 150 mb.

Hence, as the observed mean air temperature is about -65°C at 150 mb, the rapid decrease in the incidence of pixels colder -65°C is consistent with comparatively small areas of actively overshooting convective plumes at these -65°C temperatures.

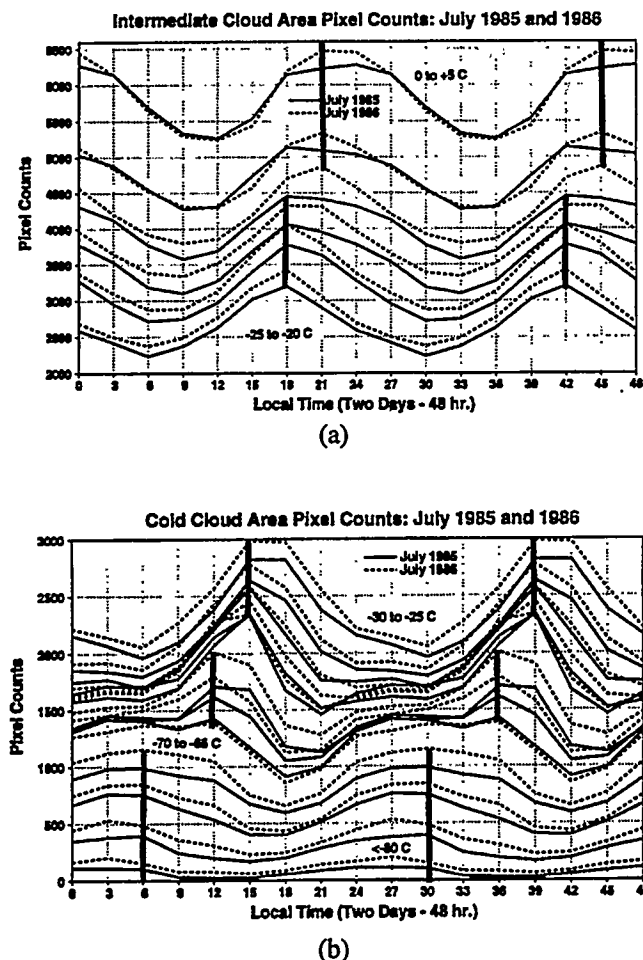


Figure 2. Diurnal variation of pixel counts for the large area enclosed by 20°N to 20°S , 150°E to 180° shown in Figure 1. For clarity, two full daily cycles (48 hours) are shown beginning at midnight local time. The 12 temperature ranges in the lower panel span five degree increments, beginning at -80°C to -85°C (bottom) and warming to -25°C to -30°C (top). The dashed lines are for July 1986 (El Niño) versus solid lines for July 1985 (La Niña). Top Panel: As in the lower panel but for seven intermediate temperature ranges extending from -20°C to -25°C (bottom) to 0°C to $+5^{\circ}\text{C}$ (top). The two sets of vertical bars highlight temperature regimes maximizing at 1800 and 2100, respectively.

The analyses in Figure 2 show the diurnal cycle of pixel counts for a total of eighteen five-degree ranges spanning temperatures from colder than -80°C , at the bottom of lower panel, through $+5^{\circ}\text{C}$, at the top of the upper panel. Figure 3 shows diurnal cycles of total outgoing longwave radiation (OLR) emittance for warm temperature ranges ($+5$ to $+30^{\circ}\text{C}$). Data for two years, 1985 and 1986, are included in Figure 2, to show the year-to-year consistency as well as differences linked to El Niño (1986). One thousand pixels correspond to about one percent of the large sub area in Figure 1; hence, all of the area colder than -65°C occupies only about 3% of the special area, whereas the emittance curve in Figure 3 represents about 70% of the area.

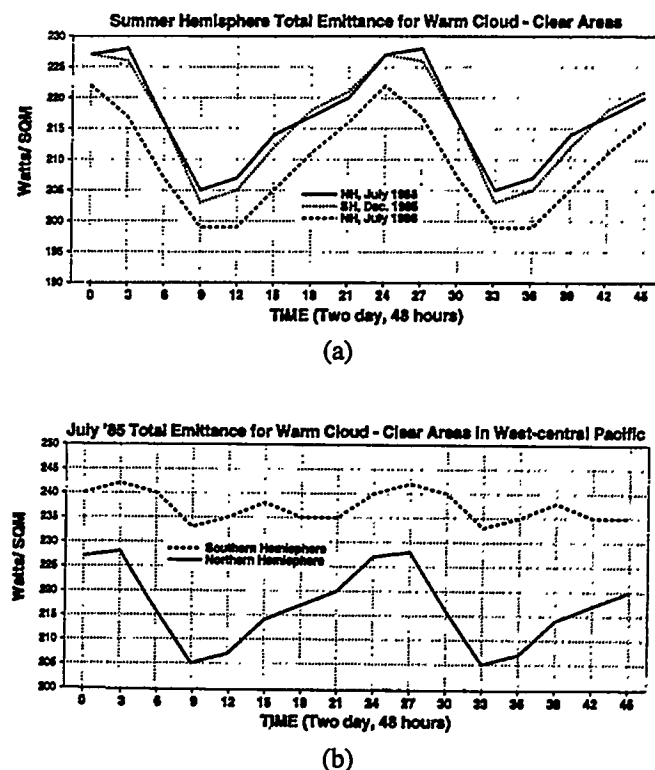


Figure 3. Comparative diurnal cycles of warm area ($+5$ to $+30^{\circ}\text{C}$) emittance for portions of the analysis area in Figure 1. Top panel shows cycles for July 1985 (La Niña), July 1986 (El Niño), and December 1985; data are for the summer hemisphere so that July plots are for $0-20^{\circ}\text{N}$, whereas December plots are for $0-20^{\circ}\text{S}$ (150°E to 180°). Bottom panel compares concurrent "summer" and "winter" hemisphere cycles for July 1985.

The very coldest clouds and most intense convection represented in Figure 2 reach their maxima around 0300 to 0600 local time, whereas somewhat warmer clouds associated with stratiform cirrus clouds (i.e., warmer than -65°C) initially increase in phase with the very coldest clouds but go on to achieve appreciably stronger maxima later in the day, in a 1200-1500 time frame. A distinct phase propagation of increased pixel incidence occurs starting from the coldest EBT values in the early morning through the -20 to 0°C temperature domains by the late afternoon. This trend is accentuated in Figure 2 by the sequence of vertical bars.

The total area of relatively clear conditions and warmer temperatures represented in Figure 3 ($\text{EBT} > +5^{\circ}\text{C}$) increases closely in phase with the intense ($< -65^{\circ}\text{C}$) convection, also reaching a morning maximum. The temporal concurrence of the maxima of coldest and warmest areas reflects the influence accelerating rates of subsidence in the "clear" areas manifest as increasingly larger areas of warm EBT; this trend begins at sunset and accelerates during the night, leading to maximum warm area extent in the late morning hours.

Discussion

The diurnal march of interactive clear versus cloudy area radiative forcing of the nocturnal maximum of intense tropical convection can be inferred directly from Figures 2 and 3. As seen in the warm area emittance cycles (Figure 3), net heat loss begins increasing at sunset (1800) and accelerates throughout the night. The associated broad scale subsidence in these areas immediately begins to enhance low level convergence into the smaller intervening cloudy areas wherein convection intensifies, creating the observed late night-early morning maximum of very cold EBT values.

During daylight hours this forcing subsides, and the morning maximum of very cold EBT systematically gives way to warmer intermediate values (-55 to -20°C) associated largely with stratiform cloud, which spreads rapidly in late morning to a maximum in late afternoon—after which the cycle begins again. This same sequence is seen to play out over the entire maritime portion of the Tropical West Pacific with only modest seasonal differences in component amplitudes. El Niño-Southern Oscillation-linked differences reflect east-west adjustments of the convective centers, but with the same late night/early morning maximum of intense convection. Future research entails expanding these analyses to include detailed validation and quantification of our observations in the Tropical Ocean Global Atmosphere Coupled Ocean Atmosphere Response Experiment data including microwave data, soundings, and regional precipitation analyses.

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Feasibility of Tropospheric Water Vapor Profiling Using Infrared Heterodyne Differential Absorption Lidar

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Continuous, high quality profiles of water vapor, free of systematic bias and of moderate temporal and spatial resolution, acquired over long periods at low operational and maintenance cost, are fundamental to the success of the Atmospheric Radiation Measurement (ARM) Cloud and Radiation Testbed (CART) program. The development and verification of realistic climate model parameterizations for clouds and net radiation balance and the correction of other CART site sensor observations for interferences due to the presence of water vapor are critically dependent on water vapor profile measurements. Application of profiles acquired with current techniques have, to date, been limited by vertical resolution and uniqueness of solution (e.g., high resolution infrared [IR] Fourier transform radiometry), poor spatial and temporal coverage, and high operating cost (e.g., radiosondes), or diminished daytime performance, lack of eye-safety, and high maintenance cost (e.g., Raman lidar). In the study reported here, we develop system performance models and examine the potential of IR differential absorption lidar (DIAL) to solve some of the shortcomings of previous methods using parameterizations representative of current technologies. These models are also applied to diagnose and evaluate other strengths and weaknesses unique to the DIAL method for this application.

The DIAL technique has been implemented to determine the concentration of water vapor and other species at many laboratories. In its simplest form, DIAL is based on comparing the transmission through the atmosphere of beams at two wavelengths which are absorbed differently by the species being monitored. The ratio of the attenuations of the two beams is given by the Lambert-Bouguer law as $\exp[-n\sigma r]$ where n and σ are the concentration and differential absorption cross-section of the specie and r is the path length. In a lidar, the transmitted beams are reflected back to the receiver by atmospheric backscatter, resulting in a measurement of the column content $2nr$ within the (identical)

outgoing and return paths along the line of sight; returns over different column lengths can be compared to determine the range-resolved concentration. Most DIAL systems used for monitoring water vapor have operated at wavelengths of about 700 nm (Grant 1990a).

Recent developments in infrared laser and detector technology make possible compact systems at eye-safe wavelengths. The problems with earlier infrared lidars operating at 10 μm that used direct detection were in part that photodetectors in the infrared are relatively insensitive. Their principal noise sources arose from background radiation on the photodetector and thermal noise. To overcome noise sources, large, high-pulse-power lasers were needed to act as transmitters. Even so, the ranges achieved have been short. This difficulty can be overcome by use of heterodyne receivers. Heterodyne systems are insensitive to background light and are essentially photon-limited so that pulses of rather low energy can be transmitted. Their principal noise source is coherent fading or speckle in the return signal itself, and a heterodyne lidar must be designed to average this out. It is usual, therefore, to consider that these lidars operate at high pulse repetition frequency (PRF). Only recently have suitable laser sources become available (Pearson 1992; Grund 1995).

Simulations have been used here to compare the likely performance of direct detection and heterodyne lidars using what are essentially currently available infrared lasers operating at 2 μm and 10 μm . Simplifying assumptions are made where possible and where attempts to be unnecessarily exact prejudice understanding of the fundamental problems. We concentrate on statistical sources of error. Absorption data (absorption coefficients and their temperature and pressure dependence) are taken from the HITRAN data base, which may be a source of some systematic error (Grant 1990b). In the simulations shown here, atmospheric

parameters have been based on the values specified in the U.S. Standard Atmosphere Model. Water vapor concentrations have been randomized about these values to add realism and make the processing problem more challenging. Useful range-resolved estimates cannot usually be obtained simply by differencing column content values because of the noise level, so that the range-smoothing and differencing operations needed for both column content and range-resolved estimates are combined using a Kalman filter algorithm (Rye and Hardesty 1989). This also provides as an output an estimate of the standard deviation σ of the water vapor measurement. The backscatter model is one developed previously (Kavaya et al. 1989).

Properties of the systems considered in this summary and parameters of the measurements are listed in Table 1. The total transmitted energies of the systems assumed are 50 J for the CO₂ lidars and 300 J for the 2- μ m lidars. The 50-J, CO₂ systems represent data accumulation over about 10 secs, which would be appropriate for monitoring at lower altitudes (say up to 3 km), and the 300-J, 2- μ m systems accumulate data over about 5 mins, appropriate for higher altitudes. The reason for choosing CO₂ lidars for the lower altitude measurements quoted in this summary and vice-versa is arbitrary; all four types of systems having been evaluated for both applications in the full report (Rye et al. 1995). Examples of range-resolved water vapor concentrations obtained using the four systems considered are shown in Figures 1-4.

Some of the further differences between the heterodyne and direct detection systems are

- diffraction-limited optics are usually used in heterodyne lidars, and the aperture and focusing of these optics should be at least approximately matched to the strength and range of the return; the limited field of view reduces background light, but direct detection systems can make use of larger optics
- the heterodyne signal-to-noise ratio (SNR) is approximately constant out to long range and can be optimized at long range by focusing, whereas the direct detection SNR falls rapidly with range; consequently, very good DIAL results can be obtained using direct detection at short ranges, but the long range results are impaired.

Thus somewhat smaller optics can be used for the heterodyne systems (Table 1), and at 10 μ m the results from the smaller heterodyne system are superior to those from the direct detection lidar at long range but inferior at short range (Figures 1 and 2). In the absence of daylight background (not included in this assessment) and at these transmitted energy levels for the 2- μ m lidars, either of the two receivers might be used on the basis of Figures 3 and 4. A heterodyne 10- μ m system would produce similar results. Other simulations suggest that small heterodyne lidars at both wavelengths, using only 10-sec integration and small optics, should be satisfactory for near-horizontal measurements to ranges of 5 km. The larger systems of Figures 3 and 4 are needed for vertical measurements because of the rapid reduction of backscatter with altitude.

Our results, as exemplified in the figures here, suggest that the reluctance to use IR DIAL based on the likely

Table 1. Properties of simulated lidar systems.

Laser	CO ₂		2 μm	
Wavenumbers	974.62, 975.93 cm ⁻¹ (10R18, 10R20)		4943.887, 4948.046 cm ⁻¹	
Range Gate	60 m			
No. of Range Gates	100			
Receiver	heterodyne	direct	heterodyne	direct
Pulse Energy	5.0 mJ	500 mJ	5.0 mJ	100 mJ
PRF	1 kHz	10 Hz	200 Hz	10 Hz
Number of Pulses	10,000	100	60,000	3,000
Receiver Diameter	0.3 m	1.0 m	0.5 m	1.0 m
Receiver Efficiency	8%	10%	8%	10%
Sampling Frequency	2.5 MHz	2.5 MHz	12.5 MHz	2.5 MHz
Photodetector Noise Equivalent Power	--	6 x 10 ⁻¹³ W/Hz ^{1/2}	--	1 x 10 ⁻¹³ W/Hz ^{1/2}

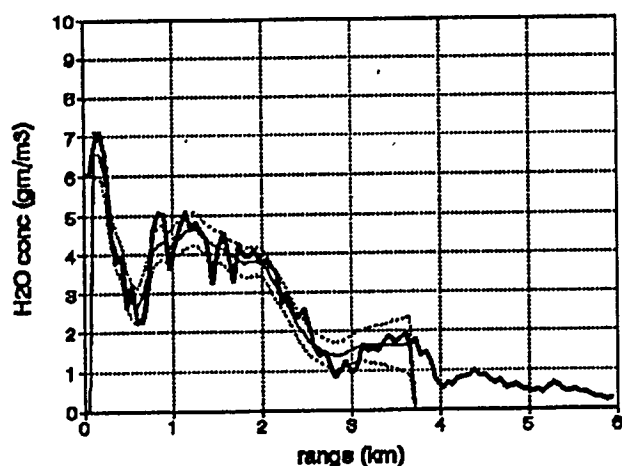


Figure 1. Performance of a "small" CO_2 heterodyne system with a 0.3-m mirror pointing vertically and focused at 3 km, integrating the return from 5-mJ, 1-kHz lidar over 10 secs (50 J total). The heavy solid line shows the "true" water vapor concentration, the solid line shows the lidar estimate, and the dotted lines are $\pm 1 \sigma$ around the estimate. Only 60 (out of 100) range gate data points are processed.

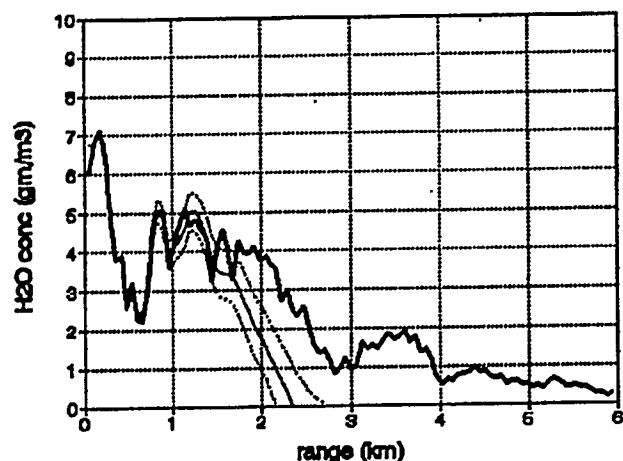


Figure 2. Performance of 50-J CO_2 direct detection lidar for comparison with data of Figure 1. Using a TEA laser, this system might be realized with a 0.5-J laser operating at 100 Hz for 10 secs. A 1-m diameter receiving dish is assumed.

performance of direct detection 10- μm lidars is well-founded, but that compact heterodyne systems operating at either 2 μm or 10 μm with modest current technologies could provide useful, eye-safe, day and night measurements at altitudes up to about 3 km. The statistical uncertainty of the water vapor estimates obtained is

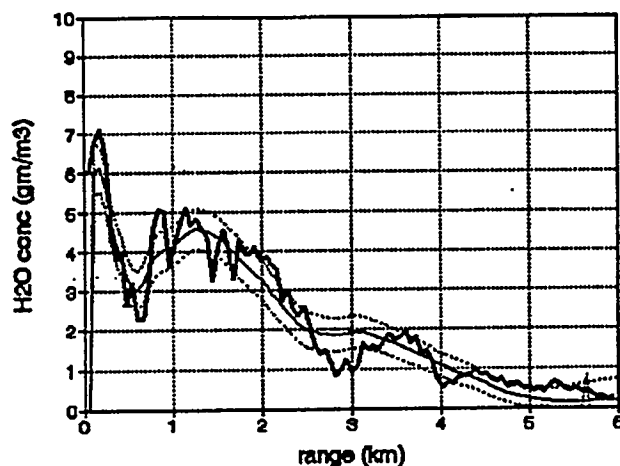


Figure 3. Performance of a 300-J 2- μm heterodyne lidar (5 mJ at 200 Hz for 5 mins) pointing vertically and focused at 5 km. Transmitter and receiver dishes are assumed to be of diameter 0.5 m. Improved results at shorter ranges could be obtained by focusing.

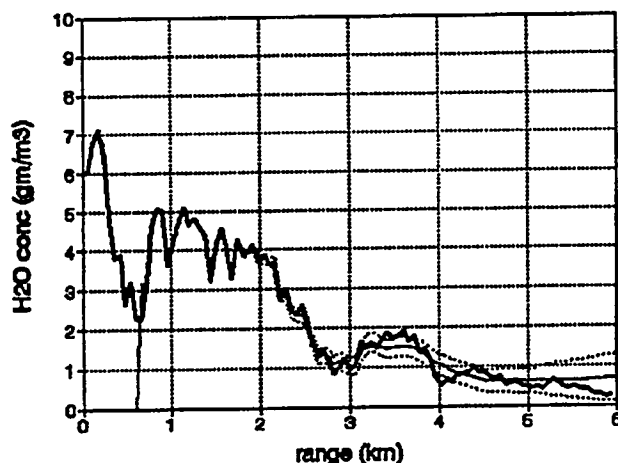


Figure 4. Performance of 300-J (e.g., 100 mJ at 10 Hz for 5 mins) direct detection 2- μm lidar for comparison with data of Figure 3. A 1-m receiver is assumed. The noise equivalent power (NEP) of the photodetector does not take account of daylight background.

predicted to be less than about 20% or 1 gm/m^3 , the range resolution less than about 200 m, and the time resolution 10 secs. The small 2- μm lidars that could be used for this purpose would have optics diameters as small as about 5 inches. At higher altitudes (up to 5 km) it is necessary to rely on long averaging times, and it would have to be demonstrated in practice whether this is

feasible. Alternatively, we shall have to wait on development of more powerful laser sources. This work is to continue in the direction of evaluating yet smaller and lower-cost laser diode-based systems for routine monitoring of the lower altitudes using photon-counting detection methods.

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Retrieving Moisture Profiles from Precipitable Water Measurements Using a Variational Data Assimilation Approach

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Introduction

Atmospheric moisture distribution is directly related to the formation of clouds and precipitation and affects the atmospheric radiation and climate. Currently, several remote sensing systems can measure precipitable water (PW) with fairly high accuracy. As part of the development of an Integrated Data Assimilation and Sounding System in support of the Atmospheric Radiation Measurement Program, retrieving the 3-D water vapor fields from PW measurements is an important problem. A new four dimensional variational (4DVAR) data assimilation system based on the Penn State/National Center for Atmospheric Research (NCAR) mesoscale model (MM5) has been developed by Zou et al. (1995) with the adjoint technique. In this study, we used this 4DVAR system to retrieve the moisture profiles.

Because we do not have a set of real observed PW measurements now, the special soundings collected during the Severe Environmental Storm and Mesoscale Experiment (SESAME) in 1979 were used to simulate a set of PW measurements, which were then assimilated into the 4DVAR system. The accuracy of the derived water vapor fields was assessed by direct comparison with the detailed specific humidity soundings. The impact of PW assimilation on precipitation forecast was examined by conducting a series of model forecast experiments started from the different initial conditions with or without data assimilation.

4DVAR System

The 4DVAR system includes the forward model and its adjoint, and a minimization algorithm. The detailed descriptions of the forward model (MM5) and its adjoint are given by Grell et al. (1994) and Zou et al. (1995). The minimization procedure used the limited-memory quasi-Newton method (Liu and Nocedal 1989). A brief description of the working procedure of the 4DVAR system is as follows:

1. The forward model (MM5) integration started from a guess of the initial condition in the assimilation time window (t_0, t_R) gives the basic states, which are used to calculate the coefficients for its adjoint.
2. The backward integration of the adjoint model with the forcing terms added will give the gradients of the cost function with respect to the initial conditions. The forcing terms are usually calculated from the observed data available within the time window (t_0, t_R) and the model variables.
3. The gradients are used to determine the descent direction, and the initial conditions are corrected based on this descent direction.
4. With the new initial conditions, steps 1 to 3. are repeated. This process continues iteratively and finally produces the optimal initial conditions, which make the model trajectory a best fit to the observations within the time window (t_0, t_R) .

Here the cost function J is defined as

$$J(x_0) = \beta(qs(t_R) - qs_{obs}(t_R))^2 + \gamma(PW(t_R) - PW_{obs}(t_R))^2 \quad (1)$$

Where x_0 represents a vector of the model condition at $t = t_0$ and qs , PW , and qs_{obs} , PW_{obs} are the model and observed surface humidity and precipitable water at $t = t_R$. β and γ are the weighting coefficients.

Observed Data

The 3-hour soundings and the surface observations collected during the SESAME I case from 1200 UTC 10 to 1200 UTC 11 April 1979 were objectively analyzed to a 40-km grid using the Cressman scheme (Figure 1). The observed precipitable water is computed from the vertical integration of the specific humidity:

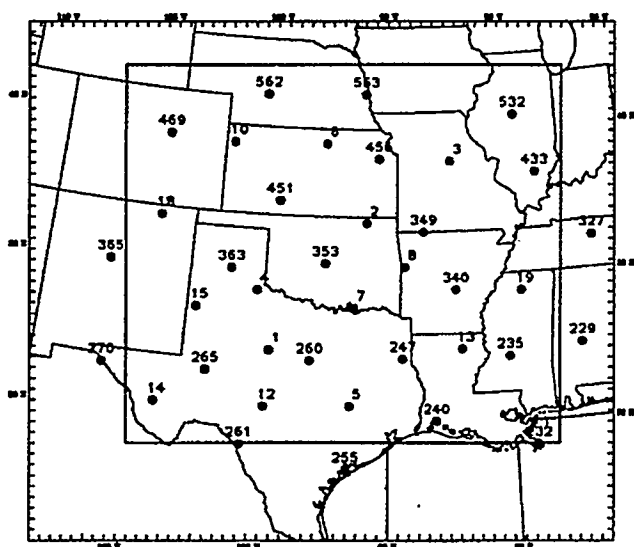


Figure 1. The model domain with the rawinsonde stations plotted. The numbers above the stations are the station ID. Inner box shows the region used for precipitation verification.

$$PW_{\text{obs}} = \frac{\bar{p}}{g} \sum_{k=1}^K q_{\text{obs}}(k) \Delta\sigma \quad (2)$$

where $q_{\text{obs}}(k)$ is the objective analysis of the specific humidity, $\bar{p} = p_s - p_t$ (where p_s is the surface pressure and p_t is the pressure at the model top [100 mb]), $K=10$ is the total number of the model layers, $\Delta\sigma$ is the thickness of the layers (0.1), and g is the acceleration due to gravity. The specific humidity at the lowest model level $q_{\text{obs}}(k)$ was used as the observed surface moisture qs_{obs} . The model PW is calculated in the same manner.

Experiment Design

The data assimilation time window is 1 hour from 1700 UTC (t_0) to 1800 UTC (t_R) 10 April. The initial guess at $t = t_0$ was interpolated from the analyses of 1200 UTC 10 and 0000 UTC 11 April. Four experiments were conducted to assess the impact of the data assimilation (Table 1).

Results

To evaluate the quality of the retrieved specific humidity via the variational data assimilation approach, we calculated the rms, bias, and non-bias errors at 1800 UTC 10 April verified against the soundings observations at 36 stations and the

Table 1. Experiment design.

Exp. Name	1-H 4DVAR		6-H Forecast
	Init. Guess at 1700 Z	Obs. Data at 1800 Z	Init. Cond. at 1800 Z
ANAL	N/A	N/A	Analysis
INTP	N/A	N/A	Interp.
PW	Interp.	PW	From 4DVAR
QSPW	Interp.	PW, qs	From 4DVAR

threat scores for 3-hour accumulated rainfall forecast during the period of 1800 UTC 10 through 0000 UTC 11 April.

Errors

Figure 2 shows that the analyzed specific humidity is very accurate compared with the station observations. The vertically integrated rms error is only 0.15 g/kg. The interpolated q_{INPT} shows errors as large as 2 g/kg at level 8 and 9 and has integrated error of 1.2 g/kg. The assimilation of PW reduced the integrated error to 0.89 g/kg, and the addition of the surface moisture further decreases the error to 0.78 g/kg, which derived mainly from the improvements in the surface moisture field.

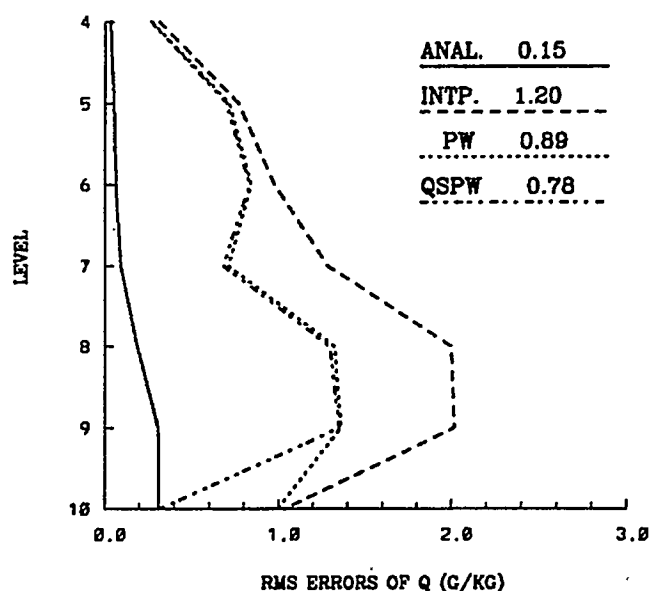


Figure 2. The rms errors of specific humidity field at 1800 UTC 10 April for Exp. ANAL, INTP, PW, and QSPW. The numbers above each of line patterns are the vertically integrated rms error.

Figure 3 shows the distribution of the errors for Exp. INTP, PW, and QSPW. Large bias (PW) errors of 1.42, 1.70, and 1.37 g/kg are found at station 353, 260, and 13 for Exp. INTP, respectively (Figure 3a). The rms errors are 2.04, 3.03, and 1.79 g/kg at these 3 stations (Figure 3b). As expected, the assimilation of PW or PW and qs effectively removed the bias errors. As a result the bias (PW) errors for Exp. PW and QSPW are close to zero (not shown), and the rms errors and non-bias errors are almost identical.

Figure 3c and 3d showed that assimilation of PW or PW and qs reduced the errors to 1.36, 1.53, 1.18 and 1.14, 1.41, 1.11g/kg at the 3 stations, respectively (Figure 3c and 3d).

We also noticed two other interesting stations, 1 and 19. The rms errors for Exp. INTP are rather large (Figure 3b). Figure 4 shows the vertical moisture profiles at these two stations. Station 1 has very small bias error (0.18 g/kg). Therefore, the assimilation of PW alone would not be effective. However, the assimilation of PW and surface moisture reduced the errors dramatically from 3.73 to 1.33 g/kg (Figure 4a). Station 19 has a small bias (PW) error (0.50 g/kg) as well as the small surface moisture error (Figure 4b). According to the cost function in section 2, the forcing terms at this station were small, so the assimilation of PW and the surface moisture would not lead to improvement on this complicated vertical moisture structure.

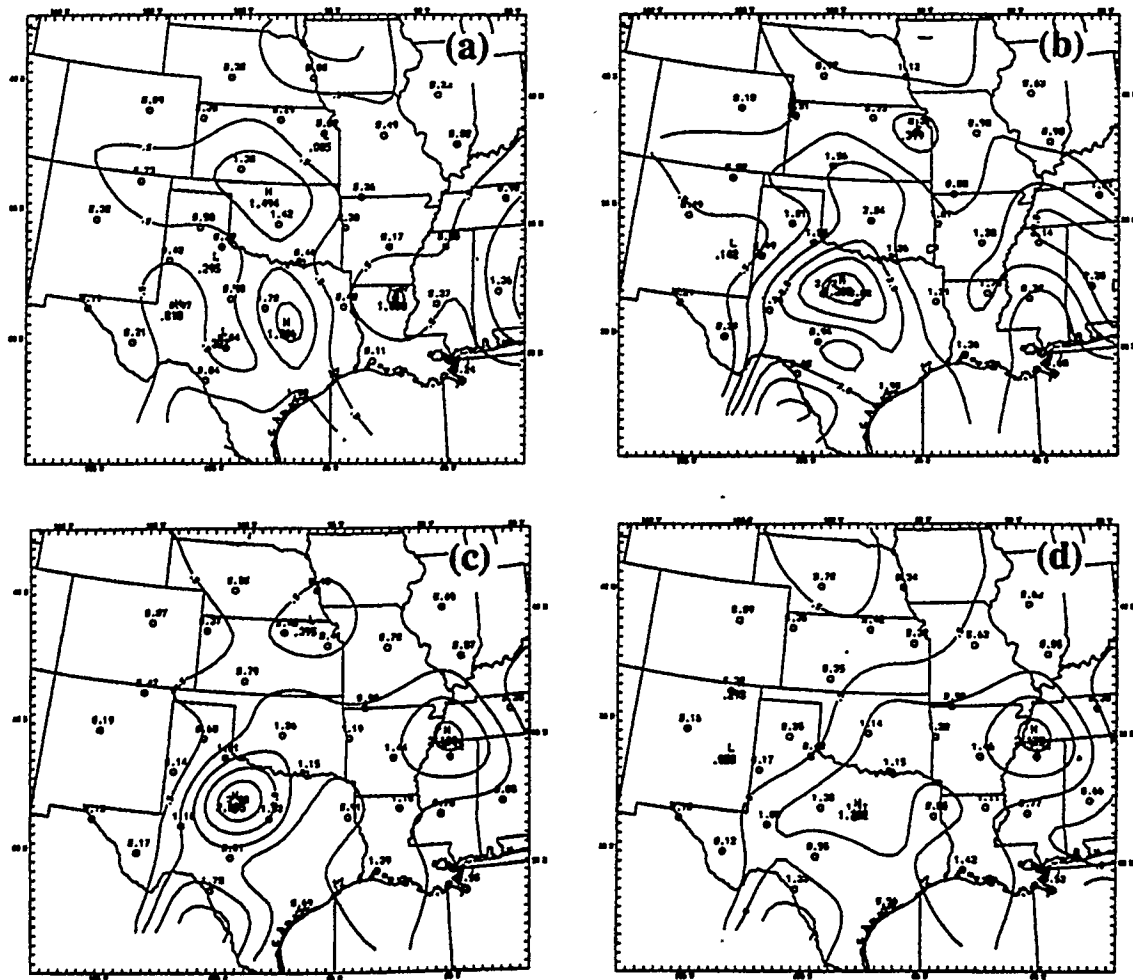


Figure 3. The distribution of the (a) bias errors and (b) rms errors of the specific humidity for Exp. INTP, and the distribution of non-bias errors of the specific humidity for (c) Exp. PW and (d) Exp. QSPW at 1800 UTC 10 April. The contour interval is 0.5 g/kg.

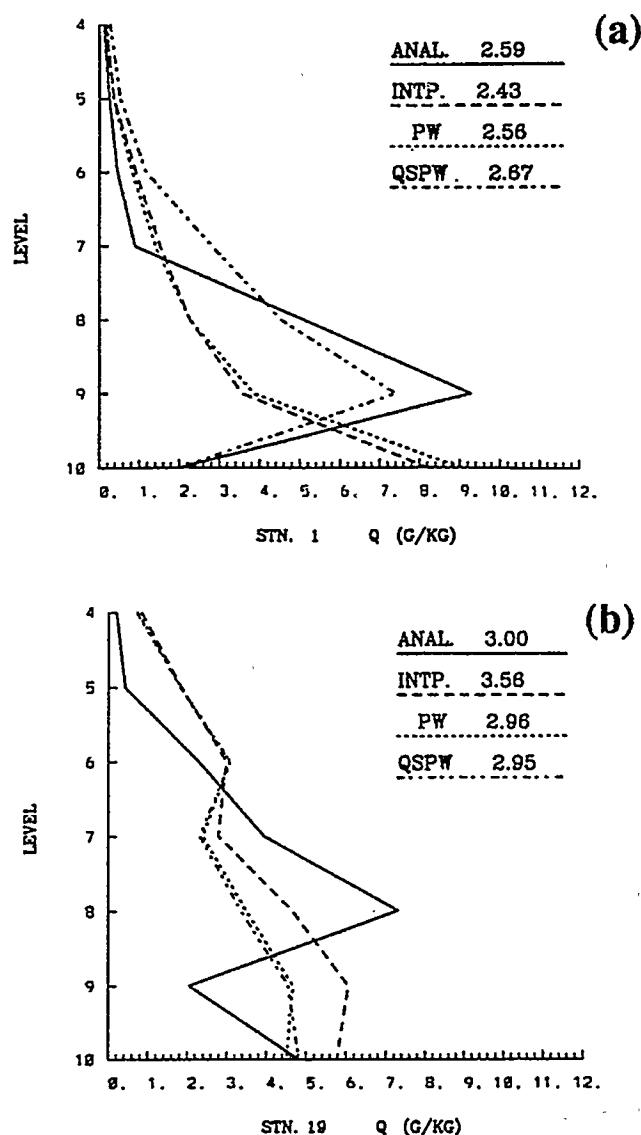


Figure 4. The vertical moisture profiles at (a) station 1 and (b) station 19 for Exp. ANAL, INTP, PW, and QSPW. The numbers above each of the line patterns are the mean values of the specific humidity (g/kg).

Verification of Precipitation Forecast

Figure 5 shows that the Exp. INTP has a very low precipitation forecast skill as verified against the forecast of Exp. ANAL, which has the detailed moisture analysis. The threat scores are close to zero. The assimilation of PW data improved the precipitation forecast considerably. Further improvements were found with the addition of surface

moisture data in the assimilation cycle. Assimilation of the

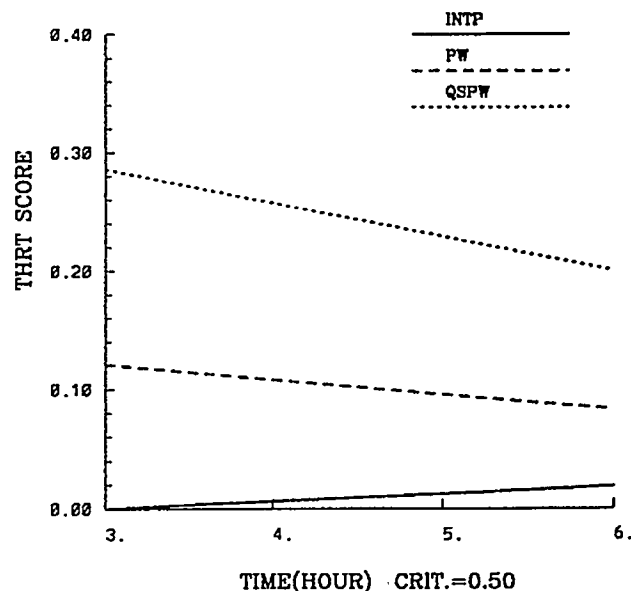


Figure 5. Threat scores of 3-h precipitation forecast at a threshold of 5 mm from 1800 UTC 10 to 0000 UTC 11 April for Exp. INTP, PW and QSPW verified against the forecast from Exp. ANAL.

moisture alone was certainly not enough to get a good precipitation forecast skills (TS is only 0.2~0.3; Figure 5). The best results were obtained when the PW and surface moisture data were assimilated in combination with the wind and temperature data (TS reached 0.5~0.6; Kuo et al. 1995).

Summary and Conclusions

In this study, the newly developed variational data assimilation system based on the Penn State/NCAR mesoscale model MM5 and its adjoint were used to conduct a series of experiments in retrieving the moisture profiles from the observed PW data. The PW observations at one time level, 1800 UTC 10 April 1979, were obtained from the vertical integration of the observed analysis of the specific humidity during the SESAME I period.

We found that the assimilation of precipitable water measurements can recover the vertical structure of water vapor. The use of surface moisture data can further improve the accuracy of moisture retrieval, particularly in the lower troposphere. The improved moisture analysis as a result of PW (and surface moisture data) assimilation can lead to improved characterization of the moisture field over the

CART domain and, in general, improved short-range precipitation forecast.

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Coupled Interactions of Organized Deep Convection Over the Tropical Western Pacific

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Introduction

The relationship between sea surface temperature (SST) and deep convection is complex. In general, deep convection occurs more frequently and with more intensity as SSTs become higher. This theory assumes that the atmospheric stability is sufficiently reduced to allow the onset of moist convection. However, the amount and intensity of convection observed tends to decrease with increasing SST because very warm SSTs may occur only under conditions of diminished convection. This indicates that convection acts to decrease SSTs. A reason for such decrease is the enhancements to surface fluxes of heat and moisture out of the ocean surface because of the vertical overturning associated with deep convection. Early studies used the radiative-convective models of the atmosphere to examine the role of the convective exchange of heat and moisture in maintaining the vertical temperature profile. In this paper we use a Coupled Ocean/Atmosphere Mesoscale Prediction System (COAMPS) to simulate a squall line over a tropical ocean global atmosphere/coupled ocean atmosphere response experiment (TOGA/COARE) area and to investigate how the ocean cooling mechanisms associated with organized deep convection act to limit tropical SSTs.

Model Description

The three-dimensional COAMPS consists of a non-hydrostatic atmospheric model and a hydrostatic ocean model (Hodur 1993). The models can be integrated simultaneously so that the surface fluxes of heat, momentum, and moisture are exchanged across the air-sea interface every time step. The models have been coupled at the interface by vertical mixing. Both the atmospheric model and the ocean model have used a time-splitting technique for computational efficiency. Short time steps have been used for the terms that govern the sound waves in the atmospheric dynamic equations and for the barotropic mode in the ocean

model. Longer time steps have been used for all other terms in the atmospheric model and for the internal modes in the ocean model.

The model domain is 120 x 120 x 17.5 km with 61 x 61 x 50 grid points in the horizontal and vertical, respectively. It includes 30 layers for the atmosphere and 20 layers for the ocean. Model resolution is 2 km for horizontal and variable resolution for vertical, stretching from 150 m for the lowest layer smoothly to 1000 m for the upper layer. For the atmospheric model, large time step is 6 seconds and the small time step 0.6 seconds. For the ocean model, large time step is 240 seconds and the small time step 6 seconds. The model is started at 1800 UTC before squall line develops. The initial thermodynamic field and the wind field used composite sounding derived from P-3 flight level data and rawinsonde data at Honiara, Guadalcanal (~9° S, 160° E) for 1800 UTC on February 22, 1993 (Trier et al. 1994). Initial thermal perturbation is oriented in the along-line direction, with a radius of 20 km in the along squall line direction, 5 km in the perpendicular squall line direction, and 1600 m in the vertical direction. Maximum magnitude of thermal is 2° K at the center, decreasing to 0° K at the edges. Two simulations are performed: 1) without the atmosphere/ocean coupling and 2) with the atmosphere/ocean coupling.

Results

Simulation without Atmosphere/Ocean Coupling

Without the atmosphere/ocean coupling, the sea surface temperature is fixed in the model. Therefore, effects of the ocean on the deep convection are not included because surface fluxes related to SSTs were not represented correctly. However, the squall line can be simulated using the atmospheric model alone since it is driven by the atmospheric

instabilities. However, the atmospheric instabilities are affected by the surface forcing. Horizontal cross sections of wind vectors and model-calculated radar reflectivity at an altitude of 1 km from a 4-hour simulation are shown in Figure 1. Squall line has been simulated with a pronounced "bow-shaped" bulge in the leading convective line, which is consistent with the observations. A low-level cyclonic gyre near the North (equator ward) end of the convective line has appeared. The cross section of vertical velocity from grid point (0, 31) to (120, 31) shows two updrafts (figure is not shown). One has a maximum velocity of 7.7 ms^{-1} at about 9 km and another has a maximum velocity of 4.86 ms^{-1} at about 4 km. Observations indicate similar features. Figure 2 is the vertical cross section of radar reflectivity from grid point (0, 31) to (120, 31) at hour 5. The squall line with anvil stratus precipitation in the upshear direction and convective precipitation in the downshear direction are shown.

The new convective cell in the leading edge maintains the squall line development.

Simulation with Atmosphere/Ocean Coupling

When the simulation was made with coupled atmosphere/ocean model, the maximum surface fluxes were different from uncoupled simulation because of the surface cooling by precipitation and evaporation. Figure 3 shows the time changes of the domain averaged latent heat and sensible heat fluxes. Solid lines represent uncoupled simulation and dashed lines represent coupled simulation. In the early stages, surface latent heat flux shows no difference between the two simulations. After precipitation, less surface latent heat flux appears in the coupled simulation. Because SST

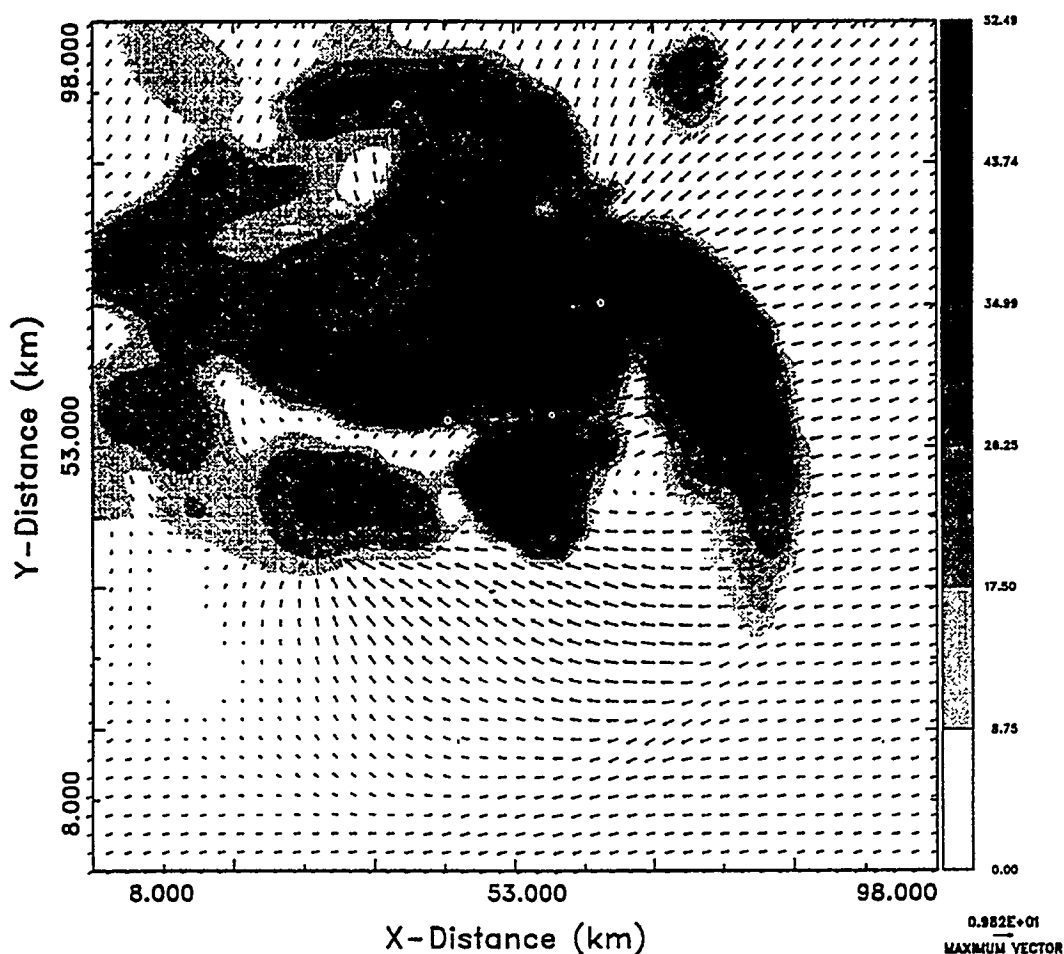


Figure 1. Horizontal cross section at 4 hrs showing the model-calculated radar reflectivity (dBz) and wind vectors at $z = 1000 \text{ m}$. Only a 90-km portion of the domain is displayed.

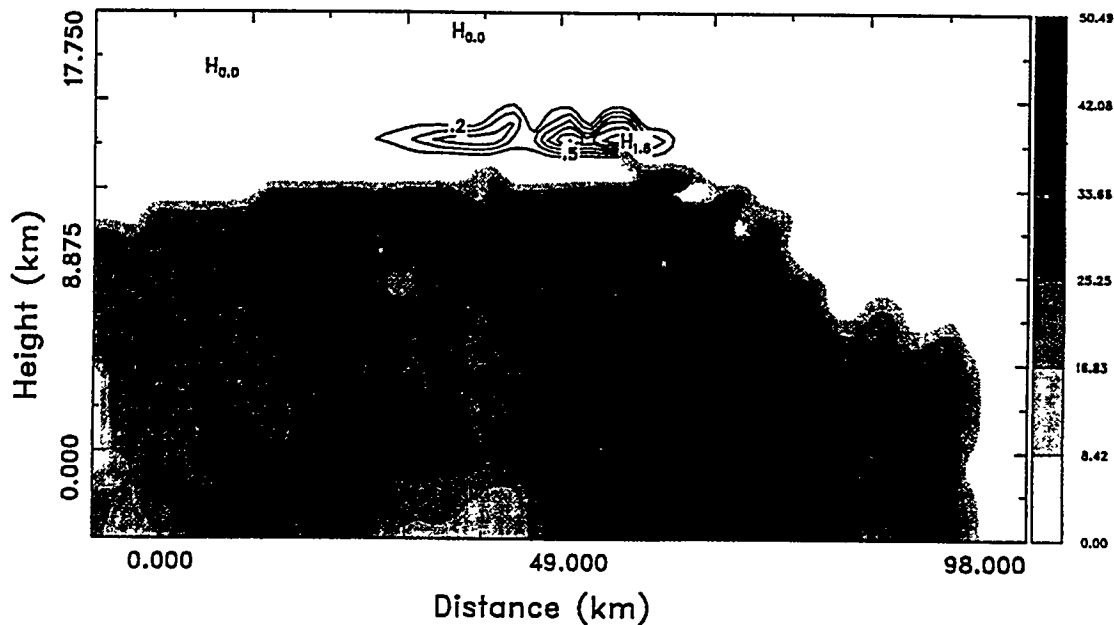


Figure 2. The vertical cross section at 5 hrs showing the model-calculated radar reflectivity (dBZ) and ice crystals (g/kg). The gray scale bar indicates the range and values for the radar reflectivity and contours represent ice crystals. The cross section is taken at $y = 31$.

has been fixed in the uncoupled simulation, unchanged SST overestimates surface heat fluxes due to a larger gradient of temperature between the surface and the first layer of the atmospheric model. Feedback from the deep convection decreases the SSTs because of evaporative cooling (figure is not shown). This indicates that the convective exchange of the heat and moisture can change the vertical temperature profile. The ocean cooling mechanisms associated with organized deep convection can act to limit tropical SSTs.

Acknowledgments

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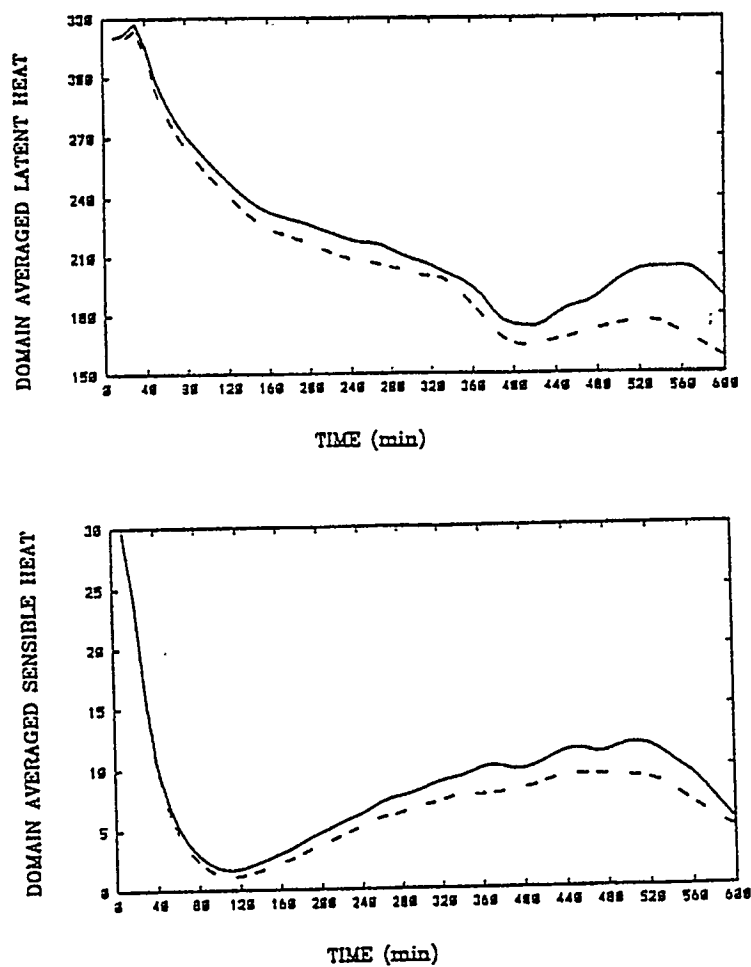


Figure 3. The time changes of domain averaged latent heat and sensible heat fluxes for the uncoupled and coupled simulations. The solid lines indicate uncoupled simulation and dashed lines indicate coupled simulation.

Radiative Transfer in Atmosphere-Sea Ice-Ocean System

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Introduction

Radiative energy is critical in controlling the heat and mass balance of sea ice, which significantly affects the polar climate. In the polar oceans, light transmission through the atmosphere and sea ice is essential to the growth of plankton and algae and, consequently, to the microbial community both in the ice and in the ocean. Therefore, the study of radiative transfer in the polar atmosphere, sea ice, and ocean system is of particular importance. Lacking a properly coupled radiative transfer model for the atmosphere-sea ice-ocean system, a consistent study of the radiative transfer in the polar atmosphere, snow, sea ice, and ocean system has not been undertaken before. The radiative transfer processes in the atmosphere and in the ice and ocean have been treated separately. Because the radiation processes in the atmosphere, sea ice, and ocean depend on each other, this separate treatment is inconsistent. To study the radiative interaction between the atmosphere, clouds, snow, sea ice, and ocean, a radiative transfer model with consistent treatment of radiation in the coupled system is needed and is under development.

Model Description

The radiative transfer equation for the coupled system has been solved by the discrete ordinate method (Jin and Stamnes 1994). This solution automatically takes into account the refractive index change at the air-ice or air-ocean interface. Therefore, the model considers the ice and ocean just as additional "atmospheric" layers; and the number of streams applied in the atmosphere, sea ice, and ocean is flexible and can be chosen to satisfy the competing requirements of computational time and accuracy. A detailed description of the radiative transfer model for the coupled atmosphere-sea ice-ocean system has been presented by Jin et al. (1994). Here we only give a brief description.

The physical processes considered in the model include

- absorption and scattering by atmospheric molecules, clouds, and aerosols
- absorption and scattering by snow
- absorption by pure ice, and absorption and scattering by inclusions in sea ice, such as brine pockets and air bubbles
- absorption and scattering by sea water and by hydrosols in the ocean.

The optical properties of each stratum are treated as follows:

- Optical properties (extinction coefficient, single scattering albedo, and scattering asymmetry factor) of clouds are parameterized in terms of the equivalent radius (ER) and the liquid water content (LWC) of the clouds (Slingo 1989; Tsay et al. 1989).
- Optical properties of snow are based on Mie calculations (Wiscombe and Warren 1980).
- The absorption coefficient for pure ice is from a compilation by Grenfell and Perovich (1981). The optical properties of brine pockets, air bubbles, and solid salts are based on Mie calculations. The volume fractions of these inclusions required in the Mie computations are based on the formulation developed by Cox and Weeks (1983). This formulation, which is based on the phase equilibrium constraints, relates the volume fraction of brine pockets, air bubbles, and solid salts to ice temperature, density, and salinity.
- Optical properties of sea water are from Smith and Baker (1981).

Input parameters for the model include

- incident spectral radiation at the top of atmosphere
- profiles of temperature, pressure, and gas concentrations in the atmosphere
- ER and LWC of clouds, cloud height, and thickness
- surface temperature and snow conditions
- profiles of temperature, salinity, and density in the ice; or profiles of volume fractions of gas and brine inclusions in the ice
- vertical distribution of hydrosols in ocean.

These input parameters are used to compute the optical properties in each layer of the atmosphere, snow, sea ice, and ocean, which are subsequently used to compute the radiative quantities.

Examples of Model Application

Figure 1 shows the solar energy partitioning in the atmosphere, sea ice, and ocean (including open ocean and ocean under the ice) as a function of the area fraction of open water, such as leads and polynya within the ice field. The solar energy is integrated over the spectral region of 0.25 μm to 4.0 μm . Here the vertical distance covered by the three different shades gives the fraction of absorption by the atmosphere, sea ice, or ocean. In the model computation, the input parameters for the atmosphere are from the McClatchey et al. (1972) model atmosphere for the subarctic summer. The ice is assumed to have a thickness of 2.0 m, salinity of 5%, and surface temperature of -10°C linearly increasing to -2°C at the ice base. The vertical variation in the properties of seawater is neglected so that the ocean is considered as a single homogeneous layer. As expected, as the fraction of the open ocean increases, the absorption in the ocean rapidly increases and absorption in the ice rapidly decreases, whereas the absorption in the atmosphere is not sensitive to the variation of the open ocean fraction.

Figure 2 shows a comparison of the observed spectral albedo for melting multiyear ice with model calculation. The observational data are from Grenfell and Maykut (1977). Parameterized profiles of salinity and air volume developed by W. F. Weeks^(a) are used in the model computation.

(a) Personnel communication, 1994.

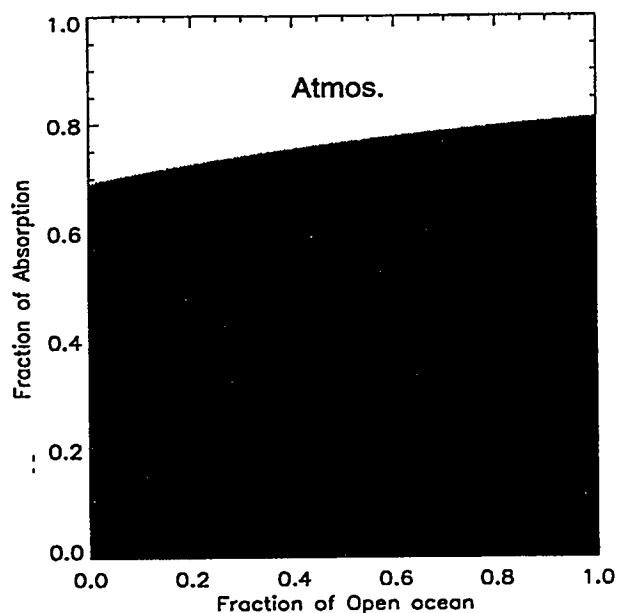


Figure 1. Distribution of the absorbed solar energy in the atmosphere, sea ice, and ocean as a function of open ocean fraction.

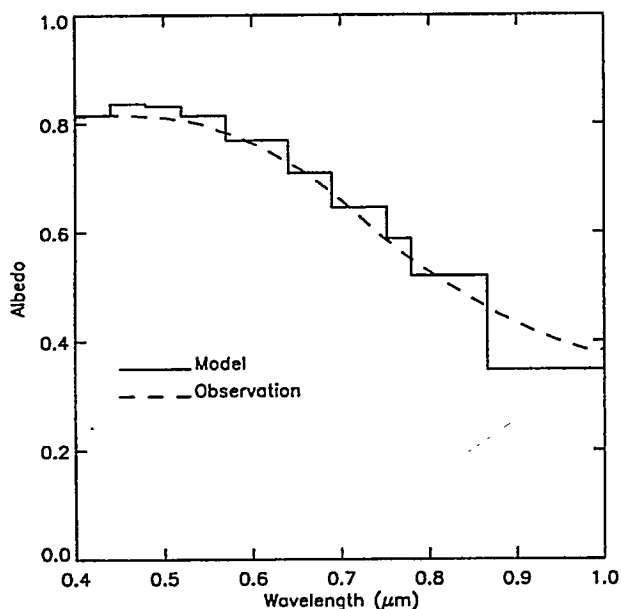


Figure 2. A comparison of the observed spectral albedo for melting multiyear white ice with model calculations.

Figure 3 presents simulated results for a set of irradiance field measurements taken in an Antarctic ice floe (Quakenbush 1994), which had a thickness of 1.24 m. The irradiance measurements in the ice were taken at depths of 20, 40, 60, 80, 100, and 120 cm. The irradiances at different depths of the ice have been normalized by the irradiance at the depth of 20 cm. We use these normalized values for comparison with modeling because 1) the measured irradiance data have not been calibrated in absolute units, so they only represent relative magnitudes; and 2) the angular distribution of the incident light on the ice surface has little effect on these normalized values. The salinity and temperature profiles used in the model for input are based on *in situ* measurements. The absorption peaks at 670 and 430 nm in the measured spectral irradiances indicate that algae exists in the ice. Unfortunately, measurements of the algae concentration were not made. We have therefore assumed a chlorophyll concentration of 10 mg/m³ in the upper 80 cm of the ice and of 40 mg/m³ in the bottom 40 cm of the ice. Results showed that if the algae is ignored, good agreement could be achieved only at wavelengths larger than 700 nm, where algae absorption is small. To obtain better agreement, algae absorption has to be taken into account.

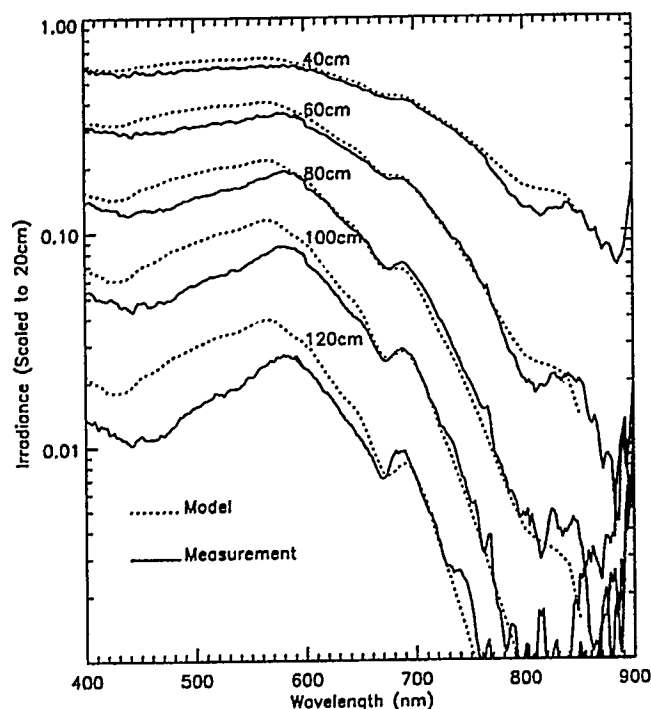


Figure 3. Comparison of the measured spectral irradiances at various depths (normalized to values at a depth of 20 cm) with model calculations.

Some Results and Implications from Modeling

We have used the present model to perform a number of sensitivity tests aimed at establishing the relative importance of each parameter. The modeling results show that the sea ice, clouds, and snow all have a significant impact on the absorption and partitioning of the solar radiative energy in the entire system and on the light transmission into the ice and ocean. Some important findings from the modeling include the following:

- Most of the radiative energy absorbed by sea ice occurs in a very thin top layer of the ice. Under clear sky conditions, only 10 cm of the top layer of ice can absorb more than 50% of the total solar radiation deposited in the entire system.
- In sea ice, it is the scattering by inclusions, especially the air bubbles, in a few centimeters of the uppermost layer that plays a vital role to the solar energy absorption and partitioning in the whole system. Enhanced scattering in this top layer will not only increase backscattering to the atmosphere, but it also increases the fraction absorbed in this top ice layer itself and decreases the radiation penetrating to the deeper layers of the ice and into the ocean.
- Because air bubbles scatter light much more efficiently than brine pockets, the radiative absorption is more sensitive to air volume variations than to brine volume variations.
- Increasing the ice thickness would result in not only an increase of solar absorption in the ice and a corresponding decrease in the ocean, but also a decrease of the absorption in the entire system. On the other hand, the absorption in the atmosphere is not sensitive to the ice thickness. However, the total absorption in the entire system remains almost constant once the ice thickness exceeds about 70 cm.
- Both clouds and snow reduce the solar energy absorption in the ice and in the ocean, as well as in the entire atmosphere-sea ice-ocean system.
- The absorption of solar radiation in sea ice shows a greater sensitivity to the cloud microphysics than the absorption in the atmosphere and in the ocean. The absorption in the entire system will increase as the equivalent radius of cloud droplets increases and will decrease as the liquid water content increases.

- Different from the case in atmosphere, the absorption in sea ice and the ocean is not sensitive to cloud height.
- Clouds not only change the energy disposition in the system, they also reduce the sensitivity of the energy disposition in each layer to the ice and snow thickness change. However, they increase the sensitivity of the surface albedo to the ice or snow thickness variation.
- Surface albedo is determined by approximately 10 cm of the uppermost layer of snow for a snow covered surface and by about 50 cm of the uppermost layer of ice for a bare ice surface.
- Generally, increasing cloud thickness will increase the surface albedo, unless the solar elevation is low and the cloud is thin. Of all the possible cases, the combination of new snow and thick clouds yields the highest surface albedo.
- The occurrence of ice algae has significant effects on the light transport in sea ice and ocean and on the spectral distribution of light transmitted through the ice.

The coupling that we have considered here is purely radiative and does not allow the radiation to alter the snow/ice/cloud properties resulting in changes which could then affect subsequent radiative transfer. To investigate interactions and feedbacks in the polar environment, the present radiative transfer model should be coupled to models treating the evolution of ice, snow, and clouds. To accomplish this goal the computational efficiency of the present model has to be improved, so that it can be coupled to models which require integration over a long time scale.

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Evaluating Aerosol Indirect Effect Through Marine Stratocumulus Clouds

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Introduction

During the last decade much attention has been focused on anthropogenic aerosols and their radiative influence on the global climate. Charlson et al. (1992) and Penner et al. (1994) have demonstrated that tropospheric aerosols and particularly anthropogenic sulfate aerosols may significantly contribute to the radiative forcing exerting a cooling influence on climate (-1 to -2 W/m^2) which is comparable in magnitude to greenhouse forcing, but opposite in sign.

Aerosol particles affect the earth's radiative budget either directly by scattering and absorption of solar radiation by themselves or indirectly by altering the cloud radiative properties through changes in cloud microstructure. Marine stratocumulus cloud layers and their possible cooling influence on the atmosphere as a result of pollution are of special interest because of their high reflectivity, durability, and large global cover.

In this paper we present an estimate of the aerosol indirect effect. More specifically, we focus on the forcing due to anthropogenic sulfate aerosols. In our evaluation of the short wave forcing, we use the data by Warren et al. (1988) on global stratocumulus cloud climatology, the data on pollution due to sulfate aerosols from the 3-D chemical model of Langner and Rodhe (1991), and the parameterization of stratocumulus cloud albedo susceptibility developed using the Cooperative Institute of Mesoscale Meteorological Studies (CIMMS) large-eddy simulation (LES) cloud model with explicit microphysics and radiation (Kogan et al. 1994).

Model Description

In the study we employed the 3-D fields of cloud drop size distributions generated by the CIMMS/National Center for Atmospheric Research (NCAR) cloud model (Kogan et al. 1994). The dynamical framework of the model is based on a

LES code developed by Moeng (1984). The cloud physics formulation follows that of Kogan (1991) and includes explicit formulation of the processes of nucleation, condensation, evaporation, and coalescence. The long wave radiation is parameterized according to Herman and Goody (1976), while the radiation code developed by Bretherton (1994) is used to compute the short wave heating rates. The simulations have been made with 40^3 grid points covering the $(2 \text{ km})^3$ domain with equal spacing of 50 m in all directions. From each cloud simulation, we extracted 1600 vertical profiles of cloud drop spectra corresponding to each cloud grid-column. These profiles represented a wide range of dynamical conditions existing at various spatial locations within the cloud layer.

The developed cloud albedo susceptibility parameterization can be rather accurately represented by the formula:

$$dA/dN = CN^k \quad (1)$$

where $C = 0.044$ and $k = -0.86$. As can be seen from Figure 1, the developed parameterization agrees very well with the calculations based on measurements by Taylor and McHaffie (1994) made during the Atlantic Stratocumulus Transition Experiment field program. The total column sulfate mass loading was obtained from the 3-D chemical Eulerian transport model of Langner and Rodhe (1991). The model simulations were made with horizontal resolution of 10 by 10 degrees in both longitude and latitude and with ten layers in the vertical between the surface and 100 hPa. The difference between distributions of total sulfate aerosols and sulfate aerosols from natural sources only provides the distributions of sulfates due to anthropogenic sources.

Finally, to calculate the spatial distribution of low layer cloud amounts over the globe, we used the two-dimensional climatological data for stratus/stratocumulus clouds (St, Sc), and fog situations by Warren et al. (1988).

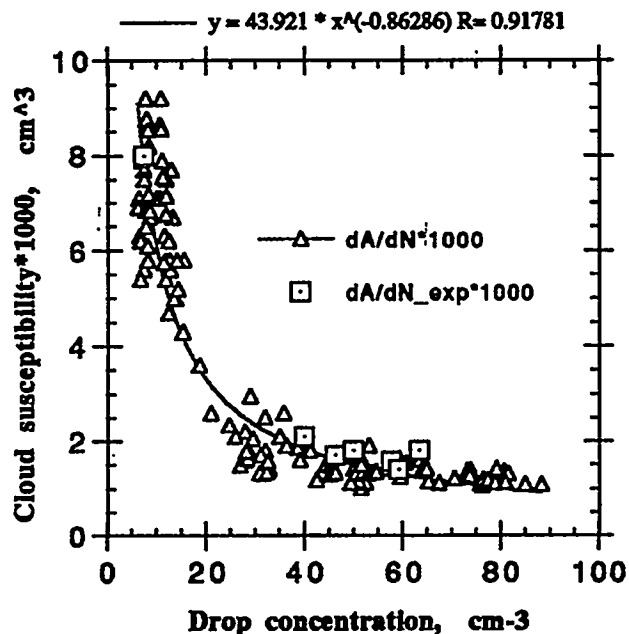


Figure 1. Cloud albedo susceptibility versus drop concentration derived using LES model data (triangles) and the experimental data of Taylor and McHaffie (1994) (squares).

Results

Figure 2 shows zonal averages of sulfate aerosols from all sources as well as from natural sources only. As one may expect, the strongest pollution is in the Northern Hemisphere within the 30°–70°N zone. These areas correspond to the pollution caused by major industrial sources from Europe and the East Coast of the United States. The cleanest air is in the Southern Hemisphere within the 40°–80°S zone; the concentration of particles there is smaller, by approximately a factor of 2, than in the Northern Hemisphere. The level of pollution which may be estimated by the difference between the two lines in Figure 2 is about 5 times less in magnitude in the Southern Hemisphere compared with the Northern Hemisphere. These well-pronounced differences result in a much larger cloud albedo susceptibility of the Southern Hemisphere clouds. As Figure 3 shows, the susceptibility of the Southern Hemisphere clouds is more than twice as large as the susceptibility of the Northern Hemisphere clouds. This result is consistent with our parameterization in Equation 1. Namely, for very clean clouds with drop concentrations on the order of 20 cm⁻³ the susceptibility is about 2 times larger than for the Northern Hemisphere clouds with drop concentrations on the order of 50 cm⁻³.

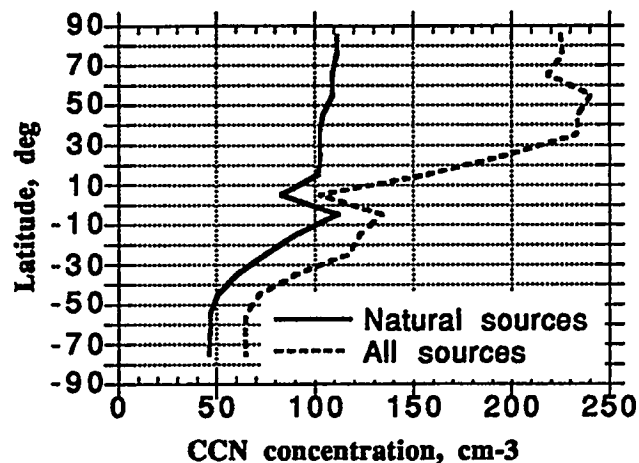


Figure 2. The zonally averaged profiles of the sulphate aerosols. The solid line shows contribution from natural sources only, while the dashed line from all sources.

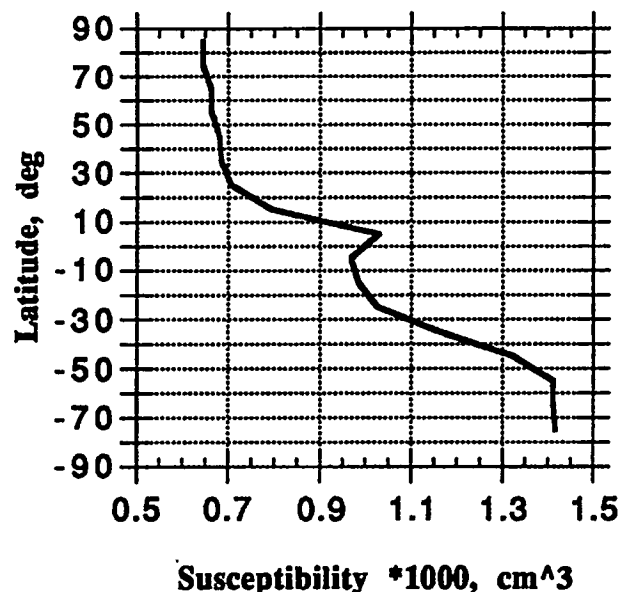


Figure 3. The zonally averaged profile of the cloud albedo susceptibility.

Figure 4 shows the variation with latitude of zonally averaged short-wave forcing due to sulfate aerosol pollution, while Figure 5 shows the distribution of this forcing over the globe (ocean areas only). Figure 4 shows that the maximum forcing is given by regions within the 50°–60°N zone, which is consistent with the high pollution in these areas.

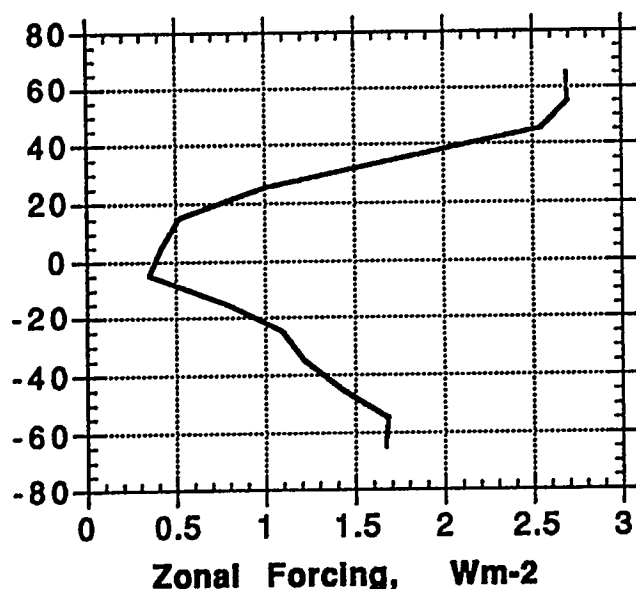


Figure 4. The zonally averaged profile forcing due to the indirect effect of anthropogenic sulphate aerosols.

A striking feature of this figure is the rather large contribution from the high latitudes in the Southern Hemisphere. This is evidently due to the high cloud albedo susceptibility in that zone (Figure 3). Figure 4 shows that, although the pollution level in the Northern Hemisphere is about 5 times larger than in the Southern Hemisphere, the peak forcing here is only 1.6 times larger than the peak forcing in the Southern Hemisphere. Again, this fact can be explained by the negative feedback between sulfate loading and albedo susceptibility.

Our calculations show that the total contribution to the forcing term from each hemisphere differs insignificantly. Namely, the average cooling rate over the whole globe is 1.1 W/m^2 , while the averages for Northern and Southern Hemispheres are 1.3 W/m^2 and 0.9 W/m^2 , respectively. Thus, in spite of the large difference in the pollution loading between the two hemispheres, the difference in climate cooling rates is rather small.

Figure 5 shows that forcing (cooling) may vary substantially over the globe from 0.08 W/m^2 near the equator to about 5.6 W/m^2 near the major pollution sources in the United States and Europe.

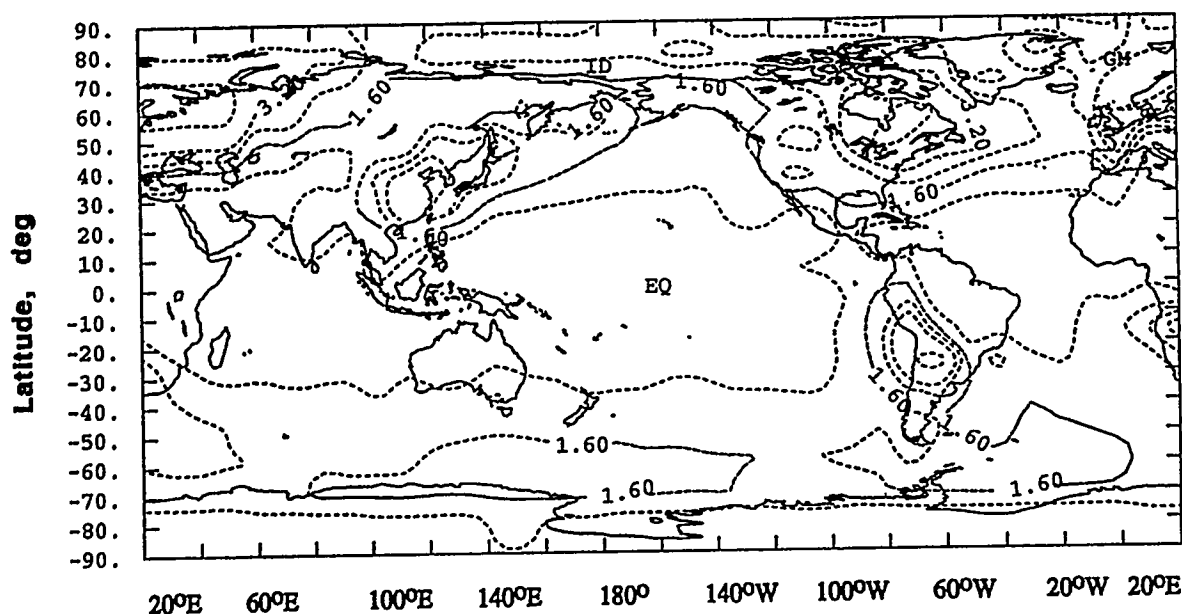


Figure 5. Global distribution of the radiative forcing due to the indirect effect of anthropogenic sulfate aerosols. The isoline increment is 0.8 W/m^2 .

Conclusions and Future Work

We have evaluated the indirect radiative forcing caused by sulfate aerosols using the data on pollution from the 3-D chemical transport model of Langner and Rodhe (1991), cloud climatological data by Warren et al. (1988), and parameterization of cloud albedo susceptibility developed using the CIMMS/NCAR LES cloud model with explicit microphysics (Kogan et al. 1994). The results show that the global average cooling due to indirect forcing is 1.1 W/m^2 , with averages 1.3 W/m^2 and 0.9 W/m^2 over the Northern and Southern Hemispheres, respectively. Our results also show that forcing (cooling) may vary over the globe from 0.08 W/m^2 near the equator to 5.6 W/m^2 near major pollution sources in United States and Europe. These estimates are consistent with those obtained previously by Charlson et al. (1992) and Jones et al. (1994). Our results also show a large difference in cloud albedo susceptibility between the two hemispheres. This leads to a rather small difference in the cooling rates (0.4 W/m^2) between the two hemispheres. As a result, in spite of the large differences in pollution loading, the contribution to the climatic forcing due to anthropogenic aerosols from each hemisphere differs only by about 25%. In our future work we will study the sensitivity of the climatic forcing due to anthropogenic sulfur aerosols to the uncertainties and variability in the input data.

Acknowledgments

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A General Circulation Model (GCM) Parameterization of Pinatubo Aerosols

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The June 1991 volcanic eruption of Mt. Pinatubo in the Philippines is by far the largest and the best documented global climate forcing experiment in recorded history. The time development and geographical dispersion of the Pinatubo aerosol has been closely monitored and sampled by a wide variety of satellite, airborne, and ground based instruments using visible, near-infrared (IR), and thermal wavelength measurements. The principal information regarding the optical depth, particle size, and vertical and geographic distribution of the aerosol is derived from the solar occultation measurements at 385, 453, 525, and 1020 nm taken by the Stratospheric Aerosol and Gas Experiment (SAGE) II aboard the ERBS satellite (McCormick et al. 1995). Balloonborne measurements (Deshler et al. 1992) have identified the aerosol composition as concentrated sulfuric acid and have also characterized the particle size distribution. Limb-viewing measurements taken by ISAMS aboard UARS (Lambert et al. 1993) at thermal wavelengths provide key information about the variance (width) of the particle size distribution. The Pinatubo signature on the global energy balance was also recorded in the Earth Radiation Budget Experiment (ERBE) measurements of top of the atmosphere (TOA) solar and thermal fluxes (Minnis et al. 1993). Additional information obtained from a variety of lidar, sunphotometer, and *in situ* measurements is still being incorporated into the Pinatubo stratospheric aerosol data base to fill in remaining gaps and to cross-calibrate measurement results.

Given the above observational data, the spherical shape of the sulfuric acid droplets, and the real and imaginary refractive indices of sulfuric acid, Mie scattering can be used to calculate the aerosol radiative properties at all wavelengths of the spectrum for a broad range of particle size distributions, thus making possible the accurate determination of the radiative forcing generated by the Pinatubo aerosols. Other observational constraints that help define the Pinatubo climate forcing experiment are provided by MSU measurements of stratospheric and tropospheric temperature (Spencer and Christy 1993) and by the global network of meteorological measurements of surface air temperature.

Based on preliminary estimates of the Pinatubo aerosol loading, Hansen et al. (1992) made general circulation model (GCM) predictions of the impact on global climate due to the Pinatubo aerosol. The predicted decrease in the global surface air temperature by 0.5°C by the end of 1992 with recovery to normal by 1995 has proven to be remarkably accurate, though not all for the right reasons. Since the presentation of these preliminary model calculations, significant improvements have been made to the radiative forcing input and to the Goddard Institute for Space Studies GCM. Model improvements include a more physically based cloud prediction scheme and an improved parameterization for the radiative transfer modeling of aerosols.

The radiative input to the GCM is obtained by least-squares fitting of a mono-modal size distribution to the zonally averaged SAGE II column extinction data at 385, 453, 525, and 1020 nm. The spectral range of the SAGE data is too limited to constrain the size distribution variance. However, extinction at $12.1\text{ }\mu\text{m}$ from ISAMS limb-viewing observations can be used to infer a size distribution variance of order $v_{\text{eff}} = 0.35$. Since nearly all of the stratospheric heating is produced by the absorption of upwelling thermal radiation, the size distribution variance is closely related to the amount of stratospheric heating and also to the amount of greenhouse warming that counteracts the surface cooling effect due to reflected solar radiation.

The time-latitude evolution of the Pinatubo aerosol cloud is shown in Figure 1. The zonal mean column optical depth at $\lambda = 0.55\text{ }\mu\text{m}$ derived from SAGE II measurements is displayed in Figure 1a, and the time-latitude evolution of the effective particle size is shown in Figure 1b. The Pinatubo aerosol was found to increase steadily in size following the eruption, reaching a maximum effective radius of over $0.7\text{ }\mu\text{m}$ near the end of 1993. It is interesting to note that the particle size maximum does not coincide with peak optical depth, occurring instead nearly a year later. The constraints that are placed on the aerosol size distribution by the ISAMS $12.1\text{-}\mu\text{m}$ measurements are illustrated in Figure 2.

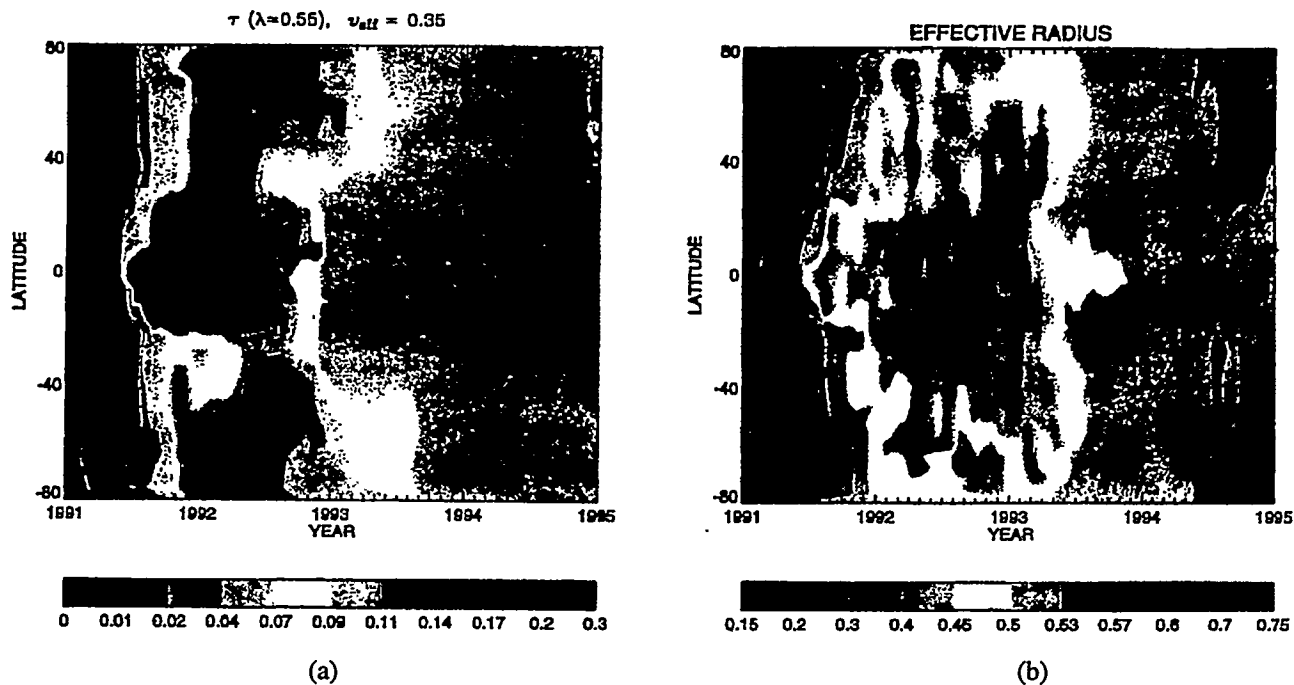


Figure 1. Zonally averaged latitude-time dependence of Pinatubo aerosol optical depth (a) at reference wavelength $\lambda=0.55 \mu\text{m}$, and (b) effective particle radius derived from SAGE II data for effective size distribution variance $v_{\text{eff}} = 0.35$.

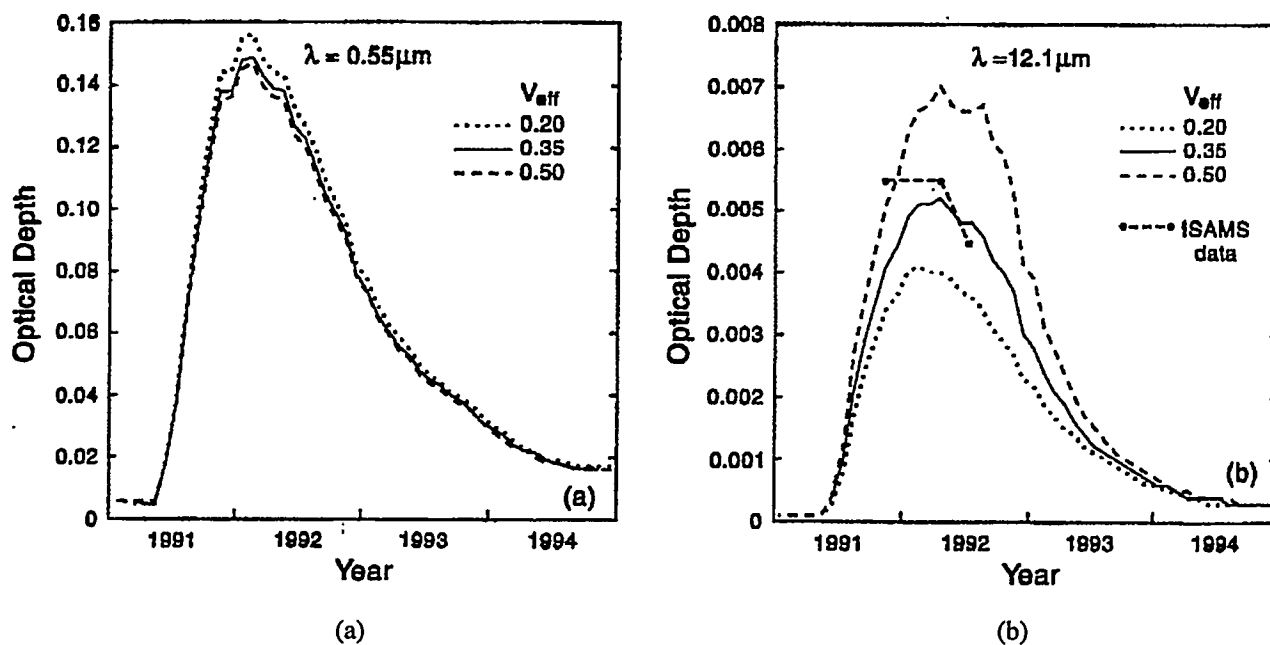


Figure 2. Global mean optical depth at (a) $0.55 \mu\text{m}$ and (b) $2.1 \mu\text{m}$ for several values of effective variance based on multi-spectral extinction measured by SAGE II.

The radiative transfer model used in the calculations is based on the doubling/adding method (Lacis and Hansen 1974). Six spectral intervals are used to represent the wavelength dependence of the aerosol radiative properties in the solar spectrum. The single gauss point adaptation of the doubling/adding method has been parameterized to yield the same solar zenith angle dependence for conservative scattering as obtained by the full angle integrated doubling/adding results. The thermal spectrum is subdivided into 25 non-contiguous k-distribution intervals (Lacis and Oinas 1991). The flux calculations are done without scattering, using Mie-calculated absorption cross-sections and a 3-point numerical quadrature to integrate radiances over emission angle.

Using the radiative modeling parameterization outlined above, the radiative forcing by the Pinatubo aerosol is computed interactively at each gridbox as a function of optical depth, particle size, and variance based on the SAGE II measurements. The distribution of the aerosol with height is prescribed with 5-km vertical resolution, also based on SAGE II measurements. The globally integrated instantaneous radiative forcing is shown in Figure 3 where the

earlier forcing used in the Hansen et al. (1992) study is compared with current results. The broader size distribution used in the current treatment yields a somewhat greater greenhouse effect and a somewhat smaller solar forcing. With the present Pinatubo data base and GCM radiative treatment, the maximum net Pinatubo forcing is about 3.5 Wm^{-2} compared with 4.0 Wm^{-2} obtained previously.

While comparison of the GCM calculated radiation balance anomalies due to the Pinatubo aerosol shows qualitatively good agreement with ERBE measurements (Minnis et al. 1993), some TOA flux differences need further investigation. Some of these differences might be due to the significant changes in stratospheric ozone that were associated with the Pinatubo volcanic aerosol, but have not yet been included in model simulations. Stratospheric water vapor may also have been affected. The possibility of other feedback interactions with the Pinatubo aerosol that have an observable radiative effect clearly cannot be discounted. Thus the Pinatubo eruption with its growing data base of detailed observational constraints presents intriguing possibilities for further investigation and climate model validation.

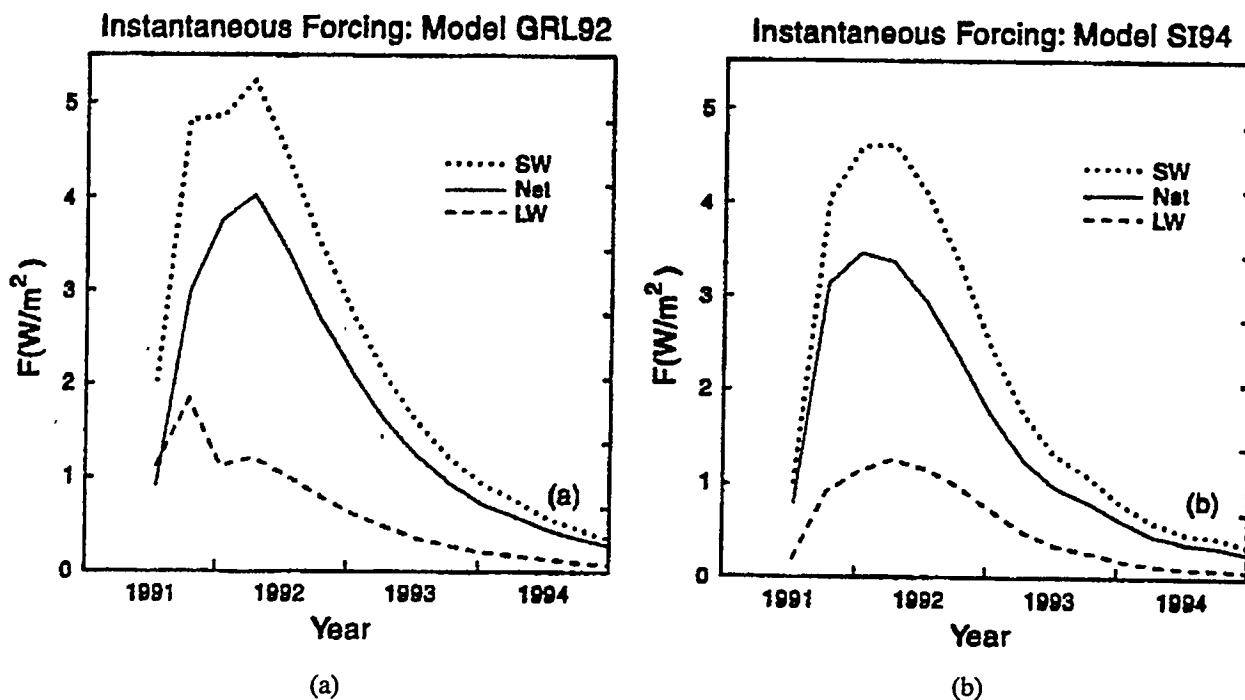


Figure 3. Seasonally averaged instantaneous radiative forcing assumed in the (a) Hansen et al. (1992) aerosol model and (b) based on present analysis of SAGE II data.

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Aerosol Measurements at 60 m During the April 1994 Remote Cloud Study Intensive Operating Period (RCS/IOP)

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In response to a request from Dr. Harvey Melfi of the University of Maryland for aerosol measurements during the April 1994 Remote Cloud Study/Intensive Operating Period (RCS/IOP), the Environmental Measurements Laboratory (EML) responded by modifying an existing EML aircraft aerosol sampler that had been previously tested. This new sampler was installed on the 60-m tower at the Southern Great Plains (SGP) site of the Atmospheric Radiation Measurement (ARM) Program. A 2-hour, battery-operated instrument package was modified into a 30-day automated package capable of sampling at 1-minute intervals. The package was returned to the ground daily for data downloading and calibration of the integrating nephelometer. Data were obtained from the integrating nephelometer ($\lambda = 530$ nm); a six-channel optical particle counter; a condensation particle counter; pressure, temperature, and relative humidity sensors; and mini-impactors. Independent computers inside the integrating nephelometer provided 15-min average data. In addition, twice daily back trajectory analyses (BTAs) were calculated providing some insight into the origins of the air mass sampled.

In the middle of the experiment, two of the controlling computers were destroyed by lightning. The lightning caused loss of all but the integrating nephelometer data which had its own internal computer set to record at a 15-min average interval.

Among one of many reasons for the original selection of the location of the SGP site at Lamont, Oklahoma was the fact that many different types of air masses pass over the area. As seen from Figure 1, air masses come from the Pacific Ocean, central and western Canada, and the Gulf coast. During the early period of the April RCS/IOP, extremely low aerosol scattering coefficients (Figure 2) were seen.

This correlated very well with the BTA which showed air coming from very clean sectors (continent and ocean). During the middle part of the experiment, extremely polluted air, containing very high scattering coefficients, impacted the site. The BTA indicated the transport of air from the Illinois area during this period. During the month, the scattering coefficient varied by more than an order of magnitude. Local and regional plumes can be seen passing over the site. This can be seen in the daily plots of aerosol scattering coefficients and condensation particle concentrations (Figure 3).

A major effort was placed on providing some characterization of the aerosol size distribution for the RCS/IOP. Because of equipment malfunction and the lightning damage, we obtained only one full day of surface and tower measurements. In addition, limited surface measurements for four days were obtained before the instrument package was raised to 60 m. We programmed the computer to provide 1-min samples of six channels of size data during the first 6 hours of the day and then continued with 2-min sampling for the rest of the day.

The data from our sampler is now available from the experimental center. A description of the data, along with a graphical presentation of the data will be available as part of a U.S. Department of Energy/ARM publication.

Acknowledgments

The authors would like to thank the Instrumentation Division for their fine work in developing the tower aerosol package in the limited amount of time given them. Without the help of Dr. Jeff Griffin of Pacific Northwest National Laboratory, who handled the daily downloading of the data, calibration and trouble shooting, and the site operations support staff, we could not have achieved our goals.

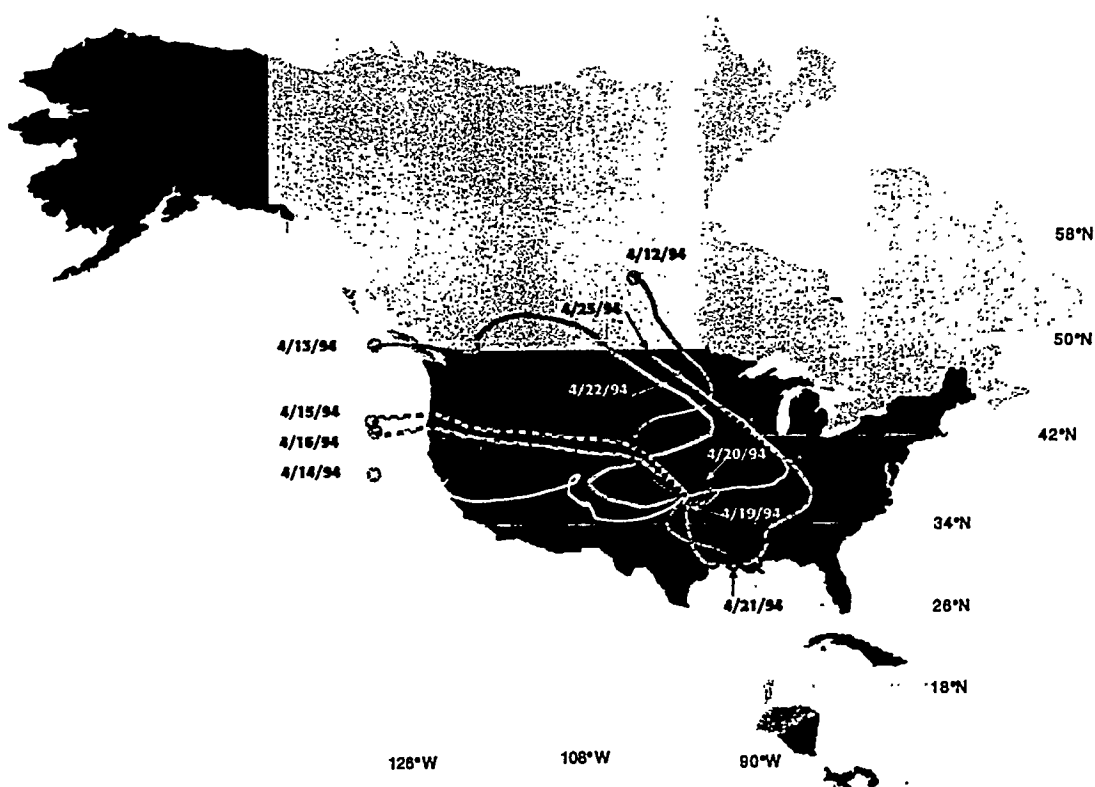


Figure 1. Backward trajectories (BT) for Lamont, Oklahoma. All dates shown are starting times for the BT. The length of the trajectories represents the following times: 4/12/94 lasted 3.5 days, 4/13/94 lasted 9.5 days, 4/14/94 lasted 7.3 days, 4/15/94 lasted 3.6 days, 4/19/94 lasted 0.5 days, 4/20/94 lasted 1.5 days, 4/21/94 lasted 2.5 days, 4/22/94 lasted 3.5 days, 4/23/94 lasted 4.3 days, 4/24/94 lasted 4.5 days, 4/25/94 lasted 6 days.

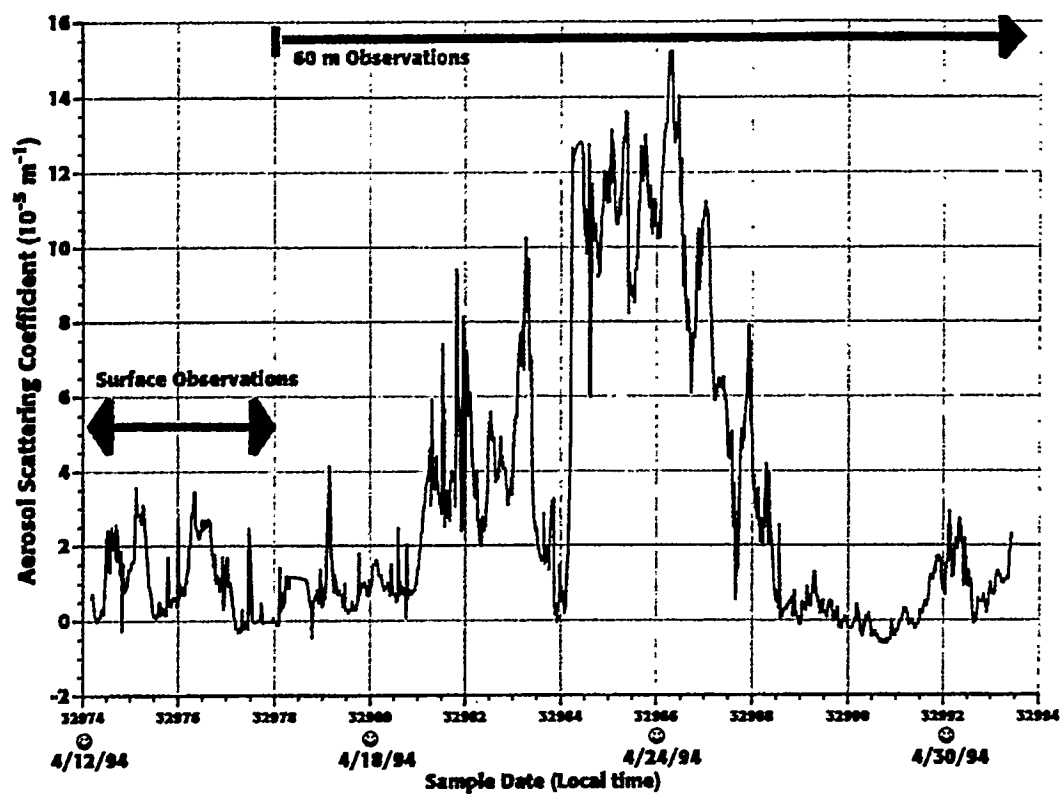


Figure 2. Fifteen-minute average aerosol scattering coefficient ($\lambda = 530 \text{ nm}$) observations during the April 1994 Remote Cloud Study Intensive Operating Period experiment at the ARM SGP site at Lamont, Oklahoma.

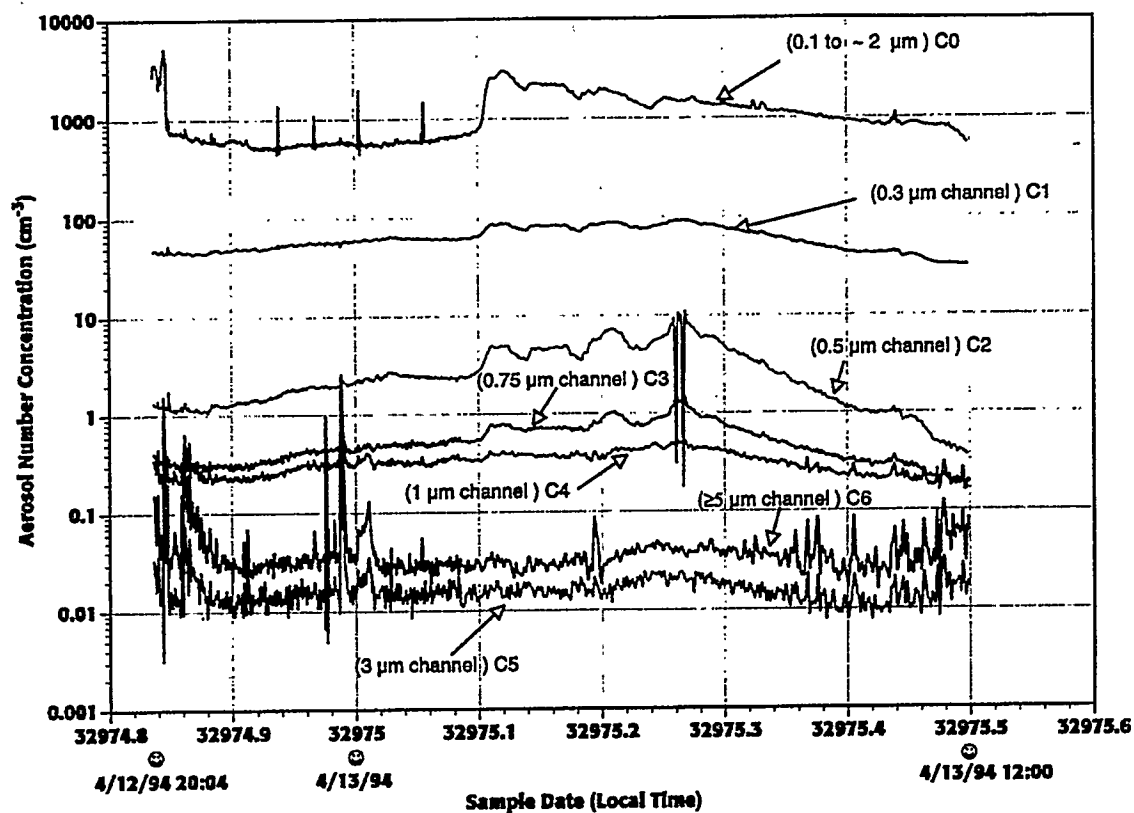


Figure 3. One-minute surface observations of the aerosol scattering coefficients and condensation particle concentration for April 12 and April 13, 1994, at the Remote Cloud Study Intensive Operating Period experiment at the ARM SGP site at Lamont, Oklahoma.

Aerosol Measurements at the Southern Great Plains Site: Design and Surface Installation

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To improve the predictive capabilities of the Atmospheric Radiation Measurements (ARM) Program radiation models, measurements of aerosol size distributions, condensation particle concentrations, aerosol scattering coefficients at a number of wavelengths, and the aerosol absorption coefficients are needed at the Southern Great Plains (SGP) site in Lamont, Oklahoma. In addition, continuous measurements of the ozone concentrations are needed for model validation.

The Environmental Measurements Laboratory (EML) has the responsibility to establish the surface aerosol measurements program at the SGP site. To provide representative aerosol measurements at SGP, EML has designed a special sampling manifold. A stainless steel stack with an inlet set at 10 m provides ambient air flow to the aerosol sampling equipment located in a special trailer dedicated to aerosol measurements. The air entering the aerosol instruments is dried by heating to a controlled temperature in order to reduce the relative humidity (RH) of the sample to below 40%. A separate sampling line made of Teflon is mounted next to the stainless steel aerosol sampling stack. This inlet is set at 10 m to provide ambient air to the ozone monitor. ARM's operating protocol requires that all of the observational data be placed online and sent to the main computer facility in real time. To accomplish this goal, a field data ingestor (FDI; Sun Workstation) located in the aerosol trailer will interrogate and control instruments output via RS-232 serial interface ports. In addition, the field data ingestor will sample analog signals and perform control functions.

A major effort was expended in 1994 by EML to develop the drawings and statement of work for the exterior support structure of the aerosol sampling stack. The documents were completed and submitted to the SPG Site Program Manager. In addition, a sampling manifold was built and setup at EML for testing and interfacing to the FDI. A new three-wavelength nephelometer ($\lambda = 450, 550, 700$ nm) with backscatter capability (90° to 170°) was purchased in 1994 from TSI, Inc., and was tested and incorporated into the existing aerosol manifold. The specifications for this instrument and its integration into the manifold are shown in Table 1 and Figure 1, respectively.

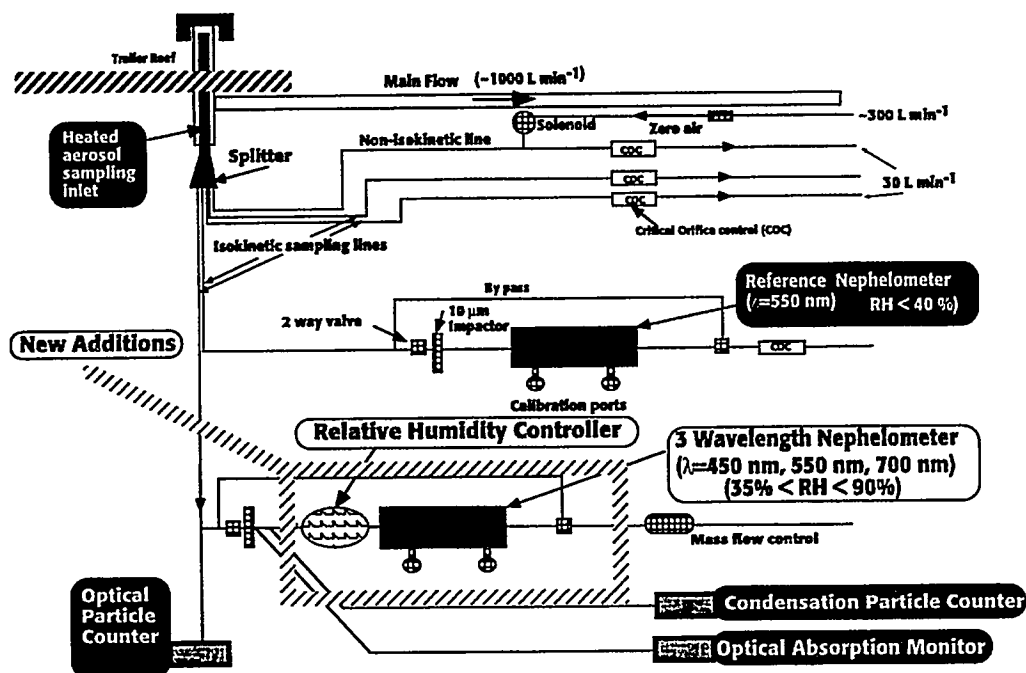
An adjustable RH system is under development at EML (Figure 2) and, together with the three-wavelength nephelometer, will provide a measure of the scattering and backscattering coefficients as a function of RH and wavelength. A second single wavelength ($\lambda = 550$ nm) nephelometer (Radiance Research, Inc.) will be operated as a reference in one of the extra isokinetic sampling lines (Figure 1) under dry conditions (RH < 40%). The humidifier is based on the research efforts of Dr. Mark Rood of the University of Illinois at Urbana-Champaign.

Acknowledgment

The authors would like to thank the Instrumentation Division for their exceptional work in drafting of the necessary working drawings; construction of the sampling stack, support structure, and various manifold components; and completion of the electrical wiring.

Table 1. Specifications for the three-wavelength integrating nephelometer.

Aerosol Scattering Coefficient Operating Range:	10^{-7} to 10^{-3} m^{-3}
Sensitivity:	10^{-7} m^{-1} (450 nm, 550 nm) 3×10^{-7} m^{-1} (700 nm) at 30 s averaging time. Defined as the standard deviation of the measured aerosol scattering coefficient when sampling filtered, particle free air.
Drift:	$<2 \times 10^{-7}$ m^{-1} (30 s average time) for up to 1 h after filtered air reference measurement at 550 nm
Measuring Wavelength:	450 nm, 550 nm, 700 nm, center wavelength (50 nm band width)
Angular Integration:	7° to 170°
Zero Calibration:	particle free air
Zero Calibration Frequency:	user adjustable (1 to 24 h)
Output Analog:	at least 8 channels at 0 to 10 V DC, 0.05% resolution
Output Analog:	eight data channels (aerosol scattering coefficients, Rayleigh corrected, backscattering coefficient for 3 wavelengths); separate pressure temperature and RH channels.
Output Serial:	instrument operating parameters, date (dd, mm, yy), time (h, min), Rayleigh corrected aerosol scattering (σ_{sp} , σ_{bsp} , at 450 nm, 550 nm and 700 nm), pressure, temperature, RH, status
Backscatter Shutter:	changes angular integration from 7° to 170° to 90° to 170°
Backscatter Sensitivity:	10^{-7} m^{-1} (450 nm, 550 nm, 700 nm) at 1 min averaging time.
Instrument Operating Temperature Range:	5°C to 35°C
Power requirements:	120 VAC 60 counts

**Figure 1.** The SGP site aerosol sampling manifold, including the new additions to the system.

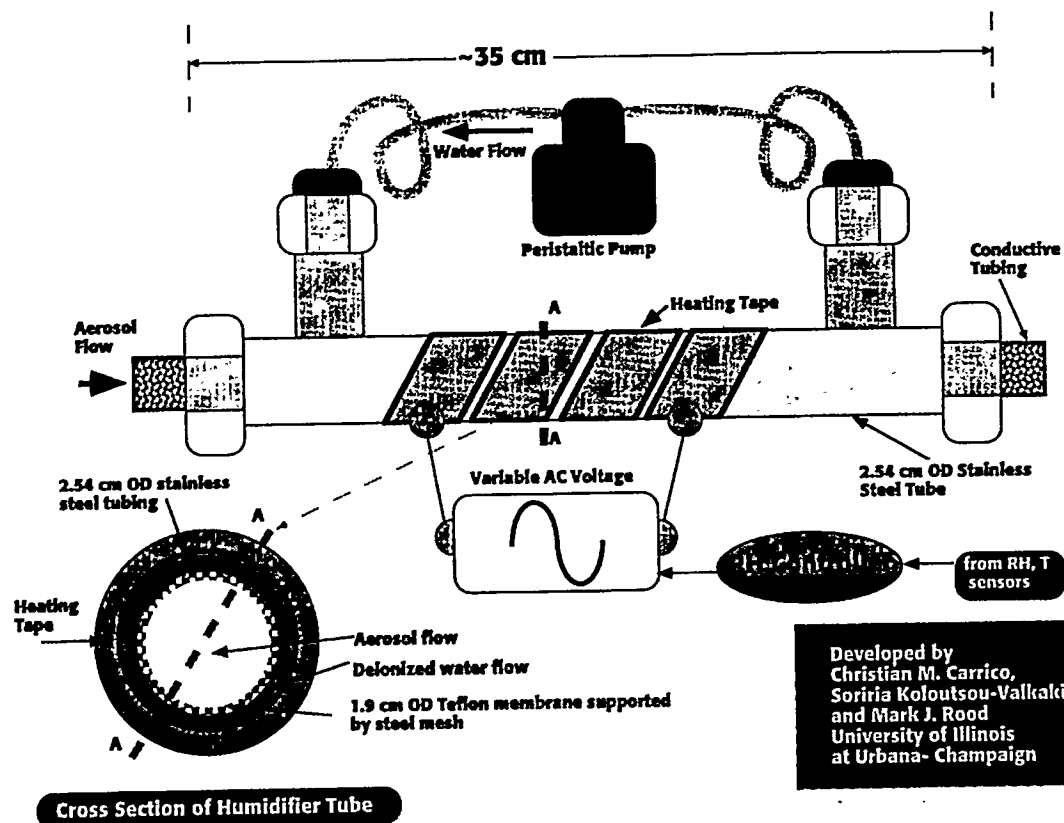


Figure 2. Schematic of the humidifier used to vary the relative humidity of the dried ambient aerosol entering the three-wavelength nephelometer. Based on the research efforts of Dr. Mark Rood of the University of Illinois of Urban, Champaign.

Comparison of Satellite-Derived and Observer-Based Determinations of Cloud Cover Amount at the Southern Great Plains Cloud and Radiation Testbed Site

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Cloud-climate feedback is one of the most important factors in predicting the timing and magnitude of global climate change and its regional effects. Recent satellite measurements indicate that global effects of clouds on solar and infrared radiation are large (Ramanathan et al. 1989). The experimental objective of the Atmospheric Radiation Measurement (ARM) Program is to characterize, empirically, the radiative processes in the earth's atmosphere with improved resolution and accuracy (ARM 1990). Therefore, the effective treatment of cloud formation and cloud properties is crucial for reliable climate prediction.

This study focuses on the analysis of cloud cover data for the ARM Southern Great Plains (SGP) Cloud and Radiation Testbed (CART) site central facility. The data set was obtained from the Advanced Very High Resolution Radiometer (AVHRR) on National Oceanic and Atmospheric Administration (NOAA) Satellites 11 and 12, and cloud observations made by SGP CART site operators. Such an analysis provides a basis for future evaluations with whole-sky cameras and provides a means of assessing the reliability of surface-based observations of cloud cover at the SGP CART site.

Satellite radiometer data have been analyzed previously to detect clouds (Coakley and Bretherton 1982; Arking and Childs 1985). Most techniques use a defined threshold of reflectance to distinguish clear and cloudy skies (Girolamo and Davies 1994). In this study, AVHRR data were used to determine the amount of cloud cover over the SGP CART site central facility, located near Lamont, Oklahoma (36.61°N and 97.49°W).

During the months of March through August, the SGP CART site central facility and the surrounding area are

mostly covered by green vegetation (Gao 1994). Winter wheat is the predominant crop. Because green leaves absorb most of the incident visible radiation at wavelengths of 0.4–0.7 μm and reflect or transmit most of the incident radiation at longer wavelengths, the normalized-difference vegetative index (NDVI) can be used to determine the extent of vegetation cover. The NDVI is defined as

$$\text{NDVI} = (\rho_{\eta} - \rho_r) / (\rho_{\eta} + \rho_r) \quad (1)$$

where ρ_{η} and ρ_r are the canopy reflectances in the near-infrared and red portions of the spectrum, respectively. Green vegetation and other land surfaces usually have a positive NDVI value, while bodies of water and clouds have negative values because water molecules reflect and transmit more strongly in the red wave band than in the near-infrared. Gao (1994) showed that the NDVI value for background vegetation and soil at the SGP CART site is positive from March through August. Therefore, for the time period of this study (April 8 to July 24, 1994), we set the NDVI threshold for cloud detection at zero. A negative NDVI value was assumed to be due to cloud cover, although this may be a conservative choice according to Girolamo and Davies (1994). Because the resolution of the AVHRR at the nadir view position is 1 km^2 , clouds smaller than this scale may not be detected, leading to errors in the estimates of the actual cloud cover (Shenk and Salomonson 1972).

Routine operation of the SGP CART site central facility includes the hourly observation and recording of cloud type and cloud cover, plus the estimation of cloud base height for 1) the total sky by quadrant and 2) the sky overhead within a 30° arc from zenith. The amount of coverage is reported to the nearest tenth (10%), and the cloud base height is

obtained from the *International Cloud Atlas* (1956), which provides a standard cloud base height as a function of cloud type. Table 1 summarizes those cloud types and their standard heights. The radius of the total-sky observational field of view is estimated to be 50 km for high clouds, 40 km for middle clouds, and 15 km for low clouds. The radius of each 30° arc observation is determined individually from

$$\gamma = h \tan 30^\circ \quad (2)$$

where h is the cloud height from Table 1.

For April 8 to July 24, 1994, the cloud cover fraction (in tenths) calculated from NDVI values was compared with SGP CART site cloud observations. Because the times when NOAA 11 and 12 pass over the CART site vary from day to day, the time usually was not precisely on the hour, but the CART site operator cloud observations were. Therefore, comparisons were performed with the SGP CART site cloud observations made closest to the time of the satellite observations. A total of 58 cloud cover comparisons were made for the 30° arc observations and 57 for the total-sky observations.

For the 30° arc observations, 63% of the low-cloud observations agreed within 0.1 with satellite observations, while only 26% of high-cloud observations agreed within 0.1 with satellite observations. Seventy percent of the low-cloud observations agreed and 58% of the high-cloud observations agreed within 0.2 of the satellite observations. For all cloud level observations combined (high, middle, and low), 43% agreed within 0.1 and 64% agreed within 0.2 with satellite observations. These results are shown in Table 2.

Specifically, for the low-cloud 30° arc observations, 9 out of the 27 cases resulted in disagreement between the NDVI and ground observations. In 3 of those cases, the time of the two observations differed by more than 30 min. In 5 of the remaining 6 cases, cloud cover was underestimated by the NDVI. These 5 cases were either type 1 (sparse cumulus) or type 8 (stratocumulus and cumulus at different heights). This result suggests that small cumulus clouds with little vertical extent may not be detected by the NDVI because of the limited resolution of the AVHRR. Furthermore, ground observers may sometimes confuse low clouds of types 1 and 8.

For middle clouds, twelve 30° arc observations were made. In 5 of those cases, the NDVI and ground observations disagreed, with the NDVI underestimating cloud coverage in 2 cases and overestimating cloud coverage in 3 cases. No meaningful conclusion can be drawn from this result.

For the total-sky observations, 37% of the low-cloud observations and 26% of the high-cloud observations agreed within 0.1 with satellite observations. Sixty-seven percent of the low-cloud observations and 63% of the high-cloud observations agreed within 0.2 of the satellite observations. For all cloud level observations combined, 30% agreed within 0.1 and 62% agreed within 0.2 with satellite observations. These results are also shown in Table 2.

Because of the conservative NDVI threshold selected, the satellite data generally underestimated high, thin cloud coverage for both the 30° arc and the whole-sky observations. For example, in 8 cases of high cloud 30° arc, the NDVI detected no clouds, while the ground observer recorded cloud coverage at 0.3-1.0. Of the 8 cases, 7 were accompanied by ground observation of cloud type 9 (thin cirrocumulus). Because cloud type 9 is thin, the green wavelengths are probably transmitted fairly well, allowing the satellite radiometer to detect green vegetation on the ground. Raising the NDVI threshold to a small positive number (i.e., 0.1) might improve agreement with surface-based observations for high, thin clouds.

In summary, the ground observation data, as an hourly cloud coverage mapping, provide a valuable continuous picture of the change in cloud characteristics during the day in the absence of other measurements. Furthermore, as the whole-sky-imager data become available, the ground observation data can be compared with ground truth. Surface-based observations and satellite observations of clouds did not agree as well for the whole-sky view as for the 30° arc for AVHRR satellite-derived NDVI data. The best overall agreement was found for a 30° arc view, or center-weighted view, with low clouds being detected quite well by the NDVI. This study suggests that the hourly cloud observations provided by the SGP CART site operators for the past several years are important in interpreting radiation data fields from the first year of SGP CART site operations when whole-sky imagery and cloud base measurements were not available.

Table 1. Cloud type, reportable height, and description.		
Cloud Type	Cloud Base Height (m)	Cloud Description
Low		
1	1,500	Cumulus, sparse, flattened, ragged
2	1,200	Cumulus, moderate to strong vertical extent
3	1,200	Cumulonimbus, no anvil
4	1,200	Stratocumulus, from spread-out cumulus
5	1,000	Stratocumulus, not from spread-out cumulus
6	500	Stratus, continuous or ragged
7	400	Broken stratus below altostratus or nimbostratus
8	1,200	Cumulus and stratocumulus at different levels
9	1,200	Cumulonimbus, with anvil
Mid		
1	5,000	Altostratus, semi-transparent
2	5,000	Altostratus, hides sun or moon
3	6,000	Altostratus, semi-transparent
4	6,000	Altostratus patches
5	5,500	Altostratus bands or multiple layers
6	5,500	Altostratus from spread-out cumulus
7	5,500	Altostratus, multiple layers, opaque
8	5,000	Altostratus, vertical extent in cells
9	5,000	Chaotic sky, numerous altostratus types, layers
High		
1	9,000	Cirrus filaments, strands, hooks
2	9,000	Cirrus, dense, patches, sheaves
3	10,000	Cirrus, dense, cumulonimbus anvil remains
4	9,000	Cirrus, dense, hooks, filaments
5	9,000	Cirrostratus or cirrus bands, < 45° high
6	9,000	Cirrostratus or cirrus bands, > 45° high
7	9,000	Cirrostratus veil, whole-sky coverage
8	9,000	Cirrostratus veil, not whole-sky coverage
9	10,000	Cirrocumulus alone or with cirrus and cirrostratus

Table 2. Comparison of satellite-derived and ground-based determinations of cloud cover amount for two types of area averaging and three cloud height levels.

Observation	Number of Events	Agree Within 0.1	Agree Within 0.2	Disagree
30° Arc Observations				
High cloud	19	26	58	42
Mid cloud	12	25	58	42
Low cloud	27	63	70	30
Total	58			
Total Sky Observations				
High cloud	19	26	63	37
Mid cloud	11	18	45	55
Low cloud	27	37	67	33
Total	57			

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Numerical Simulations of Altocumulus with a Cloud Resolving Model

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Introduction

Altocumulus and altostratus clouds together cover approximately 22% of the earth's surface (Warren et al. 1986; Warren 1988). They play an important role in the earth's energy budget through their effect on solar and infrared radiation. However, there has been little altocumulus cloud investigation by either modelers or observational programs.

Starr and Cox (SC) (1985a,b) simulated an altostratus case as part of the same study in which they modeled a thin layer of cirrus. Although this calculation was originally described as representing altostratus, it probably better represents altocumulus stratiformis (Houze 1993). In this paper, we simulate altocumulus cloud with a cloud resolving model (CRM). We simply describe the CRM first. We calculate the same middle-level cloud case as SC to compare our results with theirs. We will look at the role of cloud-scale processes in response to large-scale forcing. We will also discuss radiative effects by simulating diurnal and nocturnal cases. Finally, we discuss the utility of a 1D model by comparing 1D simulations and 2D simulations.

Description of Model

The CRM is a 2D (x - z) numerical model. It is based on the anelastic set of equations. It includes third-moment turbulence closure, a turbulent scale condensation scheme, a bulk microphysics parameterization, and an advanced radiation code. The model is more fully described in Krueger (1988) and Xu and Krueger (1991). The radiative transfer parameterization used in the CRM is described in Fu (1991), Fu and Liou (1992), and Krueger et al. (1995). Neither ice-phase nor precipitation is considered in our simulations.

We also use a 1D version of the CRM in which the cloud-scale circulations and the small-scale turbulence are both parameterized. In such a 1D model, a major difficulty is determining the turbulent length scale, $L(z)$. Bougeault

and Andre (1986) proposed an "upward/downward free path" method to determine $L(z)$, which is adopted in our 1D model.

The model domain we used for our numerical simulations is 6.4 km long and 8.9 km high. The horizontal grid interval is 100 m, while the vertical grid interval is 1 km from surface to 5 km, 500 m to 5.5 km, and 100 m from 5.5 km to 8.9 km. The time step is 10 seconds. The reference state is the U.S. Standard Atmosphere (USSA). The temperature and water vapor mixing ratio (q_v) from 5.5 km to 8.9 km is initialized the same as SC. At the levels outside this range, the temperature and q_v are the same as the USSA. Random perturbations are used to initialize the potential temperature θ in the region between 7.2 and 7.5 km. The maximum initial magnitude of the perturbation is 0.1 K. The resulting surface flux is near zero. The ground temperature is fixed at the USSA surface air temperature.

Simulations and Results

A series of numerical simulations, which are listed in Table 1, are performed by using the reference and initial conditions above. In Table 1, w_0 is the large-scale vertical velocity.

Table 1. Numerical simulations.

Case	Dimension	Duration (h)	w_0 (cm/s)	Radiation
Q24	2D	3	2	nocturnal
Q05	2D	6	0	nocturnal
Q26 ^(a)	2D	6	0	nocturnal
Q27	2D	3	-2	nocturnal
Q28 ^(a)	2D	6	0	diurnal
X25	1D	6	0	nocturnal
X26 ^(b)	1D	6	0	nocturnal
(a) Continued from Q05.				
(b) Continued from X25.				

Comparison to the Middle-Level Cloud Case of SC

We simulate a nocturnal case (Q24) which is the same as the case of SC (1985b) with $w_0 = 2$ cm/s. In Figure 1, the time-dependent behavior of the domain-averaged liquid water content ($\overline{\rho_0 l}$) is shown for SC and Q24. The agreement between SC and Q24 is good. In the first 25 minutes, the liquid water content increases quickly, and after 25 minutes, it increases slowly.

In Figure 2a and Figure 2b, profiles of horizontally averaged liquid water content ($\overline{\rho_0 l}$) are shown at various times during SC and Q24, separately. The agreement is also reasonable. In Q24, the cloud top and base height increase faster than those in SC. We compare the turbulent kinetic energy (TKE) in both cases. The TKE in Q24 is stronger. The maximum TKE (at 60 minute) is 0.145 J kg^{-1} in SC and 0.177 in Q24. Considering our third-moment turbulence closure and SC's simple turbulence scheme, we believe that the TKE in Q24 is more realistic. The stronger TKE should be associated with stronger entrainment, so that the faster increase of cloud top height in Q24 is reasonable.

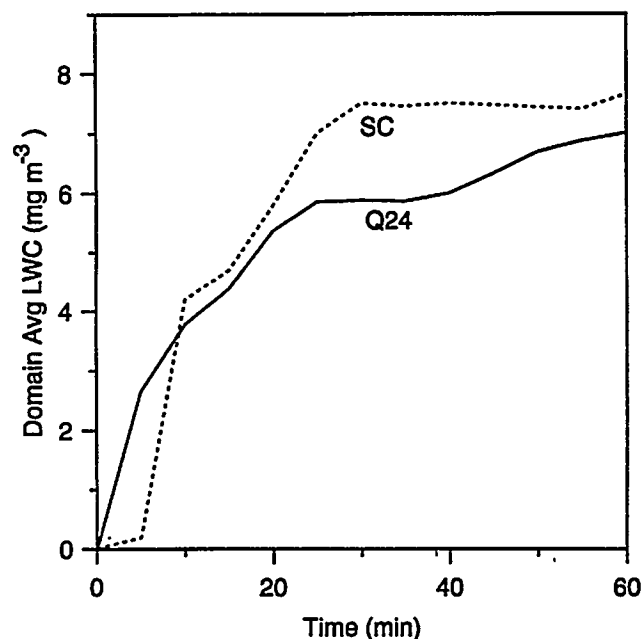


Figure 1. Time-dependent behavior of domain-averaged liquid water content ($\overline{\rho_0 l}$) in SC and Q24.

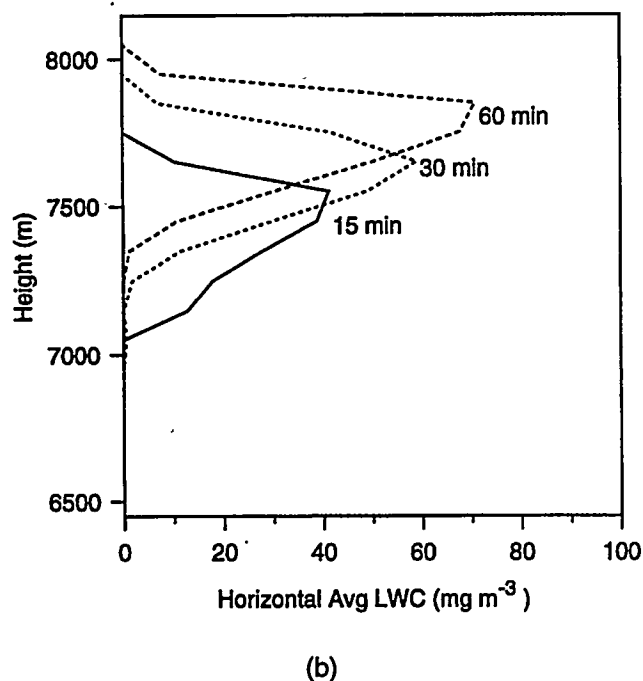
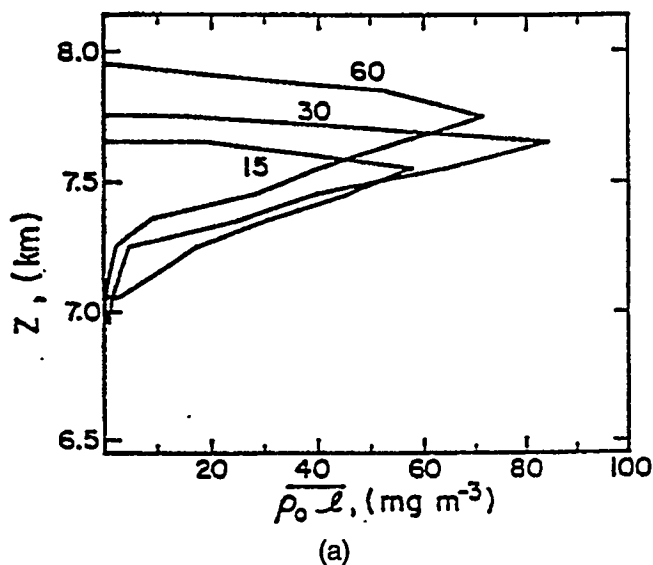


Figure 2. Vertical profile of horizontally averaged liquid water content ($\overline{\rho_0 l}$) at various times for (a) SC (from SC) and (b) Q24.

Effect of Large-Scale Vertical Motion

In this section the effects of large-scale control via w_0 are briefly examined. In Figure 3, the temporal behavior of $\overline{\rho_0 l}$ is shown for three simulations (Q24, Q05, Q27) with various

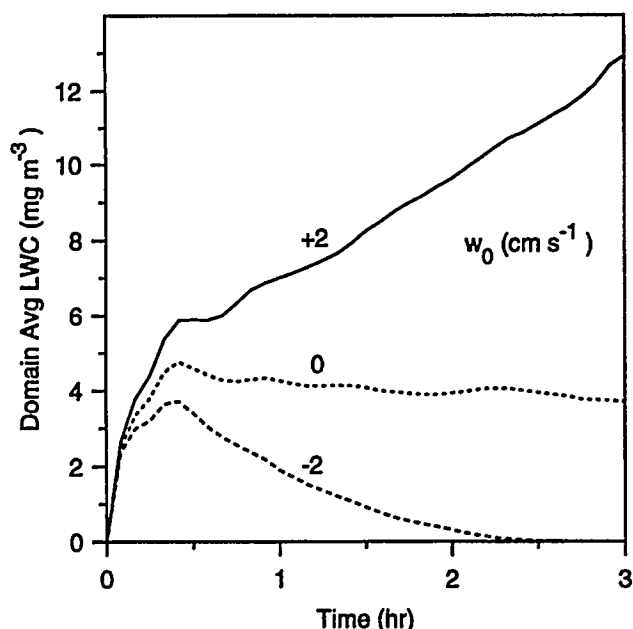


Figure 3. Time-dependent behavior of domain averaged liquid water content ($\bar{\rho}_0 l$) for different large-scale vertical velocities (Q24, Q05, and Q27).

specified values of w_0 . Similar to the analysis of SC (1985b) for cirrus, each simulation may be also partitioned into three stages. During the initial stage, the response to the initial conditions dominates. This is followed by a transition or adjustment period. Then the solutions are dominated by the response to the continuing production of available water vapor and latent heat via w_0 as modulated by cloud-scale processes. With $w_0 = 2 \text{ cm/s}$, $\bar{\rho}_0 l$ increases as the additional water vapor is available. With $w_0 = -2 \text{ cm/s}$, $\bar{\rho}_0 l$ decreases to zero. It is interesting that $\bar{\rho}_0 l$ is almost steady when $w_0 = 0$. It shows that altocumulus may last a long time when there is no large-scale vertical motion.

Effect of Radiation

We set $w_0 = 0$ for the discussion of radiative effects. Two cases are considered: one is a nocturnal case (Q05 + Q26), another is a diurnal case (Q05 + Q28). Q26 and Q28 are restarted from Q05 so that the simulated time period is from 6 to 12 hours. The date is July 15, and the latitude is 30°N . In Figure 4, $\bar{\rho}_0 l$ is shown for each case. In Q28, $\bar{\rho}_0 l$ decreases sharply, which indicates that the cloud will disappear. In Q26, $\bar{\rho}_0 l$ decreases slowly, which indicates that the cloud can last for a longer time. In Figure 5, the total heating rates are shown at various times in Q26 and Q28. The cloud top region is a cooling region, and the

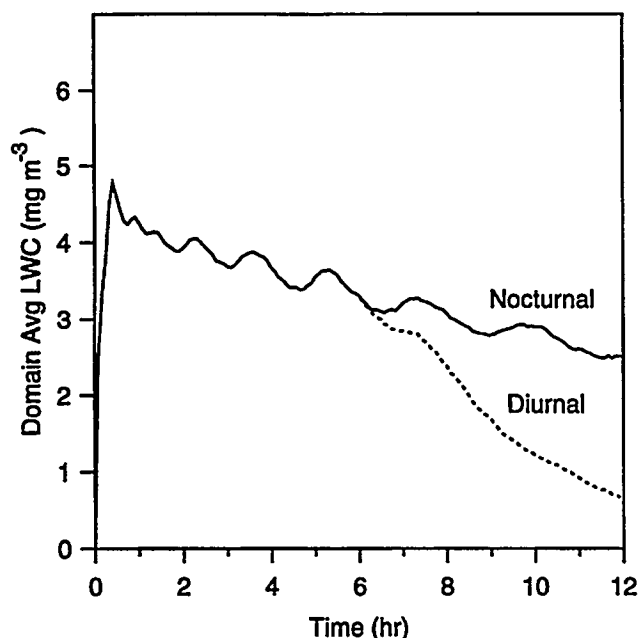


Figure 4. Time-dependent behavior of domain averaged liquid water content ($\bar{\rho}_0 l$) for a nocturnal case (Q05 + Q26) and a diurnal case (Q05 + Q28).

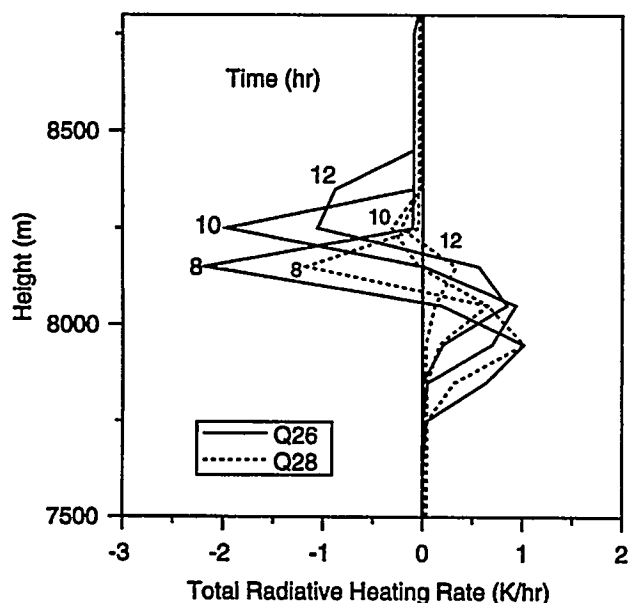


Figure 5. Total radiative heating rate profiles at various times for Q26 and Q28.

cloud base region is a heating region. In both cases the cooling rates decrease with time. However, in Q26 the cooling rates are very large. In Q28, it becomes very small at 10 and 12 hours. This shows that nighttime favors a longer lasting cloud.

Comparison of 1D and 2D Simulations

In Figure 6, the temporal behavior of $\overline{\rho_0 l}$ is shown for a 1D simulation (X25 + X26) and a 2D simulation (Q05 + Q26). In the first hour, the agreement is good. After the first hour, $\overline{\rho_0 l}$ decreases slowly for the 2D case. For the 1D case it decreases quickly from 1 to 2 hours, then decreases slowly. The $\overline{\rho_0 l}$ for the 1D case is maximum about 20% less than that for the 2D case. In Figure 7, $\overline{\rho_0 l}$ is shown at various times in both the 1D and 2D cases. In both cases, the cloud ascends. The cloud ascends more slowly in the 1D case than in the 2D case, and the cloud is thinner in the 1D case. The maximum values of $\overline{\rho_0 l}$ are slightly larger in the 1D case.

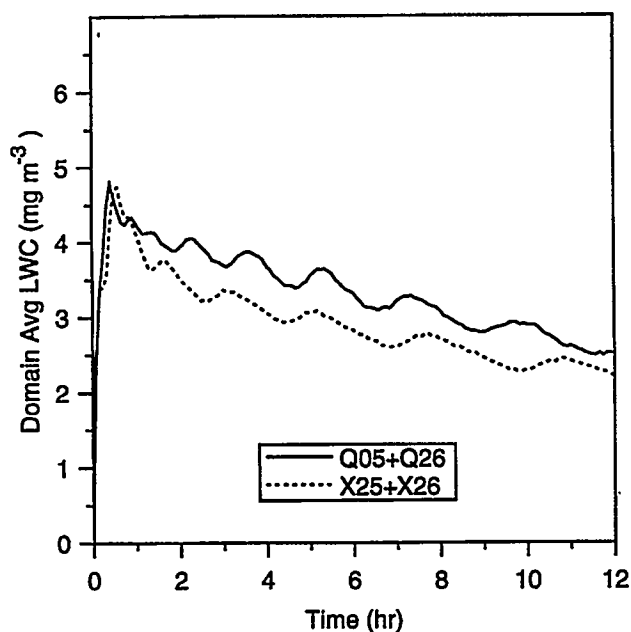


Figure 6. Time-dependent behavior of domain averaged liquid water content ($\overline{\rho_0 l}$) for a 2D simulation (Q05 + Q26) and a 1D simulation (X25 + X26).

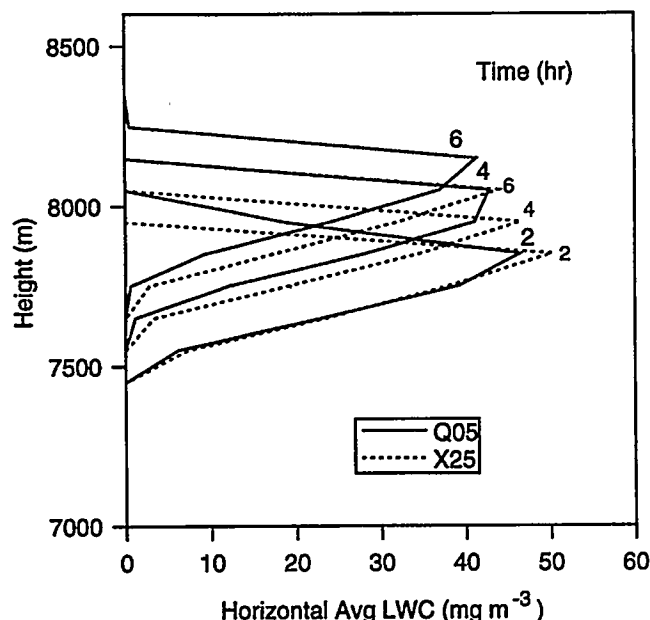


Figure 7. Vertical profiles of horizontally averaged liquid water content ($\overline{\rho_0 l}$) at various times for a 2D simulation (Q05) and a 1D simulation (X25).

Summary

A series of numerical simulations is carried out with a 2D CRM and a 1D turbulence closure model. A good agreement between our 2D simulation and SC's altocumulus cloud case is achieved. After a middle-level cloud forms, weak large-scale vertical motion may allow for a long lifetime of the cloud. In nocturnal conditions, altocumulus can last longer than in diurnal conditions. The comparison of 1D and 2D simulations suggests that 1D simulations can portray the basic characteristics of altocumulus clouds.

Acknowledgments

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A Simple Formula for Determining Globally Clear Skies

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Introduction

Surface measurements to serve as "ground truth" are of primary importance in the development of retrieval algorithms using satellite measurements to predict surface irradiance. The most basic algorithms of this type deal with clear sky (i.e., cloudless) top-to-surface shortwave (SW) transfer, serving as a necessary prerequisite towards treating both clear and cloudy conditions (Cess et al. 1991). Recently, Cess et al. (1995) have used a ratio of surface and top of atmosphere SW cloud forcing to infer the possibility of excess atmospheric absorption (compared with model results) in cloudy atmospheres. The surface component of this ratio relies on inferring the expected clear sky SW irradiance to determine the effects of clouds on the SW energy budget. Solar renewable energy applications make use of clear and cloud fraction climatologies to assess solar radiation resources. All of the above depend to some extent on the identification of globally clear sky conditions and the attendant measurements of downwelling SW irradiance.

Traditionally, cloud fraction information has been provided by surface observer reports and thus is subjective and not continuous. Use has also been made of the downwelling SW record, depending on the magnitude and "smoothness" of the plotted data to estimate clear skies. Because broken cloud and thin cloud effects can equal or actually exceed clear sky downwelling SW irradiance, total downwelling irradiance measurements alone should not be used to determine clear sky.

Other methods used to infer periods of globally clear skies include narrow field-of-view measurements made by instruments such as lidars, ceilometers, and sun photometers. A time series of this data showing no clouds in the field of view is assumed to represent a lack of clouds in the hemisphere, a problematic assumption in many cases. Recently, digital whole-sky imaging systems have been developed that can produce cloud fraction data, though these systems have difficulty in detecting thin cloud layers. In addition, these systems (like lidars and sun photometers) can be expensive to build and operate, requiring on-site personnel and significant data processing to extract the cloud fraction

information. Thus, these types of instruments cannot economically be used in remote areas.

Pyranometers, on the other hand, are relatively inexpensive and robust. The diffuse irradiance can be measured by shading a pyranometer from the direct solar irradiance. The ratio of the diffuse irradiance to the total downwelling SW irradiance is a minimum under clear sky conditions. An empirical fit of the expected diffuse ratio for clear skies for the atmospheric conditions at a given time can be compared with the measured diffuse ratio, thus providing a means to detect periods of globally clear skies. Long and Ackerman (1995) have shown that, for mid-latitude frontal regimes at least, there is a high correlation of pyranometer measurements up to a radius of 90 km. This suggests that identifying periods of clear sky conditions using the global diffuse ratio method can provide an exceptional basis for ground truth comparisons in SW satellite retrieval algorithm development. The diffuse ratio method provides a relatively simple means of identifying globally clear sky conditions in long term data sets such as those archived by the National Renewable Energy Laboratory for the Solar Radiation Research Laboratory in Golden, Colorado, and the Historically Black Colleges and Universities Solar Measurements Network. In addition, methods of using the diffuse ratio to estimate cloud fraction are currently under development. Data sets such as the above can then be processed to retrieve cloud fraction climatologies for solar renewable energy applications.

The Diffuse Ratio

The direct component of solar irradiance is defined as that part of the downwelling SW irradiance reaching the surface which has not been affected by scattering or absorption within the atmosphere. The diffuse component originates from SW radiation that is scattered from the direct solar beam. The total downwelling SW irradiance at the surface is then the sum of the diffuse and the direct components. We define the diffuse ratio as

$$\text{Diffuse ratio} = (\text{diffuse})/(\text{diffuse} + \text{direct}) \quad (1)$$

This ratio has a value ranging from zero to unity.

A base state for the diffuse ratio value at the surface can be defined as that ratio produced from scattering by the homogeneously mixed gases of the atmosphere and absorption by oxygen, ozone, and CO₂ molecules at a given solar zenith angle. We define clear skies to be those which are cloudless. The clear sky diffuse ratio includes the base state, but also becomes a function of the variable aspects of the atmosphere, which are aerosol and water vapor content, and the underlying surface that affect absorption and scattering of SW radiation. To illustrate this dependence, Blunthaler and Ambach (1994) show both a decrease in the direct component and an increase in the diffuse component with increased aerosol optical depth from the Mount Pinatubo eruption. These changes result in an increase in the diffuse ratio from the base state.

The Department of Energy's Atmospheric Radiation Measurement (ARM) Program operates the Southern Great Plains (SGP) Cloud and Radiation Testbed Site located in Lamont, Oklahoma. The ARM Program also conducted a Pilot Radiation Observation Experiment (PROBE) in conjunction with the Tropical Ocean Global Atmosphere/Coupled Ocean Atmosphere Response Experiment (TOGA/COARE) from November 1992 through February 1993 at a site near Kavieng, Papua New Guinea. Available data from these sites include measurements of downwelling SW and diffuse SW irradiance, column water vapor amounts, and (in the case of the SGP site) Whole Sky Imager (WSI) images. Because of the propensity for cloudy conditions experienced during PROBE and lack of verification by WSI images, downwelling clear sky SW irradiances were calculated using a delta two-stream model. Profiles of temperature, pressure, and water vapor from sondes serve as the basic input for the computations, along with climatological ozone profiles and an aerosol size distribution determined using sun photometer measurements. Comparisons between model output and measured values of downwelling SW for clear skies give a standard deviation of less than 15 Wm⁻². The model total and diffuse downwelling SW irradiance were used in the development of the diffuse ratio method and sensitivity studies. For the SGP site, WSI images from the April 1994 intensive observation period were used to determine periods of globally clear skies for comparison with measurements of total and diffuse SW irradiance. Coincident column water vapor amounts were used for model calculations. This site provided a range of column water vapor amounts from roughly 0.5 to 1.5 precipitable cm, while the PROBE water vapor amounts ranged from roughly 3.5 to 6.5 cm.

The observations and model results have been used to derive an empirical fit to estimate the diffuse ratio for clear sky conditions. This fit includes the cosine of the solar zenith angle and column water vapor amount as independent variables in an equation of the form:

$$DR = C_1 + C_2 * (Wvp) + C_3 * \exp[(C_4 * (\cos Z) + C_5 * (Wvp))] \quad (2)$$

where C_n are constants, DR is the clear sky diffuse ratio, Wvp is the column water vapor amount in cm, and CosZ is the cosine of the solar zenith angle. This fitted equation has been tested with encouraging results as shown in the poster presentation given for the ARM Science Team Meeting. However, model sensitivity studies show the need for further development of the method, particularly in the areas of surface albedo and aerosol effects. The cosine of the solar zenith angle is the dominant factor in the clear sky diffuse ratio magnitude and indications are that the method can only be considered useful for zenith angles less than 80°. The sensitivity of the ratio to surface albedo is such that an albedo change of 10% affects the ratio about as much as a change in column water vapor amount of 1 cm. However, with the exception of snowfall, changes in surface albedo generally occur at a much slower rate than column water vapor changes. Thus once the sensitivity to albedo is incorporated it is expected that seasonal values can be used. The effects of aerosol on the ratio have yet to be investigated, primarily due to lack of aerosol optical depth data.

Further Research

With the continuing operation and improvement in data dissemination at the SGP site and the planned installation and long-term operation of an ARM site at Manus, Papua New Guinea, we propose to continue refinement of the diffuse ratio technique for identifying globally clear skies. The Manus site is particularly important in this refinement due to the need for actual clear sky data for high column water vapor amounts to preclude the necessity of using model output. Using upwelling SW measurements available from both sites will aid in the investigation of the seasonal albedo dependence of the ratio. We also propose to use sun photometer or rotating shadow band radiometer data to investigate the effects of aerosols on the method.

A rotating arm mechanism to be used in conjunction with a standard pyranometer is currently under development and testing. This will allow the measurement of both the total and diffuse irradiance by a single instrument, thus removing instrument calibration offset errors in the ratio. This arrangement is envisioned as making possible inexpensive deployment at many sites.

Finally, we propose to use WSI cloud fraction information and time series of both measured and clear sky diffuse ratio data to develop a method of estimating cloud fraction. Along with the deployment of the rotating arm instrument, this will facilitate the retrieval of clear/cloud fraction climatologies from existing and future data sets.

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Diffusion and Scattering in Multifractal Clouds

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Turbulent cascade processes generically give rise to extremely variable multifractals, and there is now abundant empirical evidence that both cloud liquid water density and cloud radiance fields are multifractal over wide ranges of scale. Since general circulation models and NWP's assume that clouds are on the contrary horizontally homogeneous, it is therefore important to study the transport properties (particle, photon) of multifractals and compare the results with the homogeneous plane parallel situation. Since the plane parallel and multifractal radiative properties diverge for increasingly thick clouds, it becomes fundamentally important to characterize the scaling properties and optical thickness of real clouds.

We investigate these radiative properties of multifractal clouds using two different approaches. In the first, we examine the simplest nontrivial transport problem, diffusion, by considering the scaling properties of one dimensional random walks on media with multifractal diffusivities. We show both theoretically and numerically that the anomalous scaling depends on the scaling of the reciprocal spatially averaged multifractal resistance to diffusion. We show that the anomalous scaling will always be subdiffusive; the walkers are effectively trapped in a hierarchy of barriers. In the small-scale multifractal limit, the trapping is dominated by contributions from a specific order of singularity; this leads to a phase transition between anomalous and normal diffusion as the extreme resistivities (barriers) are reduced.

A second approach considers the scattering statistics associated with radiative transport. We develop an (analogue) co-dimension formalism for dealing with this problem. The basic quantities are the size of the medium measured in units of mean free paths (K), the co-dimension function of the media density, and the scattering (analogue) co-dimension function. We obtain simple relations asymptotic in K between these functions and hence obtain the single

scattering statistics. In some special cases, it is possible to renormalize the multifractal so as to obtain an equivalent homogeneous medium. We then estimate anomalous diffusion exponents.

Diffusion

Consider a one-dimensional multifractal optical density field denoted $\rho\lambda(x)$ where $\lambda > 1$ is the ratio of the largest scale of interest to the smallest scale of homogeneity (see Silas 1994 and Lovejoy et al. 1995a for more details). We have

$$\rho_\lambda \sim \lambda^\gamma$$

where the statistical exponents γ are orders of singularity satisfying a well-defined probability distribution at each scale.

$$\Pr(\rho_{\lambda^c} > \lambda^\gamma) \approx \lambda^{-c(\gamma)}$$

where $c(\gamma)$ is the co-dimension function and equality is to within slowly varying prefactors. The statistical moments of the multifractal field are described by the moment scaling function $K(q)$

$$\langle \rho_\lambda^q \rangle = \lambda^{K(q)}$$

where q is the order of the moment and the brackets indicate ensemble averaging.

The coefficient of diffusion $D\lambda(x)$ is taken to be

$$D_\lambda(x) = \frac{1}{\rho_\lambda(x)}$$

that is, $p\lambda$ is a resistance to diffusion; regions of high resistance to diffusion are likely to correspond to rare, dense (impenetrable) regions of the medium. In the diffusion approximation to radiative transfer, $p\lambda(x)$ is proportional to the (multifractal) optical density (Figures 1 and 2).

We now use a one-dimensional long-time

$$\rho_{\text{eff}} = \frac{1}{D_{\text{eff}}} = \frac{t}{x^2} \approx \frac{1}{N} \sum_{i=1}^N \frac{1}{D_i} = \frac{1}{N} \sum_{i=1}^N \rho_i$$

where N is the number of distinct sites visited by the random walker and the ρ_i are the resistances associated with those sites, t is the time taken for the walk, and (equation) is the variance of walks on a single realization of the (multifractal) process (overbars indicate means over walks, angle brackets indicate means over ensembles of multifractal realizations of $\rho[x]$). The simple interpretation is that the random walker experiences an effective resistance $\rho_{\text{eff}} = (D_{\text{eff}})^{-1}$ equal to the mean resistance of the sites it has visited.

We now define the exponent d_w by

$$\overline{\langle x^2 \rangle} \sim t^{\frac{2}{d_w}}$$

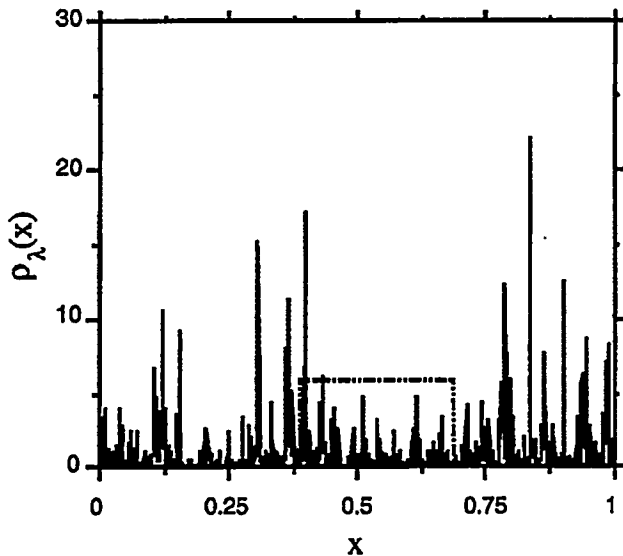


Figure 1. Multifractal field with $CI = 0.2$ and $\lambda = 1024$.

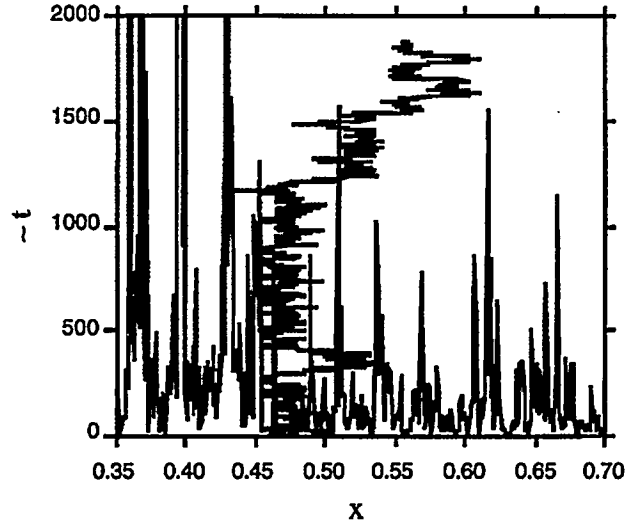


Figure 2. A random walk performed on the inset of Figure 1. A superposition of the trail of the walk (121,080 steps taken) upon the region of the field explored by the walker is pictured here. The walker is "trapped" between large values of the field, hence a slow-down of the diffusion process.

and estimate $\langle D\lambda \rangle$ from the average of the λ intervals:

$$\langle D_\lambda \rangle = \langle \rho_\lambda^{-1} \rangle = \lambda^{K(-1)}$$

where we have assumed that $K(-1)$ is finite (hence will not always apply to the universal multifractals with $\alpha < 2$). We finally obtain

$$d_w = 2 + K(-1)$$

Hence, since $K(-1) > 0$, $d_w > 2$, we will have subdiffusive behavior; the particles are trapped in a hierarchy of barriers. In the limit $\Lambda \rightarrow \infty$ the diffusive behavior is therefore totally dominated by structures with resistivity singularity $\gamma-1 = K'(-1)$ distributed over a fractal set with co-dimension $c(\gamma-1)$. The higher order singularities are too rare to affect the transport, and the lower order singularities are too weak to significantly trap the particles. This critical singularity is associated with a phase transition: if the resistivity field is replaced by a thresholded field with all values exceeding a fixed T reset to the value T , then, in the limit $\Lambda \rightarrow \infty$, there will be a transition from anomalous diffusion (with the above exponent) to normal diffusion when T is reduced below the critical value $\Lambda\gamma-1$ (Figure 3).

Particle Scatter/Radiative Transfer

Here we outline recent results which provide the basis for systematic study of radiative transport in multifractal media (see Brösamlen 1994 and Lovejoy et al. 1995b). Specifically, we indicate how formulae analogous to the multifractal optical density field arise for radiative properties. Consider the following definitions:

k = extinction coefficient [$\text{m}^2 \text{Kg}^{-1}$]

$\langle \rho \rangle$ = mean cloud density [kgm^{-3}]

l = random photon path distance [m]

L = size of cloud [m]

l_m = mean free path (m.f.p.) of a photon in the equivalent homogeneous cloud = $(k\langle \rho \rangle)^{-1}$ [m]

$\lambda = L/l = \text{scale ratio} \leq \Lambda$ (Λ = maximum cascade resolution)

$x = l/L = \lambda^{-1}$ random photon nondimensional distance, (fraction of cloud)

$\tau_p = 1/l_m$ = random photon distance, (no. of homogeneous cloud m.f.p.'s = kx)

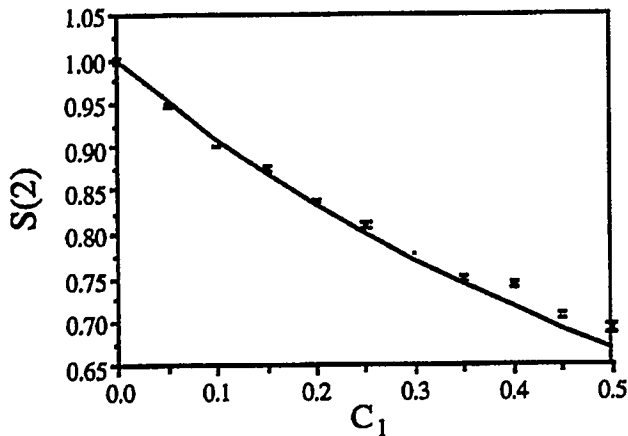


Figure 3. Dependence of the scaling exponent $S(2)$ of the second order moment of x ($\langle x^2 \rangle \sim t^{S(2)}$) on C_1 . The solid line is the theory $(1+C_1)^{-1}$; the data points were obtained from simulations.

$K = L/l_m$ = extinction parameter = no. of m.f.p.'s across cloud = mean optical depth = extinction coefficient in units such that $L = 1$, $\langle \rho \rangle = 1$.

Figure 4 shows an illustration in one dimension (c.f. Figure 2 for diffusion). The optical distance over a physical distance 1 is thus

$$\tau(l) = \int_1 k \rho_A(z) dz = k \rho_1 l$$

where we have written ρ_1 for the average density at resolution 1. We obtain

$$\tau(\gamma, \lambda) = \frac{\rho_\lambda}{\langle \rho \rangle} \lambda^{-1} \frac{L}{l_m} = \lambda^{\gamma-1} K$$

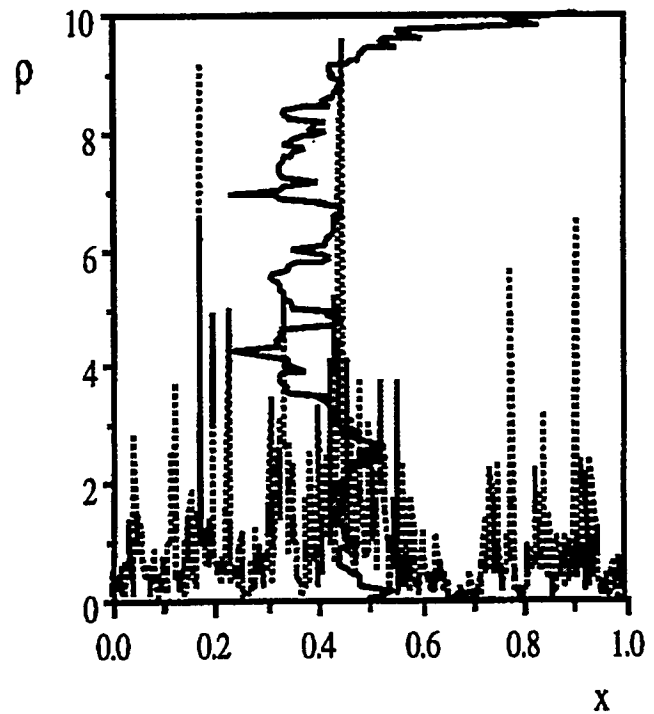


Figure 4. Photon "random walk" in a multifractal cloud ($C_1 = 0.1$; $A = 512$) with extinction coefficient $K = 32$. The y-axis represents the cloud density ρ and the number of scatters of the photon (1 unit corresponds to 20 scatters). The x-axis represents the position in the 1-d cloud. With increasing extinction coefficient K , the mean free path length of the photon decreases.

which is the optical thickness over a distance 1 through a singularity of order γ . The direct (unscattered) transmission T across this distance is thus

$$T(l) = e^{-\tau(l)}$$

Since the transmittance is the probability distribution for photon path lengths, we can average over the singularities and obtain

$$\Pr(1' > l) = \langle T(l) \rangle = \langle e^{-\tau(l)} \rangle$$

Take τ_p as the dimensionless photon path and write it as a scaling function with an order of singularity τ_p defined as follows

$$\tau_p = \kappa^{\gamma_p} = \frac{1}{l_m} = \kappa x = \kappa \lambda^{-1}$$

or

$$\lambda = \kappa^{(1-\gamma_p)}$$

since $\lambda > 1$, $K > 1$, we have $\gamma_p \leq 1$; also $\Lambda > K^{1-\gamma_p}$.

We can now obtain a multifractal scattering formalism in which the extinction coefficient K takes the place of the scaling parameter λ . Instead of the co-dimension function $c(\gamma)$ of the singularities of the cloud density γ , we rather talk about an analogue co-dimension function $c_p(\gamma_p)$ which describes how the single photon path distance singularity γ_p varies with the extinction coefficient:

$$\Pr(\tau_p \geq \kappa^{\gamma_p}) = \Pr(1' > l) \approx \kappa^{-c_p(\gamma_p)}$$

$$\langle \tau_p^{q_p} \rangle = f_p(q_p) \kappa^{K_p(q_p)}$$

$f_p(q_p)$ is the prefactor

$$K_p(q_p) = K(q_p) = q_p + q(q_p)$$

$$f_p(q_p) = \frac{q_p \Gamma(-q(q_p))}{1 - K'(q(q_p))} (\log \kappa)^{-1}$$

with q , q_p related as above.

Renormalization

We can now relate the transmission statistics of lognormal multifractal clouds to those of homogeneous clouds. At first sight this seems to be a difficult task since, in the thick limit (K large), both types of clouds will result in a completely different behavior of the radiative transfer properties. However, we find that the photon statistics of a multifractal cloud can be approximated by the photon statistics of a "renormalized" homogeneous cloud in a certain range of photon singularities. This cloud has the direct transmission given by $T(x) = e^{-\kappa_{\text{eff}} x}$ where κ_{eff} is the equivalent "effective" extinction coefficient. For a log-normal multifractal cloud, we obtain

$$\kappa_{\text{eff}} = \kappa^{\frac{1}{1+c_1}}$$

We now test that this idea works for multiple scattering using the numerical results of Davis et al. (1991) (with $C1 = 0.5$; Figure 5). These simulations were made using two dimensional discrete lognormal cascades with scale ratio factor 2 per step, total range of scales 2^{10} . Cyclic boundary conditions were used in the horizontal and photons were vertically incident. Using the standard thick cloud plane

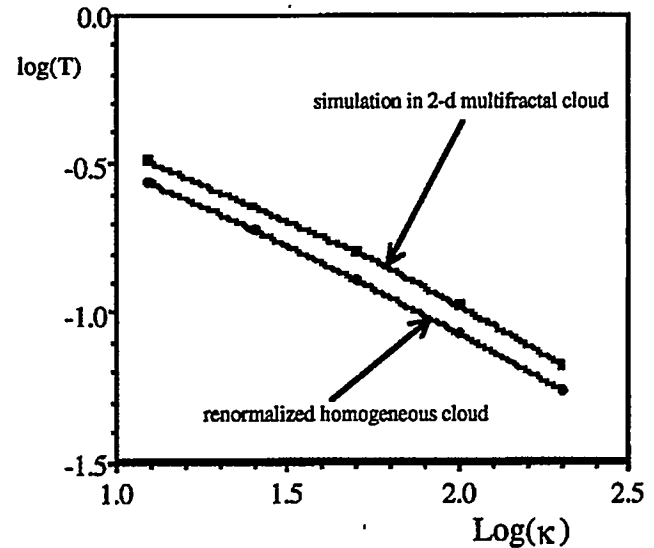


Figure 5. Result on total transmission after multiple scattering through 2-d multifractal cloud ($C1 = 0.5$) compared with the thick cloud limit of the transmission through a homogeneous cloud with renormalized extinction coefficient K_{eff} .

parallel result with the effective extinction coefficient in place of the true optical thickness $\tau = K_{\text{eff}} \langle p \rangle = K_{\text{eff}}$, we obtain

$$\langle T \rangle = \frac{1}{1 + \frac{1}{2} K^{\frac{1}{1+C_1}}} \approx 2K^{-\frac{1}{1+C_1}}$$

Figure 5 shows the result of superposing this function on the numerics, which are nearly power law even for K as low as 12.5. The total transmittances through the renormalized homogeneous cloud show for all values of K only less than 20% difference from the total transmittances through the multifractal cloud.

Discussion

The surprisingly accurate prediction of thick cloud numerics can perhaps best be understood by considering the relation between radiative transfer and diffusion on multifractals. In general, there will be two significant limits: the large Λ (wide cascade range) and large K (thick cloud) limits. Clearly, for fixed and finite Λ , if the cloud is made thick enough ($K \gg \Lambda$), the mean free path will be much smaller than a single resolution element, and the photons will diffuse through each homogeneous region of size Λ^{-1} . The overall result will be photons diffusing through the multifractal cloud. In actual fact, diffusion can still occur under somewhat less stringent conditions when K is large, the main requirement being that weak density regions become so rare that direct photon transmittance across a large fraction of

the cloud is statistically negligible. The multifractals with $\alpha < 2$ have precisely the property that they are dominated by weak events (negative singularities) called "Levy holes." It is a priori possible that even with large K , if Λ is sufficiently large (the order of the limits $\Lambda \rightarrow \infty$ and $K \rightarrow \infty$ is important, i.e., with K fixed, but with $\Lambda \rightarrow \infty$), they will have large regions dominated by the holes and hence lead to nondiffusive transfer.

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Examination of Objective Analysis Precision Using Wind Profiler and Radiosonde Network Data

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Introduction

One of the principal research strategies that has emerged from the science team of the Atmospheric Radiation Measurement (ARM) Program is the use of a single column model (SCM). The basic assumption behind the SCM approach is that a cloud and radiation parameterization embedded in a general circulation model can be effectively tested and improved by extracting that column parameterization from the general circulation model and then driving this single column at the lateral boundaries of the column with diagnosed large-scale atmospheric forcing. A second and related assumption is that the large-scale atmospheric state, and hence the associated forcing, can be characterized directly from observations. One of the primary reasons that the Southern Great Plains (SGP) site is located in Lamont, Oklahoma, is because Lamont is at the approximate center of the NOM Wind Profiler Demonstration Array (WPDA). The assumption was that hourly average wind profiles provided by the 7 wind profilers (one at Lamont and six surrounding it in a hexagon) coupled with radiosonde launches every three hours at 5 sites (Lamont plus four of the six profiler locations forming the hexagon) would be sufficient to characterize accurately the large-scale forcing at the site and thereby provide the required forcing for the SCM. The goal of this study was to examine these three assumptions.

Evaluation of Objective Analysis Techniques

The wind profilers and radiosondes provide vertical profiles of wind speed and direction and thermodynamic variables. A diagnostic determination of the large-scale forcing requires specification of the horizontal gradients of these wind, temperature, and moisture fields. An accepted technique for deriving the horizontal gradients is a least-squares fitting of polynomial surfaces to the spatially distributed profile observations. In this study, five objective analysis techniques were used to derive the horizontal gradients: 1) exact fitting of a plane surface to 3 observations (L3),

2) least-squares fitting of a plane surface to 4 observations (L4), 3) least-squares fitting of a plane surface to 5 observations (L5), 4) least-squares fitting of a plane surface to 6 observations (L6), and 5) least-squares fitting of a quadratic surface to 7 observations (Q7). Once the surface-fitting is completed, gradients are determined analytically from the fitted functions. Error in the diagnosed gradients can arise from two principal sources. The first is uncertainty in the input observations arising from instrument noise and small scale variability. The second is misspecification of the fitted surface, i.e., the chosen functional form of the fitted surface does not match that of the actual atmospheric field.

Influence of Observational Uncertainty

Observational errors propagate into the diagnostic results through the fitting process. The resulting uncertainty in the objective analysis results can be estimated if the errors are assumed to be spatially uncorrelated and of known magnitude. We have carried out this estimation using a prescribed rms uncertainty of 1.3 m/sec in wind speed and 0.5 K in temperature and all reasonable combinations of existing profile observation locations at the SGP site. The average magnitudes of the uncertainty in horizontal divergence (HDiv), horizontal advective acceleration (HAcc), and temperature advection (TAdv) for all 5 objective analysis techniques are given in the Table 1. The table values show that the objective analysis uncertainty due to observational error is nearly the same magnitude as the expected synoptic-scale signal.

Table 1. Magnitudes of objective analysis errors.

	L3	L4	L5	L6	Q7
HDiv (10-5/sec)	0.832	0.819	0.741	0.663	2.03
HAcc (10-4/sec)	0.845	0.845	0.767	0.663	1.95
TAdv (10-4/sec)	0.325	0.325	0.295		

Influence of Functional Misspecification

This error arises primarily from using too simple a surface to represent a particular meteorological quantity. Thus, it depends on the characteristics of the field of the particular quantity and becomes important when that field contains variability at higher order than is in the objective analysis model. The logical approach would seem to be to fit the field with the highest order surface possible. However, increasing the complexity of the approximating polynomial guarantees increasing errors due to observational uncertainty (see the Q7 values in Table 1).

We have carried out a case study of this error using the 6 profilers in a hexagon around Lamont. The profilers are divided into two separate, overlapping triangles. Thus, there are no common profilers between the two triangles and both triangles are centered on Lamont, at which location we would expect the analysis results from the two triangles to be nearly identical. The case study period was November 26-27, 1991, using data acquired during the FIRE II field program. The results from the two triangles are compared in Figure 1 to a third curve computed from a quadratic fit to all 6 profilers. The range of magnitudes of the difference between the two triangle curves is as large as the range of variability expected over a full synoptic wave.

Conclusions

Our research leads us to conclude that objectively analyzed data from the profiler array and ARM sonde network are generally not suitable for initialization of SCMs. Significant errors are introduced by a combination of uncertainty in the observations and higher order variability in the actual fields not accounted for in the least squares fits. We see three possible steps that can be taken to improve the situation. The first is to carefully screen meteorological situations in order to identify the cases that are suitable for SCM forcing. This will limit the breadth of the data base and the data available for parameterization improvement. The second is to use more sophisticated objective analysis techniques. While this may lead to a reduction in errors, the techniques will be time-consuming and the error uncertainty difficult to quantify.

The third step is to use operational assimilation models such as the Mesoscale Analysis and Prediction System (MAPS) to provide initialization data. These models have the virtue of providing standardized, mathematically balanced forcing fields. If the SCM experiments are going to follow this path, we question the utility of the ARM radiosonde program. We recommend that the most efficient use of the resources expended on the boundary facility radiosondes is to make sure that the observations are immediately available for assimilation into whatever model (such as the MAPS) is used.

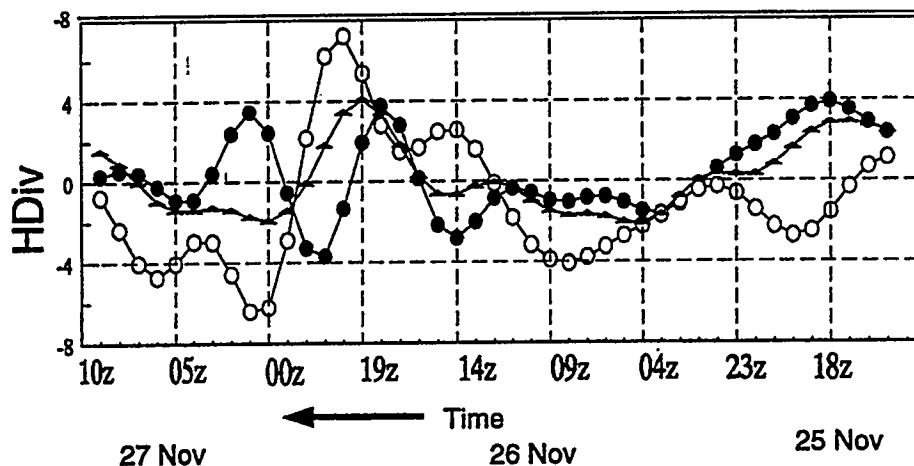


Figure 1. Time series of horizontal divergence at 9-km altitude for case study period. Circles denote results from profiler triangles. Hatched line is computed from a quadratic fit to 6 profilers.

A Comparison of Radiometric Fluxes Influenced by Parameterized Cirrus Clouds with Observed Fluxes at the Southern Great Plains (SGP) Cloud and Radiation Testbed (CART) Site

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As the data stream from the ARM Southern Great Plains (SGP) site matures, the value of this resource is becoming more readily apparent. For the first time, research on the coupling of clouds and radiation is not limited to data from short duration field campaigns. While data from these campaigns are necessary, the full spectrum of naturally occurring variability must be considered in the development and verification of parameterizations for global models.

To exploit the SGP data stream fully, we have developed an operational data processing and analysis methodology that allows us to examine continuously the influence of clouds on the radiation field and to test new and existing cloud and radiation parameterizations. Our approach is sketched out in Figure 1. In order to place the observations into a large-scale context and obtain hourly descriptions of the profiles of temperature, humidity, and vertical motions over the SGP site, we acquire the Mesoscale and Analysis Prediction

System (MAPS) output on an hourly basis from the National Meteorological Center. The MAPS profiles serve as input to a radiative transfer model, which is used to determine clear sky fluxes of solar and infrared radiation. These computed fluxes are compared to observed fluxes. By then examining the micropulse lidar or Belfort ceilometer output, the presence of a cloud is ascertained. If a cloud is detected, the difference between the observed fluxes and computed clear sky fluxes (the surface cloud forcing) is linked to the observed cloud. Initial results from this method demonstrate that, as expected, infrared cloud forcing is well described by the cloud base temperature. The solar cloud forcing, however, shows a large degree of variability among cloud events that is likely attributable to variations in cloud microphysics and water path. It is this variability that must be understood and captured if the parameterization of clouds in climate models is to be improved.

Some results of our approach to validating cloud parameterizations are shown in Figure 2. In this case, the MAPS data were used to force the cirrus cloud parameterization of Heymsfield and Donner (HD) (1990). The upper panel shows the parameterized ice water contents for a cloud event that occurred on December 11, 1994. The predicted cloud base compares well with observations from the multipulse lidar (MPL). The ice water contents derived from HD are incorporated into the Ebert and Curry (1992) radiative parameterization to determine cloud optical properties. These properties serve as input to a radiative transfer model to estimate the cloudy sky radiative fluxes. For this case, the computed surface radiative fluxes are in good agreement with the observed fluxes (lower panel). Using this ongoing analysis in conjunction with improved cloud diagnostic algorithms (see Long et al., this volume) and detailed case studies (see Albrecht et al., this volume), we intend to compute seasonal and annual statistics of cloud forcing and compare them with the results derived from cloud parameterizations.

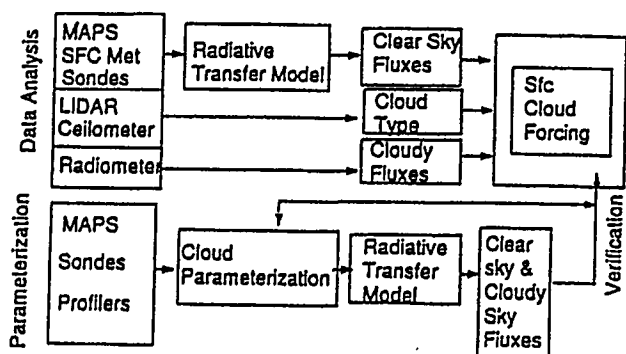


Figure 1. Schematic diagram of an operational methodology for cloud parameterization testing. The upper portion depicts the data analysis procedure, while the lower portion depicts parameterization forcing and evaluation.

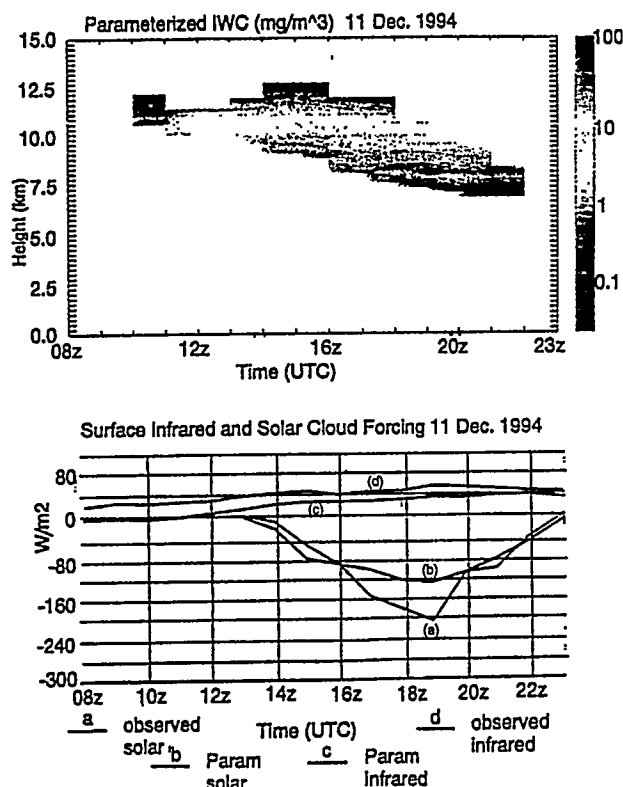


Figure 2. Example of the application of the operational methodology for a cirrus case on December 11, 1994. The upper panel shows ice water content from the Heymsfield-Donner parameterization as a function of time. The lower panel shows the observed (a) and computed (b) downward solar fluxes and the observed (c) and computed (d) downward infrared fluxes as a function of time. The ice water path for the upper panel is used as an input for the computed fluxes.

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A One-Year Cloud Climatology Using Data from the Southern Great Plains (SGP) Site Micropulse Lidar

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Data Acquisition and Analysis

The micropulse lidar (MPL) has been operational at the Southern Great Plains site of the Atmospheric Radiation Measurement Program for the past 15 months. The compact MPL is unique among research lidar systems in that it is eyesafe and operates continuously, except during precipitation. Unlike commercial laser ceilometers, the MPL is capable of detecting cloud base throughout the entire depth of the troposphere. The MPL data set is an unprecedented time series of cloud base heights. It is a vital resource for understanding the frequency of cloud occurrence and the impact of clouds on the surface radiation budget, as well as for large-scale model validation and satellite retrieval verification.

The raw lidar data are processed for cloud base height at a temporal frequency of one minute and a vertical resolution of 270 m. The resultant time series of cloud base is used to generate histograms as a function of month and time of day. To create composite diurnal frequency plots, the cloud base data are sorted into three-hour bins with a vertical resolution of 2 km. For plots of the annual cycle, the data are sorted into monthly bins with the same vertical resolution. The counts are then converted to a frequency by dividing each set of vertical bins by the total number of counts in that vertical column.

Sample Results

Two plots are presented in Figure 1 to illustrate the type of information available from the MPL. The upper panel in the figure is the annual composite of the diurnal cycle. (The horizontal axis is UTC, which is approximately 6 hours ahead of local sun time.) The morning peak in boundary layer cloud occurrence is clearly evident as the most distinctive feature. Little diurnal variability in low cloud base

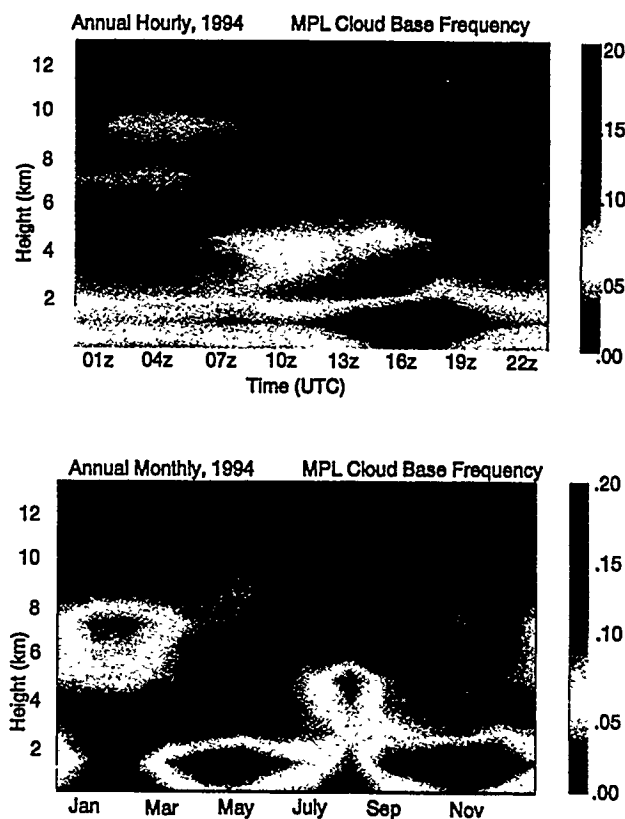


Figure 1. Micropulse lidar cloud base frequency.

height is observed, although the relatively coarse 2-km binning may obscure some variability. Middle tropospheric cloud base peaks during the late night and early morning hours. Cirrus bases tend to be observed throughout the day, but peak during the night.

The lower panel in Figure 1 depicts the annual cycle of cloud base occurrence. Well-defined maxima in boundary layer cloud base occur during the spring and autumn

months. Higher frequencies of occurrence of middle level clouds in late summer and cirrus clouds in winter may be due in part to the absence of low clouds. Note that since data for March were missing, data from February and April were used to fill the gap for display purposes.

Several general observations should be made about the two plots. First, since only the lowest cloud base can be detected by the MPL, frequencies of occurrence for middle and upper

tropospheric clouds are certainly underestimates of their actual values. Second, because of solar background noise, thin cirrus are often undetected during daylight hours. Enhanced sensitivity may also account for the apparently higher cirrus bases detected at night (upper panel in Figure 1). Lastly, although these cloud occurrence statistics are an underestimate of total cloud occurrence, they do represent an accurate accounting of the clouds that are important to the surface radiation budget.

Comparisons of Cloud Ice Mass Content Retrieved from the Radar-Infrared Radiometer Method with Aircraft Data During the Second International Satellite Cloud Climatology Project Regional Experiment (FIRE-II)

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Introduction

Comparisons of remotely sensed meteorological parameters with in situ direct measurements always present a challenge. Matching sampling volumes is one of the main problems for such comparisons. Aircraft usually collect data when flying along a horizontal leg at a speed of about 100 m/sec (or even greater). The usual sampling time of 5 seconds provides an average horizontal resolution of the order of 500 m. Estimations of vertical profiles of cloud microphysical parameters from aircraft measurements are hampered by sampling a cloud at various altitudes at different times.

Angular resolutions of the National Oceanic and Atmospheric Administration (NOAA) Environmental Technology Laboratory (ETL) K_a -band radar and precision radiation thermometer (PRT)-5 infrared (IR) radiometer used to obtain remotely sensed cloud microphysical parameters are about 0.5° and 2° . This corresponds to a horizontal resolution of about 60 m and 260 m, respectively, at a typical cirrus altitude of 7.5 km. Radar data averaging was performed in this analysis to overcome this difference. The vertical resolution of the remotely sensed profiles of cloud microphysical properties was determined by the 37.5-m transmitted pulse.

The accuracy of aircraft horizontal and vertical coordinates relative to the location of the ground-based instruments was on the order of dozens of meters. This also was contributing to the uncertainty of matching resolution volumes of aircraft and remote instrument sampling. This uncertainty has to be

taken into account when making any comparisons between cloud microphysical parameters obtained remotely and in situ from aircraft.

November 26, 1991, was one of the priority cirrus cases during the Second International Satellite Cloud Climatology Project Regional Experiment (FIRE-II) (Uttal et al. 1993). Persistent cirrus cloud was seen over a time period of more than 3 hours. Before 1830 UTC, the cloud was rather tenuous and its IR brightness temperature was rather close to the clear atmosphere background. After 1830 UTC, both the radar and IR radiometer began sensing the cloud more reliably. After about 2130 UTC, the cloud thickened and a substantial amount of liquid water appeared inside it. During these three hours, the National Center for Atmospheric Research's King Air research aircraft equipped with particle measuring instrumentation was flying in the vicinity of the experimental hub where the ETL radar and radiometer were located. This provided an ample opportunity for comparisons of aircraft in situ and ground-based remotely measured data.

Ice Mass Content Data

Remote Sensing Method Data

The remote sensing technique which was used to obtain vertical profiles of cloud ice mass content requires measurements of radar reflectivity and Doppler velocity and IR brightness temperatures in the atmospheric "window" of 10–11.4 μm . Doppler velocities are used to estimate particle

terminal fall speeds. Brightness temperature measurements are converted to cloud optical thicknesses after accounting for the intervening atmosphere. Both ground-based instruments are pointed vertically. Theoretical details of this method have been given by Matrosov et al. (1992, 1994). When making retrievals here, it was assumed that the effective density ρ of cirrus particles diminishes with particle maximum dimension L (mm):

$$\rho \text{ (g/cm}^3\text{)} = 0.1/L$$

This equation is in general accord with results of direct microphysical measurements which showed that mass of typical cirrus particles ($L > 100 \mu\text{m}$) increases approximately as L^2 rather than L^3 . It was also assumed in retrievals that particle shapes can be described by prolate spheroids with a major-to-minor aspect ratio of 2.5. Such aspect ratio is consistent with particle 2D images.

The radar-radiometer remote sensing method (Matrosov et al. 1994) uses another a priori parameter in the retrieval algorithm (i.e., the exponent in the particle terminal fall speed-size relationship). The data presented here were obtained assuming this exponent to be 1.3. However, some newer results (Mitchell 1995) show that a lower value of about 1.1 could be more appropriate for cirrus particles.

Figure 1 shows the scatter plot of ice mass content (IMC) values retrieved by the radar-radiometer method and

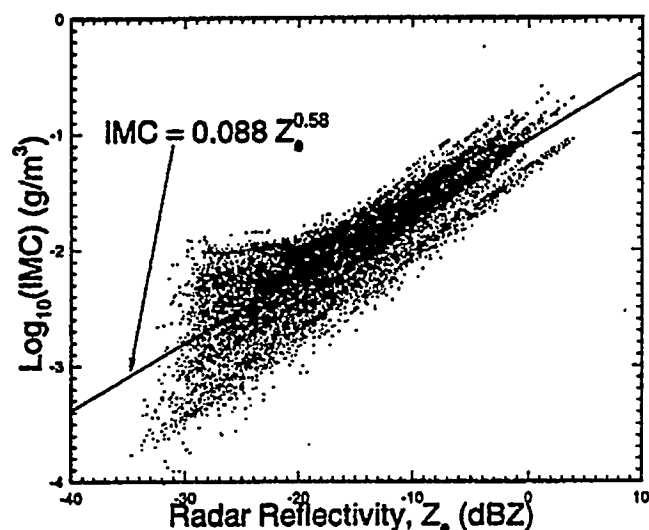


Figure 1. Radar reflectivity-IMC scatter plot for the 26-NOV-91 cirrus case. Remotely sensed data.

measured radar reflectivities. The presented data show a significant variability. Minimum and maximum retrieved values of IMC were about 0.0001 and 0.35 g/m^3 respectively. The minimum measured value of equivalent reflectivity was about -33 dBZ, which is close to the sensitivity threshold of the NOAA ETL K_a-band radar at cirrus cloud altitudes. The solid line shows the best power law regression fit through nearly 16,000 samples. This best fit was obtained with all the samples having the same statistical weight. The reflectivity values in the best fit equations are in mm^6/m^3 .

In Situ Aircraft Data

Figure 2 shows the IMC-radar reflectivity scatter plot as measured by the particle measuring system (PMS) aboard the aircraft.

2D-C probe samples were used to infer IMC values and calculate radar reflectivity. The procedures of obtaining particle microphysical information from 2D probes are described by Heymsfield et al. (1990). The in situ data sample volume contained about 1800 points and corresponded in time to the remotely measured data shown in Figure 1 (1830-2130 UTC). The information for each sample point is a 5-second average.

Comparisons of Figures 1 and 2 reveal a general agreement between remotely sensed and in situ data. The power law best fit lines are quite close. However, in situ data are mostly concentrated between 0.01 and 0.1 g/m^3 , and

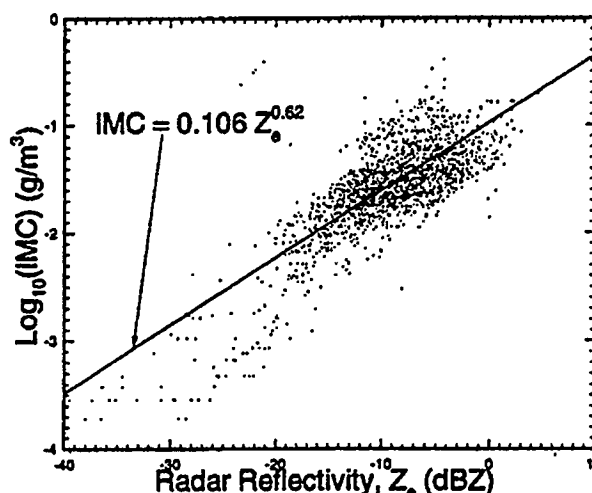


Figure 2. Radar reflectivity-IMC scatter plot for the 26-NOV-91 cirrus case. Aircraft in situ data.

remotely sensed data are distributed more uniformly between 0.001 and 0.1 g/m³. This should not be considered as a significant discrepancy because, as aircraft coordinates show, most in situ samples were taken dozens of kilometers away from the sampling volumes of the ground-based instruments. This most likely means that the aircraft was sampling cloud volumes with relatively high values of IMC and, hence, higher reflectivities.

Comparisons of Remote and In Situ Data

Most meaningful comparisons were the ones made when the aircraft was passing directly over the ground-based instruments. Dashed lines in Figure 3 show projections of aircraft tracks on the horizontal plane during the time when remote sensing data were collected (1830-2130 UTC). Position (0,0) marks the hub where the radar was located.

Figure 3 shows that the aircraft passed over the (0,0) point just once. However, because of uncertainty in determining aircraft coordinates and exact ground-based instrument location relative to the (0,0) point and also because of limited interest of comparing just one point, the area of direct comparisons should include passes in proximity of the (0,0) point.

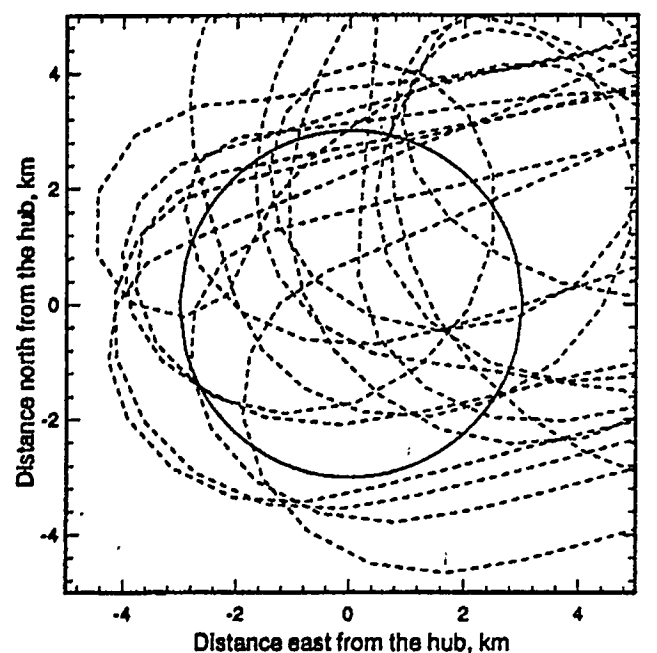


Figure 3. Aircraft flight tracks over the hub.

Another reason for enlarging the comparison area was because of the actual ground separation between the radar and the PRT-5 radiometer (about 800 m). This effect was mitigated by accounting for the cloud advection along the radar-radiometer line.

The solid line in Figure 3 shows the 3-km-radius circle with the center at the (0,0) point. Direct comparisons were made inside this circle. There were 15 aircraft passes into this circle from 1830 to 2130 UTC. Four of them occurred, however, when an intervening low altitude cloud containing liquid water phase prevented measuring the cirrus IR downwelling brightness temperature and, hence, no remotely sensed cirrus data were obtained. During one pass, the radar was performing RHI scans rather than taking data in the vertical mode required for obtaining cloud data remotely. In addition, during one more pass no PMS in situ data were recorded.

All these reasons left only 9 points for direct comparisons when in situ and remote samples were very close both in space and time.

Enlarging the circle radius to increase the number of points for direct comparisons of remotely sensed and in situ IMC values would create a risk in comparing significantly separate cloud volumes. This could lead to wrong conclusions, because, as revealed by both remote and aircraft measurements, IMC and other cloud parameters demonstrated a significant horizontal variability on a scale of few kilometers. Further reducing this radius would result in having too few points for comparisons and make the comparisons not representative.

Figure 4 shows the results of IMC comparisons. The in situ data for each aircraft pass were averaged over the time when the aircraft was inside the 3-km-radius circle. The mean time of each pass is shown near the corresponding comparison point. The one-to-one data correspondence line is also shown in Figure 4 for better reference. The variability of IMC values in the comparison samples exceeds two orders of magnitude for both remote and aircraft data.

The altitudes of compared samples covered mostly the middle and low part of the cloud and varied from about 6000 m above ground level (AGL) (2114 UTC) to about 8550 m AGL (1940 UTC).

The relative standard deviation of in situ aircraft data from remotely measured data shown in Figure 4 is about 53%. However, as seen in Figure 4, the differences between the remote and aircraft results for most of the comparison times (1950, 2020, 2037, 2100, 2105, and 2114 UTC) were better

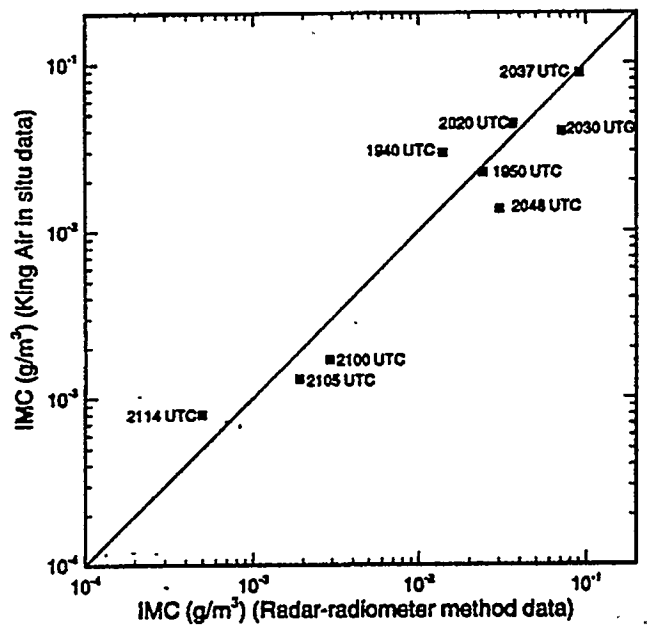


Figure 4. Comparisons of remote and aircraft data.

than that. The general agreement can be considered quite satisfactory given the uncertainty of both remote and 2D-probe measurements.

Conclusions

The November 26, 1991, cirrus case from FIRE-II provided a good opportunity to compare remotely measured cloud parameters with in situ aircraft measurements. Both remote and aircraft measurements showed a great (almost 4 orders of magnitude) variability of IMC during the experimental event from 1830 to 2130 UTC. The variability of measured

IMC values during times when the aircraft was passing in the vicinity of the hub exceeded two orders of magnitude. For these times the direct comparisons were made between remote and in situ measurements. These comparisons showed a 53% standard deviation between aircraft and remote data. Many individual comparisons showed agreement better than that.

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Raman Lidar Measurements of Water Vapor and Aerosols During the Atmospheric Radiation Measurement (ARM) Remote Cloud Sensing (RCS) Intensive Observation Period (IOP)

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The first Atmospheric Radiation Measurement (ARM) Remote Cloud Study (RCS) Intensive Operations Period (IOP) was held during April 1994 at the Southern Great Plains (SGP) Cloud and Radiation Testbed (CART) site near Lamont, Oklahoma. This experiment was conducted to evaluate and calibrate state-of-the-art, ground-based remote sensing instruments and to use the data acquired by these instruments to validate retrieval algorithms developed under the ARM program. These activities are part of an overall plan to assess general circulation model (GCM) parameterization research. Since radiation processes are one of the key areas included in this parameterization research, measurements of water vapor and aerosols are required because of the important roles these atmospheric constituents play in radiative transfer.

The National Aeronautics and Space Administration (NASA)/Goddard Space Flight Center (GSFC) Scanning Raman Lidar (SRL) acquired more than 123 hours of water vapor and aerosol profile data during 15 nights of operations. The lidar acquired vertical profiles as well as nearly horizontal profiles directed near an instrumented 60-m tower. We also deployed a multi-band automatic sun and sky scanning radiometer at this site which acquired aerosol and water vapor data throughout this experiment. Aerosol optical thickness, phase function, size distribution, and integrated water vapor were derived from a combination of sun and sky brightness measurements measured by this instrument. We used measurements from both instruments to assess the water vapor and aerosol measurements made by other instruments during the IOP, studied the structure and dynamics of a cold front which passed over the site on the night of April 15, 1994, and studied the aerosol optical and physical properties of aerosols.

Instruments

The GSFC Scanning Raman lidar uses an XeF excimer laser to transmit light at 351 nm. Light backscattered by molecules and aerosols at the laser wavelength is detected as well as Raman scattered light from water vapor (403 nm), nitrogen (382 nm), and oxygen (371 nm) molecules. The signals are gathered by a 76-cm-diameter telescope, detected by photomultiplier tubes, and recorded using photon counting. A steerable elliptical flat provides full 180 degree horizon to horizon scan capability within a single scan plane. The scan capability was used to increase the vertical resolution of the data as well as to facilitate comparisons with tower-based instruments. Water vapor mixing ratio was derived from the ratio of water vapor to nitrogen Raman signals; relative humidity was computed from the water vapor mixing ratio and radiosonde pressure and temperature measurements. Aerosol scattering ratio profiles were derived from the Raman nitrogen return signal and the signal detected at the laser wavelength; the aerosol volume backscattering cross section was then computed from the scattering ratio and from the molecular volume backscatter cross section calculated from coincident CART radiosonde density data. The aerosol extinction cross section is computed from the derivative of the Raman nitrogen return signal with respect to range. Details of these procedures are discussed by Ferrare et al. (1992, 1993) and Whiteman et al. (1992).

The Cimel Electronique 318A radiometer, which we deployed at the CART site, is a compact, weather-resistant system which operates independently of commercial electrical and communication systems. Two detectors permit the measurement of direct solar, aureole, and sky radiance.

During the RCS IOP, the instrument used six interference filters to acquire aerosol measurements at the following wavelengths: 339, 380, 441, 672, 873, and 1022 nm. Two additional filters at wavelengths near the 940-nm water vapor absorption band were used to estimate total precipitable water vapor. The sensor head is positioned by two microprocessor controlled motors to automatically acquire solar radiance observations throughout each day; aerosol optical thicknesses are derived from these observations. Additional sky radiance measurements made

at the solar zenith angle for specified azimuth angles (almucantar) are used to derive the aerosol size distribution using inversion algorithms (Kaufman et al. 1994).

Measurements

The SRL acquired vertical profiles as well as data at scan angles 5 and 10 degrees above the horizon on nine nights during the IOP. Figure 1 shows examples of images

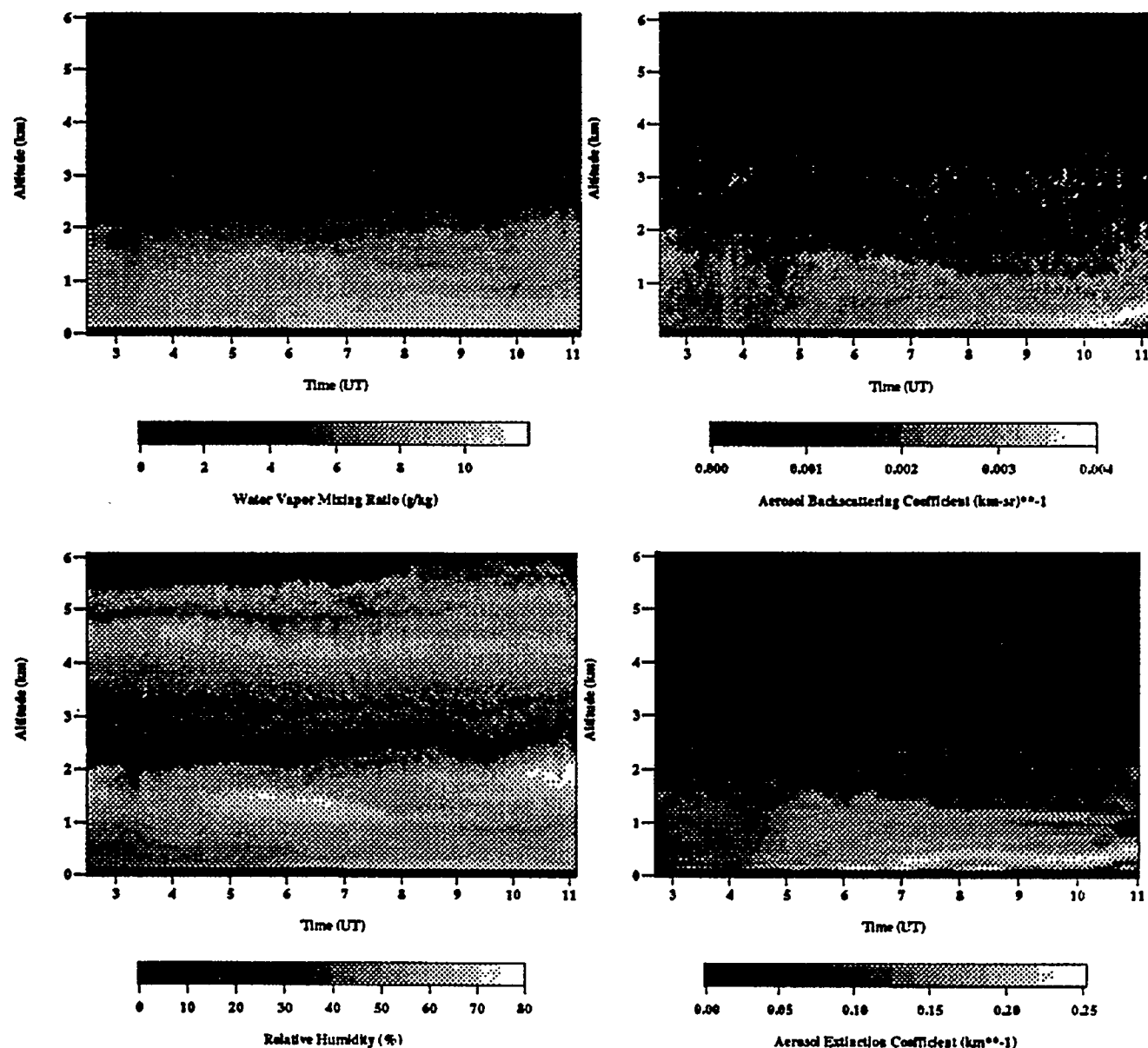


Figure 1. Water vapor mixing ratio (top left), relative humidity (bottom left), aerosol backscatter (top right), and aerosol extinction (bottom right) coefficients measured by Scanning Raman Lidar on April 21, 1994.

displaying water vapor mixing ratio, relative humidity, aerosol backscatter, and extinction coefficients acquired on April 21, 1994. These images, which were constructed using several hundred water vapor and aerosol profiles acquired simultaneously on the night of April 21, 1994, show water vapor and aerosols mainly concentrated in the residual layer below about 1.5 km. A dry, clean layer separates this residual layer from an elevated layer of water vapor and aerosols.

Figure 2 shows excellent agreement between the relative humidity profiles derived from SRL data acquired on this night and the humidity profiles measured by coincident Vaisala radiosondes launched at the CART site. On this same night, the University of North Dakota Citation aircraft flew above the CART site and acquired both water vapor and aerosol measurements. Both the lidar measurements of aerosol backscattering and extinction as well as the aerosol number density profiles measured by an optical particle counter on the aircraft showed increasing amounts of aerosols in the residual layer through this night.

The lidar and sun photometer measurements show changes in both water vapor and aerosols which occurred during this experiment. Figure 3 shows precipitable water vapor and aerosol optical thickness τ_a measured by the lidar at night as well as by the sun photometer during the day. The precipitable water and aerosol optical thickness values were

generally highly correlated during the IOP; increases in both the precipitable water vapor and aerosol optical thickness between April 19-25, 1994, were indicative of changes in air mass characteristics over the CART site at this time.

The simultaneous increase in the wavelength dependence of aerosol optical thickness measured by the sun photometer during this period was most likely caused by an increase in the number of small particles associated with a change in the air mass as shown by air trajectory analyses. The lidar measurements of aerosol backscattering and extinction also indicated changes in aerosol properties during the period.

In addition to studying changes in aerosol optical properties, the lidar measurements have been used to study atmospheric dynamics. The vertical structure and temporal evolution of a cold front was dramatically revealed by the lidar water vapor and aerosol data acquired on the night of April 15, 1994. Figure 4 shows water vapor mixing ratio and aerosol scattering ratio measured by the lidar. The lidar data show the increase of moisture ahead of the front, oscillations along the sloping frontal surface, and the decrease in moisture and aerosols after frontal passage. Figure 3 shows an increase in precipitable water and aerosol optical thickness prior to the passage of the cold front. Starr et al. (1995) have used these lidar data along with surface mesonet data to study the structure and dynamics of this cold front.

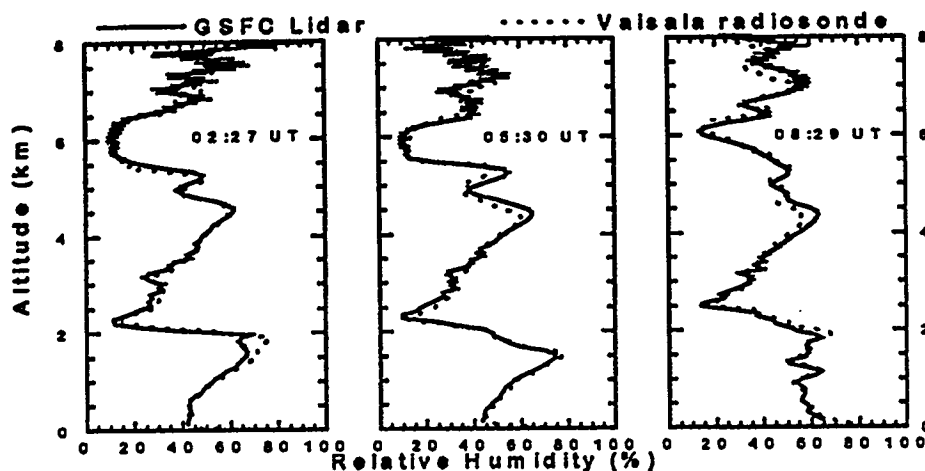


Figure 2. Relative humidity profiles derived from one-minute lidar water vapor profiles at 02:27 UT, 05:30 UT, and 08:29 UT on April 21, 1994. Profiles measured by coincident Vaisala radiosondes are also shown.

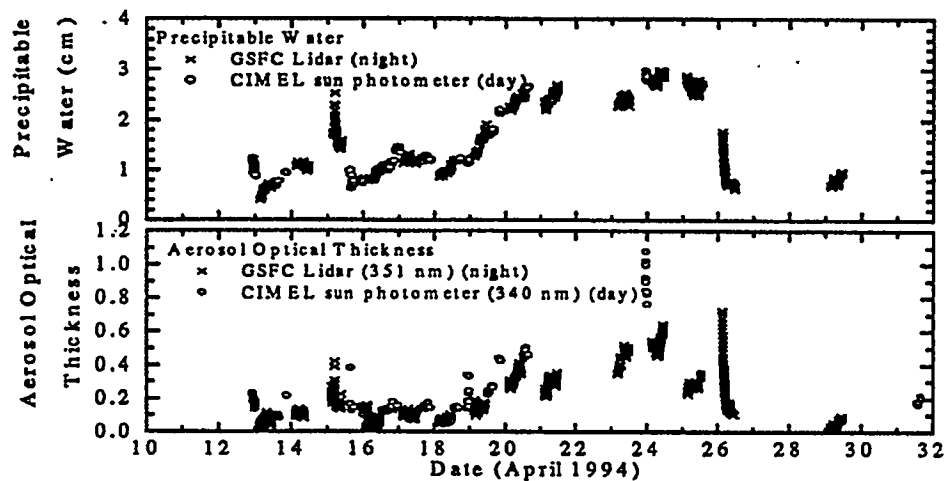


Figure 3. Precipitable water vapor (top) and aerosol optical thickness τ_a (bottom) measured by the Raman lidar and sun photometer.

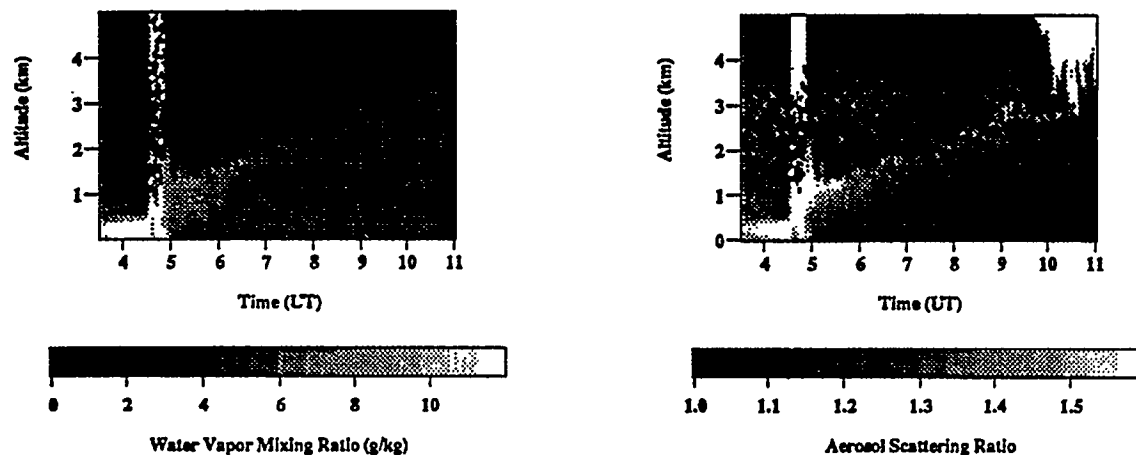


Figure 4. Water vapor mixing ratio (left) and aerosol scattering ratio (right) measured by the Scanning Raman Lidar showing the passage of a cold front over the CART site on the night of April 15, 1994.

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1. The first part of the document is a list of the names of the members of the committee.

Technical Progress Report: Completion of Spectral Rotating Shadowband Radiometers and Analysis of Atmospheric Radiation Measurement Spectral Shortwave Data

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Abstract

Our goal in the Atmospheric Radiation Measurement (ARM) Program is the improvement of radiation models used in general circulation models (GCMs), especially in the shortwave, 1) by providing improved shortwave radiometric measurements for the testing of models and 2) by developing methods for retrieving climatologically sensitive parameters that serve as input to shortwave and longwave models.

At the Atmospheric Sciences Research Center (ASRC) in Albany, New York, we are acquiring downwelling direct and diffuse spectral irradiance, at six wavelengths, plus downwelling broadband longwave, and upwelling and downwelling broadband shortwave irradiances that we combine with National Weather Service surface and upper air data from the Albany airport as a test data set for ARM modelers. We have also developed algorithms to improve shortwave measurements made at the Southern Great Plains (SGP) ARM site by standard thermopile instruments and by the multifilter rotating shadowband radiometer (MFRSR) based on these Albany data sets.

However, the major objective of our program has been the development of two spectral versions of the rotating shadowband radiometer. The MFRSR contains six filtered detectors, and one unfiltered silicon detector that serves as a surrogate total shortwave sensor. The MFRSR was completed in the first phase of this program and is deployed at the SGP cloud and radiation testbed (CART) site. It has been selected as the core instrument of the ARM solar and infrared observing system (SIROS) radiation instrument suite that is being deployed at each of twenty extended facilities within the SGP CART site. The rotating shadowband spectroradiometer (RSS) contains a 256-channel diode array that spans the wavelengths 350-1050 nm with resolution varying between 0.6 nm and

8 nm. The prototype of the RSS has acquired data on the roof of the ASRC at the State University of New York in Albany, but a field-worthy unit is not yet complete. A calibration facility for these instruments is being built at the ASRC to characterize cosine response, to determine absolute spectral calibration, and to characterize spectral response. The facility is operational, and upgrades are continuing.

The MFRSR, which was developed during the first years of the ARM Program, has become a workhorse at the CART sites in Oklahoma and Kansas, and it is widely deployed in other climate programs. We have spent most of our effort this year developing techniques to retrieve column aerosol, water vapor, and ozone from direct beam spectral measurements of the MFRSR. Additionally, we have had success in calculating shortwave surface albedo and aerosol optical depth from the ratio of direct to diffuse spectral irradiance. Using the surface albedo and the global irradiance, we have calculated cloud optical depths. From cloud optical depth and liquid water measured with the microwave radiometer, we have calculated effective liquid cloud particle radii. In each case, we have attempted to validate our approach using independent measurements or retrievals of the parameters under investigation. With the exception of the ozone inter-comparison, the corroborative measurements have been made at the SGP CART site. This report highlights these results.

ASRC Data Acquisition

A small part of our ARM funding is used to acquire a high-quality radiation data set on the roof of the ASRC using the MFRSR and the same standard thermopile pyranometers and pyrhemometers that are used at the SGP CART site. With the instruments accessible we are able to provide frequent cleaning, alignment, and calibration to ensure a

high-quality data set. The data have been used to test longwave and shortwave radiation models in a GCM. They allow testing of hardware and firmware changes to the MFRSR before they are implemented at the SGP CART site. They have enabled us to develop algorithms to improve both the accuracy of shortwave irradiance measurements made by the MFRSR and thermopile instruments.

The calibration of all of our shortwave sensors is tied to our absolute cavity radiometer. Last September we took part in the ARM absolute cavity intercomparison at the National Renewable Energy Laboratory. Although a final report has not been published, the preliminary results suggest that our absolute cavity agrees with the world standard to about 0.2%. This result validates the measurements used in the studies described here.

The shortwave and longwave downwelling radiation models in the Global Environmental and Ecological Simulation of Interactive Systems GCM have been tested for many clear days with the result that downwelling shortwave radiation is overestimated by about 20–30 W/m², and the downwelling longwave model largely agrees with measurements. The shortwave results from the ASRC GCM modeling team led by W. C. Wang are similar to those of Betts et al. (1993). Other unpublished results suggest similar conclusion.

The first-class thermopile radiation data at ASRC were used to derive a methodology to improve shortwave measurements made using the MFRSR unfiltered silicon detector. Temperature control of the detector and cosine response corrections as implemented in the MFRSR produced root-mean-square errors of about 11, 10, and 16 W/m² in the total, diffuse, and direct irradiances before spectral correction. Spectral correction further improved 1993 total and diffuse irradiance rms errors by about 20%. Mt. Pinatubo aerosols, however, no longer affected shortwave radiation in 1995, and the spectral corrections now appear to lower the errors in total and diffuse irradiance by about 50%.

Until recently, diffuse irradiance measurements made at the SGP CART site were a problem because tracking disks did not block the direct beam reliably. Improved trackers should solve the problem; however, the archived downwelling shortwave irradiances, which are best measured by summing diffuse and direct horizontal irradiances, are suspect. We have tested a procedure that uses direct and total horizontal measurements and a one-time measurement of the angular response of the total horizontal pyranometer to make improved shortwave measurements. Comparisons with our carefully monitored tracking disk diffuse measurements confirm the validity of this approach.

Instrument Development

The RSS prototype that we have developed uses a prism spectrometer and a charge-coupled device (CCD) photodiode array to measure 256 continuous wavelength intervals from 350 to 1050 nm. This instrument is intended for deployment at the central sites. A prototype of this instrument has been machined, assembled, and briefly tested in the laboratory and outdoors. The test results are very encouraging with the spectral resolution and instrument sensitivity very close to design values. With the departure of our post-doctoral student who had responsibility for this instrument, we have fallen behind schedule. Nevertheless, we expect to have the RSS operational on the ASRC roof before the end of the year.

Calibration

Radiometric calibration and characterization of our spectral instruments requires the development of three test platforms to measure the angular response, the spectral sensitivity, and the absolute response. All are computer controlled by the software Labview[®]. All of these facilities are operational; however, we expect to spend more effort improving the spectral response measurements. Currently, the output of the monochromator is below our expectations. The angular response and absolute calibration facilities are operating satisfactorily.

Analysis

Aerosol Optical Depth

Five of the six filters in an MFRSR may be used for the measurement of aerosol optical depth. When the atmosphere is clear and stable, Langley analysis is possible. The Bouguer-Lambert-Beer law is given by

$$I = I_0 e^{-\tau \cdot m} \quad (1)$$

where I is the spectral irradiance as measured by the MFRSR, I_0 is the spectral irradiance that the MFRSR would measure at the top of the atmosphere, τ is the total optical depth from all sources of extinction, and m is the air mass relative to the zenith direction. If the atmosphere is stable, a plot of the natural log of I versus m yields a linear plot whose slope is the total optical depth τ and whose intercept is I_0 . I_0 may be obtained from a standard lamp or derived from Langley plots according to the above procedure on occasions when the atmosphere is stable. When the

atmosphere is free of clouds, but unstable because of changing aerosol conditions, we may calculate τ using this measured I_0 . We have used our estimate of I_0 to calculate the averages of thirty-minute time series of optical depth. We allow the optical depth to change linearly in time, but if any point deviates by more than ± 0.01 optical depths from this linear behavior, we advance one time step and test the series again. When we find thirty minutes of data that satisfy the criterion, we advance thirty minutes and begin the search again. Figure 1 is a plot of daily averages of these thirty-minute samples.

Water Vapor

Reagan et al. (1987) introduced the modified Langley technique as an independent method of calculating total column water vapor using the 940-nm water band in the near infrared. Bruegge et al. (1992) followed the Reagan method, but used LOWTRAN7 to calculate transmission in their water vapor passband. We have followed the procedure of Bruegge et al. (1992), but used MODTRAN2 for our transmission calculations. Derived water vapor column abundances for 104 clear mornings or afternoons have been

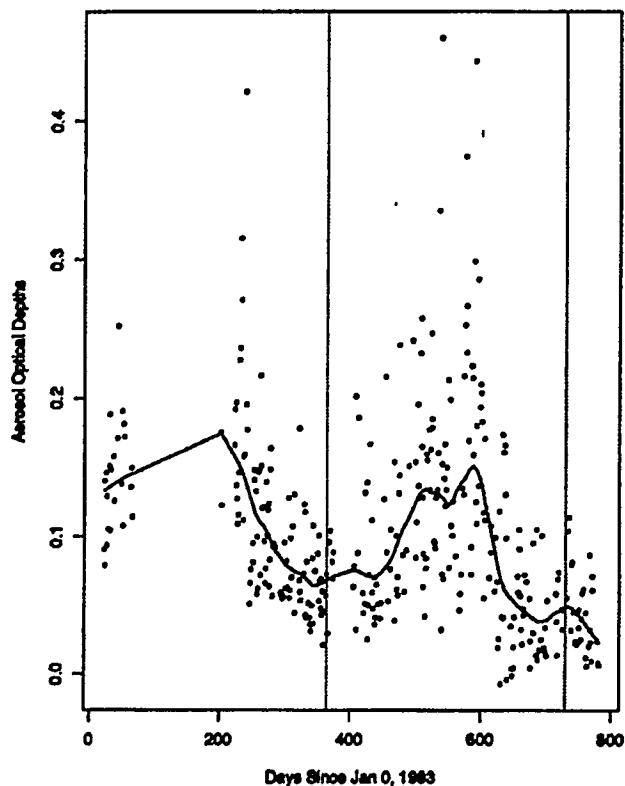


Figure 1. Daily-averaged aerosol optical depths at 500 nm for SGP central facility.

compared to water vapor columns derived from microwave measurements made simultaneously at the SGP CART site central facility. These measurements span a year and cover the range of water vapor experienced at the SGP site. Figure 2 is a plot of the MFRSR-derived water vapor versus the microwave radiometer measurements for the same times.

Ozone

Ozone may also be derived using the five filters used in the aerosol optical depth derivation above. Realistic aerosol size distributions produce little curvature on a plot of the natural log of the aerosol optical depth versus the natural log of the wavelength at which the optical depth is measured. This can be shown theoretically, and it is our experience experimentally. King and Byrne (1976) suggested that one could estimate this wavelength dependence with a quadratic function in these variables. By perturbing the ozone optical depth until a best quadratic fit is obtained, we can derive an estimate of ozone column abundance. Figure 3 is a plot of ozone column for the last four months of 1994 for the Dobson spectrophotometer at Mauna Loa Observatory and for the MFRSR. The Dobson is a standard for ozone measurements.

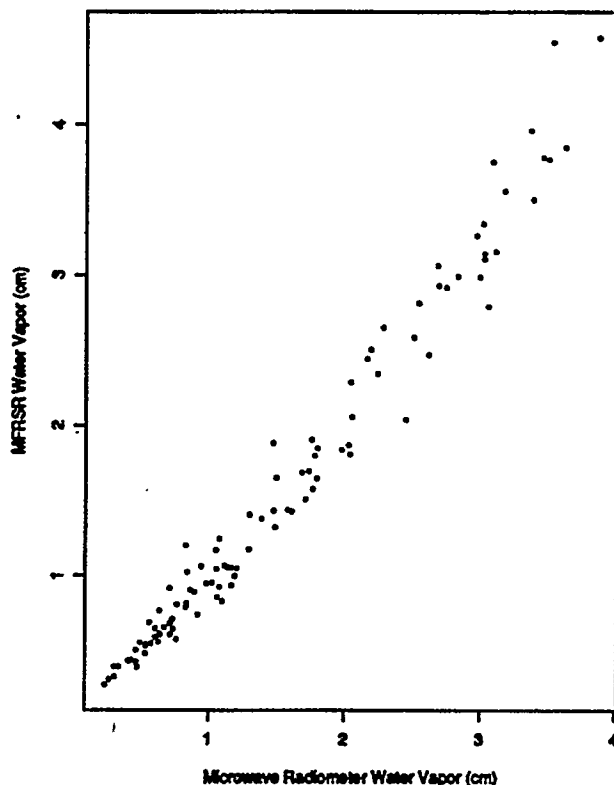


Figure 2. MFRSR-derived water vapor versus microwave radiometer water vapor at SGP.

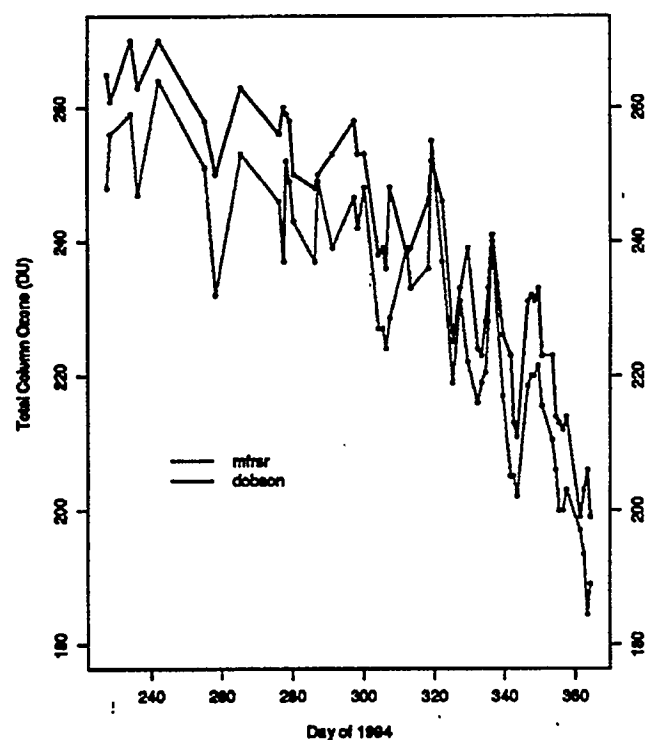


Figure 3. Mauna Loa Observatory total column ozone from the Dobson spectrometer and the MFRSR.

Radiative Transfer and Surface Albedo

A certain amount of received solar radiation is multiply scattered, requiring a more rigorous treatment of scattering processes through the use of radiative transfer models that include all orders of multiple scattering. However, full radiative transfer computations are extremely time-consuming, limiting the applications of the radiative transfer model. Noting that the response of interest is merely the downward surface flux which is characterized by very large data bases (i.e., as a function of solar zenith angle), we developed the adjoint formulation of discrete ordinate radiative transfer to understand the measurements of the MFRSR.

Based on the definition of adjoint operator, $\langle I^*, LL \rangle = \langle L^* I^*, I \rangle$, we constructed a variational principle for the forward radiative transfer equation. It led to the adjoint formulation of the radiative transfer equation, $L^* I^* = \Sigma$, with appropriate boundary conditions. The adjoint operator is

$$L^*(Z, \Omega) = -\mu \frac{d}{dz} + \beta^{\text{ext}} - \beta^{\text{sca}} \int_{4\pi} d\Omega' p(z, -\Omega' \rightarrow -\Omega) \quad (2)$$

and the adjoint source for the downward diffuse irradiance is

$$\Sigma = -\mu \sigma(z) U(-\mu). \quad (3)$$

The downward diffuse irradiance, $G[I]$, can be evaluated as equation

$$G[I] = \langle \Sigma, I \rangle = \langle L^* I^*, I \rangle = \langle I^*, LL \rangle = \langle I^*, Q \rangle \quad (4)$$

To calculate the radiative effects of m different sources, we must solve the forward radiative transfer equation m times. However, such effects can be evaluated by solving the adjoint radiative transfer equation once and undertaking m inner products instead. It is demonstrated that the accuracy of the adjoint method is equal to that of the forward discrete ordinate radiative transfer method, and the computing time is significantly reduced.

The irradiance at the surface, as well as the diffuse/direct ratio, nonlinearly depends upon several factors such as wavelength, solar zenith angle, surface albedo, and atmospheric species including ozone, cloud, and aerosol. We employed the adjoint method of radiative transfer in conjunction with the nonlinear least squares method to infer the surface albedo and the aerosol optical depth simultaneously from the ratio of diffuse to direct irradiances. We processed the data at wavelengths of 415, 500, 610, and 665 nm for the time intervals during which there is a clear stable atmosphere devoid of any cloud cover for a long enough period of time for the airmass to undergo a large change, particularly on the days in which the optical depths are large. We compared the inferred surface albedo with measurements at the SGP site of ARM and the inferred optical depth with the results of Langley regression. The data in Table 1 show reasonable agreement, when cloud-free conditions prevail.

Cloud Properties

Once the parameters related to surface albedo and aerosol have been retrieved, we can use the same approach to infer stratified cloud properties under cloud-covered sky. Specifying the single scattering albedo and the asymmetry factor based on Mie theory, we can infer the cloud optical depth from the global irradiance. We selected the transmittance of the atmosphere, the ratio of measured global irradiance to the solar constant, as the inversion function to avoid the uncertainties of absolute calibration.

Table 1. Summary of the observed and inferred surface albedo and total optical depth during August 1994 at SGP.

Time	Wavelength λ (nm)	Surface Albedo			Optical Depth	
		NLSM	A_{10m}	A_{25m}	NLSM	Langley Reg.
940816	415	0.036 ± 0.013	0.031	0.038	0.487 ± 0.009	0.503 ± 0.009
	500	0.066 ± 0.011	0.048	0.063	0.285 ± 0.006	0.299 ± 0.004
	610	0.128 ± 0.016	0.071	0.109	0.198 ± 0.007	0.214 ± 0.004
	665	0.162 ± 0.019	0.079	0.126	0.144 ± 0.007	0.162 ± 0.004
940821	415	0.094 ± 0.007	0.025	0.032	0.584 ± 0.004	0.589 ± 0.012
	500	0.088 ± 0.007	0.045	0.052	0.372 ± 0.004	0.373 ± 0.004
	610	0.122 ± 0.010	0.069	0.089	0.263 ± 0.004	0.266 ± 0.004
	665	0.137 ± 0.009	0.074	0.105	0.197 ± 0.004	0.207 ± 0.004

We compared our inferred cloud optical depth with Geostationary Operational Environmental Satellite (GOES) measurements at the SGP site during April 1994. Temporal variations of our results, shown in Figure 4, are consistent with that of GOES measurements. However, our results are slightly higher, which may be due to the effects of spatial average of GOES measurements over a 0.3° by 0.3° area.

Publications and Presentations

The following publications and presentations were completed between November 1, 1994 and October 31, 1995:

Harrison, L., J. Michalsky, and J. Berndt. 1994. Automated multifilter rotating shadow-band radiometer: An instrument for optical depth and radiation measurements, *Appl. Opt.*, **33**, 5118-5125.

Harrison, L., and J. Michalsky. 1994. Objective algorithms for the retrieval of optical depths from ground-based measurements, *Appl. Opt.*, **33**, 5126-5132.

Michalsky, J. J., L. C. Harrison, and W. E. Berkheiser III. 1995. Cosine response characteristics of radiometric and photometric sensors, *Solar Energy*, **54**.

Michalsky, J. J., J. C. Liljegren, and L. C. Harrison. 1995. A comparison of sun photometer derivations of total column water vapor and ozone to standard measures of same at the Southern Great Plains Atmospheric Radiation Measurement Site, *Journal of Geophysical Research-Atmospheres*, in revision.

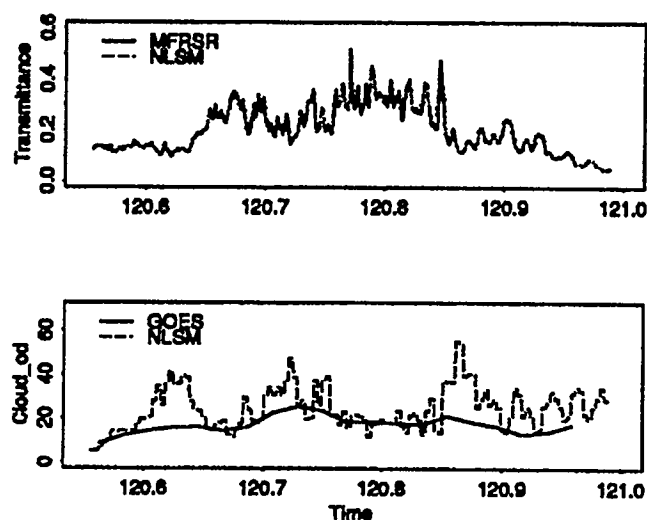


Figure 4. Measured and fitted transmittances and inferred cloud optical depths on April 30, 1994, at SGP.

Zhou, Chuan, and Joseph Michalsky. 1995. Spectral correction of silicon photodiode solar radiation detectors, *Solar Energy*, in revision.

Four papers have been presented since the last annual report:

Michalsky, J. J., J. A. Schlemmer, N. R. Larson, L. C. Harrison, W. E. Berkheiser III, and N. S. Laulainen. 1994. Measurement of the seasonal and annual variability of total column aerosol in a northeastern network. In *Proceedings of the International Specialty Conference Aerosols and Atmospheric Optics: Radiative Balance and Visual Air Quality, 26-30 September 1994, Snowbird, Utah*. W. Malm (ed). Air and Waste Management Association, Pittsburgh, Pennsylvania, pp. 247-258.

Michalsky, J., and H. Harrison. 1995. Total column ozone retrieval from the visible chappuis band: Comparisons with standard ultraviolet measurements. In *Optical Remote Sensing of the Atmosphere, Vol. 2, 1995*, OSA Technical Digest Series, Optical Society of America, Washington, D.C., pp. 195-197.

Michalsky, J. J., and E. Dutton. 1995. Comparison of standard and MFRSR sunphotometry at Mauna Loa Observatory. In *Climate Monitoring and Diagnostics Laboratory Annual Meeting*, March 8-9, 1995.

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Acknowledgment

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Bruegge, C. J., J. E. Conel, R. O. Green, J. S. Margolis, R. G. Holm, and G. Toon. 1992. Water vapor column abundance retrievals during FIFE, *J. Geophys. Res.*, **97**, 18759-18768.

King, M. D., and D. M. Byrne. 1976. A method for inferring total column ozone from the spectral variation of total optical depth obtained with a solar radiometer, *J. Atmos. Sci.*, **33**, 2242-2251.

Reagan, J. A., K. Thome, B. Herman, and R. Gall. 1987. Water vapor measurements in the 0.94 micron band: Calibration, measurements and data applications. In *Proceedings of the International Geosciences Remote Sensing Symposium*, IEEE 87CH 2434-9, 63-67.

Rotating Shadowband Radiometer Development and Analysis of Spectral Shortwave Data

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Introduction

Our goals in the Atmospheric Radiation Measurement (ARM) Program are improved measurements of spectral shortwave radiation and improved techniques for the retrieval of climatologically sensitive parameters.

The multifilter rotating shadowband radiometer (MFRSR) that was developed during the first years of the ARM program has become a workhorse at the Southern Great Plains (SGP) Cloud and Radiation Testbed (CART) site, and it is widely deployed in other climate programs. We have spent most of our effort this year developing techniques to retrieve column aerosol, water vapor, and ozone from direct beam spectral measurements of the MFRSR. Additionally, we have had some success in calculating shortwave surface albedo and aerosol optical depth from the ratio of direct to diffuse spectral irradiance. Using the surface albedo and the global irradiance, we have calculated cloud optical depths. From cloud optical depth and liquid water measured with the microwave radiometer, we have calculated effective liquid cloud particle radii.

The rest of the text will provide some detail regarding each of these efforts.

Aerosol Optical Depth

Five of the six filters in an MFRSR may be used for the measurement of aerosol optical depth. When the atmosphere is clear and stable, Langley analysis is possible. The Bouguer-Lambert-Beer law is given by

$$I = I_0 e^{-\tau \cdot m} \quad (1)$$

where I is the spectral irradiance as measured by the MFRSR, I_0 is the spectral irradiance that the MFRSR would measure at the top of the atmosphere, τ is the total optical depth from all sources of extinction, and m is the air mass

relative to the zenith direction. If the atmosphere is stable, a plot of the natural log of I versus m yields a linear plot whose slope is the total optical depth τ . The five filters that we use for aerosol measurements are sensitive to aerosol scattering/absorption and Rayleigh scattering at all wavelengths and to ozone absorption in three of the filters. Rayleigh scattering depends only on surface pressure and is easily subtracted. Ozone may either be measured independently or inferred from the data (which we discuss below) and subtracted. This yields the aerosol optical depth.

I_0 may be obtained from a standard lamp or derived from Langley plots according to the above procedure on occasions when the atmosphere is stable. When the atmosphere is free of clouds, but unstable because of changing aerosol conditions, we may calculate τ using this measured I_0 . We have used our estimate of I_0 to calculate the averages of thirty-minute time series of optical depth. We allow the optical depth to change linearly in time, but if any point deviates by more than ± 0.01 optical depths from this linear behavior, we advance one time step and test the series again. When we find thirty minutes of data that satisfy the criterion, we advance thirty minutes and begin the search again. Figure 1 is a plot of daily averages of these thirty-minute samples. This daily average is used, rather than each thirty-minute sample, when looking for seasonal behavior because it does not preferentially weight the clearest days.

Water Vapor

Reagan et al. (1987) introduced the modified Langley technique as an independent method of calculating total column water vapor using the 940-nm water band in the near infrared. Bruegge et al. (1992) followed the Reagan method, but used LOWTRAN7 to calculate transmission in their water vapor passband. We have followed the procedure of Bruegge et al. (1992), but used MODTRAN2 for our transmission calculations. Derived water vapor column abundances for 104 clear mornings or afternoons have been compared to water vapor columns derived from microwave measurements made simultaneously at the SGP CART site central facility. These measurements span a year and cover

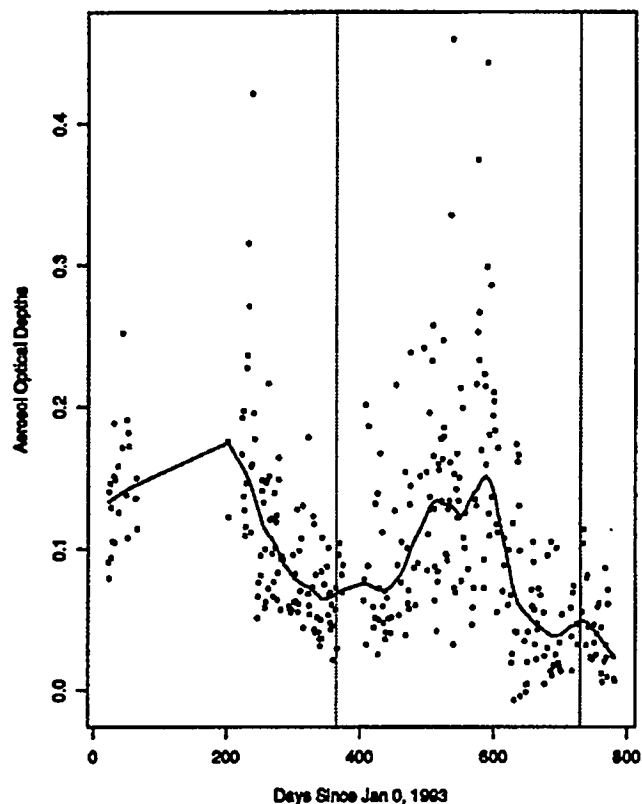


Figure 1. Daily-averaged aerosol optical depths at 500 nm for the SGP site central facility.

the range of water vapor experienced at the SGP site. Figure 2 is a plot of the MFRSR-derived water vapor versus the microwave radiometer measurements for the same periods. At high water vapor the MFRSR reports slightly higher values than the microwave radiometer. At this point the slight discrepancy is not completely resolved.

Ozone

Ozone may also be derived using the five filters used in the aerosol optical depth derivation as suggested above. Realistic aerosol size distributions produce little curvature on a plot of the natural log of the aerosol optical depth versus the natural log of the wavelength for which the optical depth is obtained. This can be shown theoretically, and it is our experience experimentally. King and Byrne (1976) suggested that one could estimate this wavelength dependence with a quadratic function in these variables. By perturbing the ozone optical depth until a best quadratic fit is obtained, we can derive an estimate of ozone column abundance. Figure 3 is a plot of ozone column for the last four months

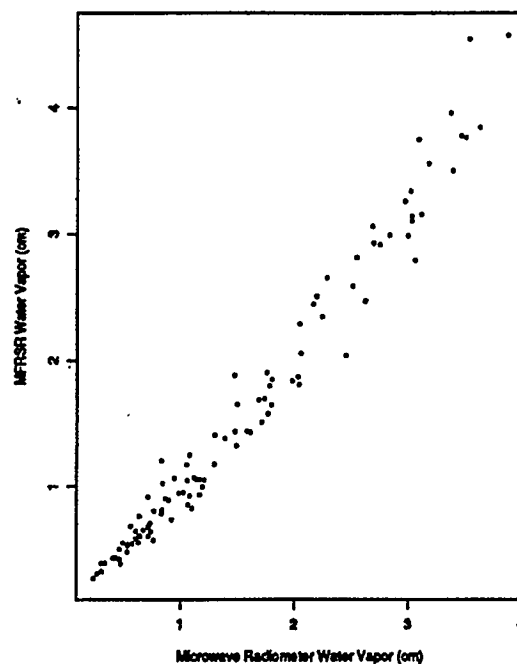


Figure 2. MFRSR-derived water vapor versus microwave radiometer water vapor at the SGP site.

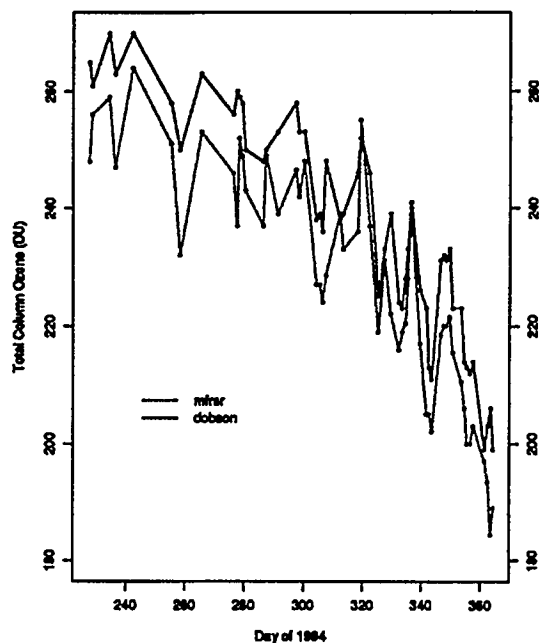


Figure 3. Mauna Loa Observatory total column ozone from the Dobson spectrometer and the MFRSR.

of 1994 for the Dobson spectrophotometer at Mauna Loa Observatory and for the MFRSR. The Dobson is a standard for ozone measurements. The MFRSR data seem to follow closely but have a 10 Dobson-unit offset from the Dobson data. Filter stability is crucial for this technique to be valid, therefore we suspect the minor offset is associated with a filter shift. We will test this hypothesis further by recalibrating and by testing the method at other sites where standard ozone measurements are made and an MFRSR is acquiring data simultaneously.

Radiative Transfer and Surface Albedo

A certain amount of received solar radiation is multiply scattered, requiring a more rigorous treatment of scattering processes through the use of radiative transfer models that include all orders of multiple scattering. However, full radiative transfer computations are extremely time-consuming, limiting the applications of the radiative transfer model. Noting that the response of interest is merely the downward surface flux which is characterized by very large data bases (i.e., as a function of solar zenith angle), we developed the adjoint formulation of discrete ordinate radiative transfer to understand the measurements of the MFRSR.

Based on the definition of adjoint operator, $\langle I^*, LL \rangle = \langle L^* I^*, I \rangle$, we construct a variational principle for the forward radiative transfer equation. It leads to the adjoint formulation of the radiative transfer equation, $L^* I^* = F$, with appropriate boundary conditions. The adjoint operator is

$$L^*(z, \Omega) = -\mu \frac{d}{dz} + \beta^{\text{ext}} - \beta^{\text{sca}} \int_{4\pi} d\Omega' p(z, -\Omega' \rightarrow -\Omega) \quad (2)$$

and the adjoint source for the downward diffuse irradiance is $\Sigma = -\mu \delta(z) U(-\mu)$. The downward diffuse irradiance, $G[I]$, can be evaluated as

$$G[I] = \langle \Sigma, I \rangle = \langle L^* I^*, I \rangle = \langle I^*, LL \rangle = \langle I^*, Q \rangle \quad (3)$$

The advantage of the adjoint method to evaluate the integral property is clearly demonstrated in the above equation. To calculate the radiative effects of m different sources, we must solve the forward radiative transfer equation m times. However, such effects can be evaluated by solving the adjoint radiative transfer equation once and undertaking m inner products instead.

The efficiency and accuracy of our adjoint method of radiative transfer are tested by comparing the results with the forward method of radiative transfer for a set of synthetic cases and for measurements with the MFRSR. It is demonstrated that the accuracy of the adjoint method is equal to that of the forward discrete ordinate radiative transfer method, and the computing time is significantly reduced.

The irradiance at the surface, as well as the diffuse/direct ratio, nonlinearly depends upon several factors such as wavelength, solar zenith angle, surface albedo, and atmospheric species including ozone, cloud, and aerosol. We employed the adjoint method of radiative transfer in conjunction with the nonlinear least squares method to infer the surface albedo and the aerosol optical depth simultaneously from the ratio of diffuse to direct irradiances.

We processed the data at wavelengths of 415, 500, 610, and 665 nm for the time intervals during which there is a clear stable atmosphere devoid of any cloud cover for a long enough period of time for the airmass to undergo a large change, particularly on the days in which the optical depths are large. We compared the inferred surface albedo with measurements at the SGP site of ARM and the inferred optical depth with the results of Langley regression. The data in Table 1 show reasonable agreement, when cloud-free conditions prevail.

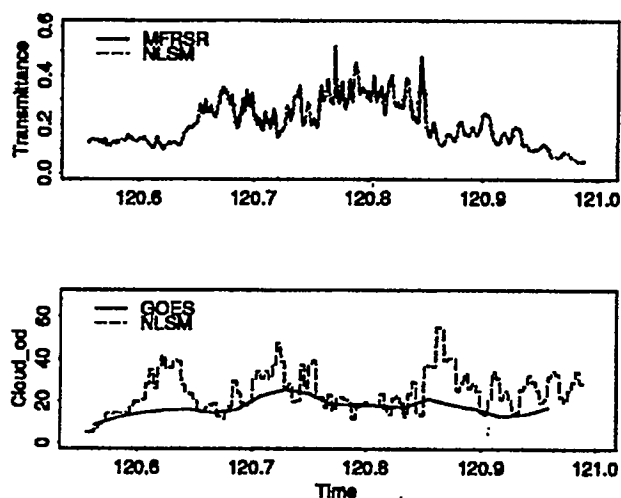
Cloud Properties

Once the parameters related to surface albedo and aerosol have been retrieved, we can use the same approach to infer stratified cloud properties under cloud-covered sky. Specifying the single scattering albedo and the asymmetry factor based on Mie theory, we can infer the cloud optical depth from the global irradiance. We selected the transmittance of the atmosphere, the ratio of measured global irradiance to the solar constant, as the inversion function to avoid the uncertainties of absolute calibration.

We compared our inferred cloud optical depth with the Geostationary Operational Environmental Satellite (GOES) measurements at the SGP site during April 1994. Temporal variations of our results, shown in Figure 4, are consistent with that of GOES measurements. However, our results are slightly higher, which may be due to the effects of spatial average of GOES measurements over a 0.3° by 0.03° area.

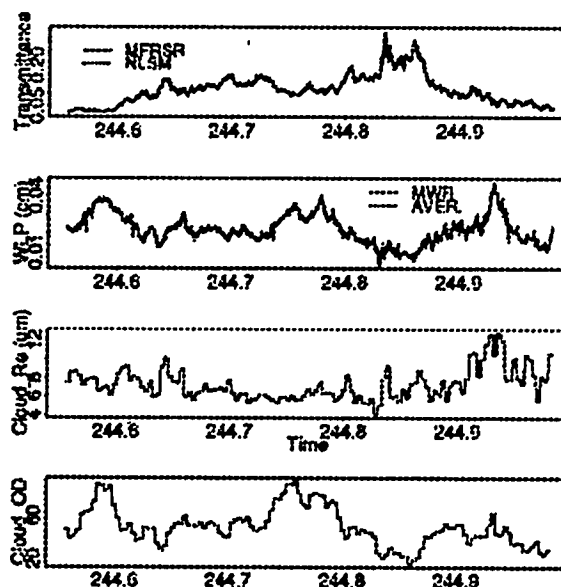
Table 1. Summary of the observed and inferred surface albedo and total optical depth during August 1994 at SGP.

Time	Wavelength $\lambda(\text{nm})$	Surface Albedo			Optical Depth	
		NLSM	Δ_{10m}	Δ_{25m}	NLSM	Langley Reg.
940816	415	0.036 ± 0.013	0.031	0.038	0.487 ± 0.009	0.503 ± 0.009
	500	0.066 ± 0.011	0.048	0.063	0.285 ± 0.006	0.299 ± 0.004
	610	0.128 ± 0.016	0.071	0.109	0.198 ± 0.007	0.214 ± 0.004
	665	0.162 ± 0.019	0.079	0.126	0.144 ± 0.007	0.162 ± 0.004
940821	415	0.094 ± 0.007	0.025	0.032	0.584 ± 0.004	0.489 ± 0.012
	500	0.088 ± 0.007	0.045	0.052	0.372 ± 0.004	0.373 ± 0.004
	610	0.122 ± 0.010	0.069	0.089	0.263 ± 0.004	0.266 ± 0.004
	665	0.137 ± 0.009	0.074	0.105	0.197 ± 0.004	0.207 ± 0.004

**Figure 4.** Measured and fitted transmittances and inferred cloud optical depths on April 30, 1994, at the SGP site.

Knowing the liquid water path from microwave radiometry, we can infer the effective radius of cloud drops through the relationship defined by Mie theory. We processed the data at the SGP site during August and September 1994. The inferred mean cloud particle radius and cloud optical depths are sensible. For example, on

September 1, 1994, the inferred mean cloud particle radii vary from $4.2 \mu\text{m}$ to $12.8 \mu\text{m}$ as cloud optical depths vary from 16.0 to 91.8, respectively (Figure 5).

**Figure 5.** Transmittances, water liquid paths, inferred cloud effective radii, and cloud optical depths on Sept. 1, 1994, at the SGP site.

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Effect of Particle Nonsphericity on Bidirectional Reflectance of Cirrus Clouds

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Previous analyses of the First International Satellite Cloud Climatology Project (ISCCP) Regional Experiment (FIRE) measurements have been interpreted as indicative of asymmetry parameters of the cirrus cloud particle phase function on the order of 0.75 or smaller (Wielicki et al. 1990; Kinne et al. 1994). Ray tracing computations for single or aggregated hexagonal ice crystals cannot produce such small asymmetry parameters (Takano and Liou 1989; Macke 1994). Furthermore, there is accumulating evidence that the idealized phase function associated with regular hexagonal particles and exhibiting pronounced halos is not representative of the reflectance characteristics of the majority of cirrus clouds (Minnis et al. 1993; Sassen et al. 1994; Wielicki et al. 1990; Francis 1995). On the other hand, ray tracing computations of the phase function for irregular fractal ice particles (Macke 1994) show no halo features and can have asymmetry parameters approaching 0.75. Therefore, in this paper we use the fractal ice particle model to study the differences in bidirectional reflectance caused by the differences in the single scattering phase functions of spherical water droplets and nonspherical ice crystals.

The solid line in Figure 1 shows the phase function for randomly oriented, irregular, fractal particles with shape depicted in the insert. The size-parameter-independent scattering component of the phase function was computed using the ray-tracing method (Macke 1994), while the size-parameter-dependent diffraction component was averaged over a size distribution with an effective radius $r_{\text{eff}} = 30 \mu\text{m}$. Remarkably, this phase function not only shows no pronounced halo features, but also has an asymmetry parameter as small as 0.752. For comparison, the dotted line shows the phase function for the water cloud particle model (gamma distribution of spherical droplets with an effective radius $r_{\text{eff}} = 10 \mu\text{m}$) used in the ISCCP (Rossow et al. 1991).

We have used the two phase functions in accurate multiple scattering calculations for plane-parallel water and ice clouds with varying optical thickness. Bidirectional cloud reflectivity was computed using the standard adding/doubling method (Hansen and Travis 1974) without

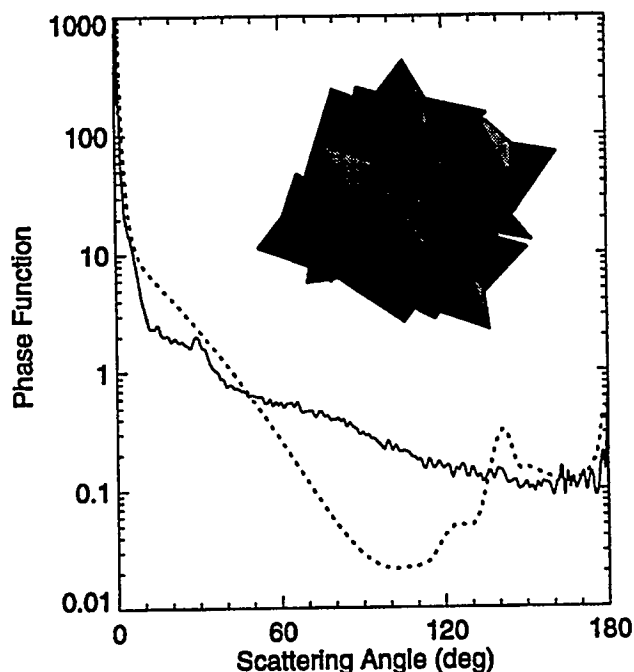


Figure 1. Phase functions of fractal ice particles (solid line) and ISCCP water droplets (dotted line) at $0.63 \mu\text{m}$.

introducing any further approximations like the truncation of the forward-scattering peak of the phase function. The number of Gaussian quadrature points in zenith angle discretization and the number of terms in the Fourier decomposition of the reflection function were increased until the relative accuracy of computing the reflection function was better than 10^{-3} .

Figure 2 shows the ratio $\rho(\tau_i)$ of the reflectivity computed for the ice cloud with optical thicknesses $\tau_i = 0.3, 0.1, 3$, and 300 relative to that of the liquid water cloud with optical thickness $\tau_w = 1$. Cosines of the sun, μ_0 , and satellite, μ , zenith angles vary from 0.2 to 1, and relative satellite-sun azimuth angles are $\varphi = 0, 90$, and 180° .

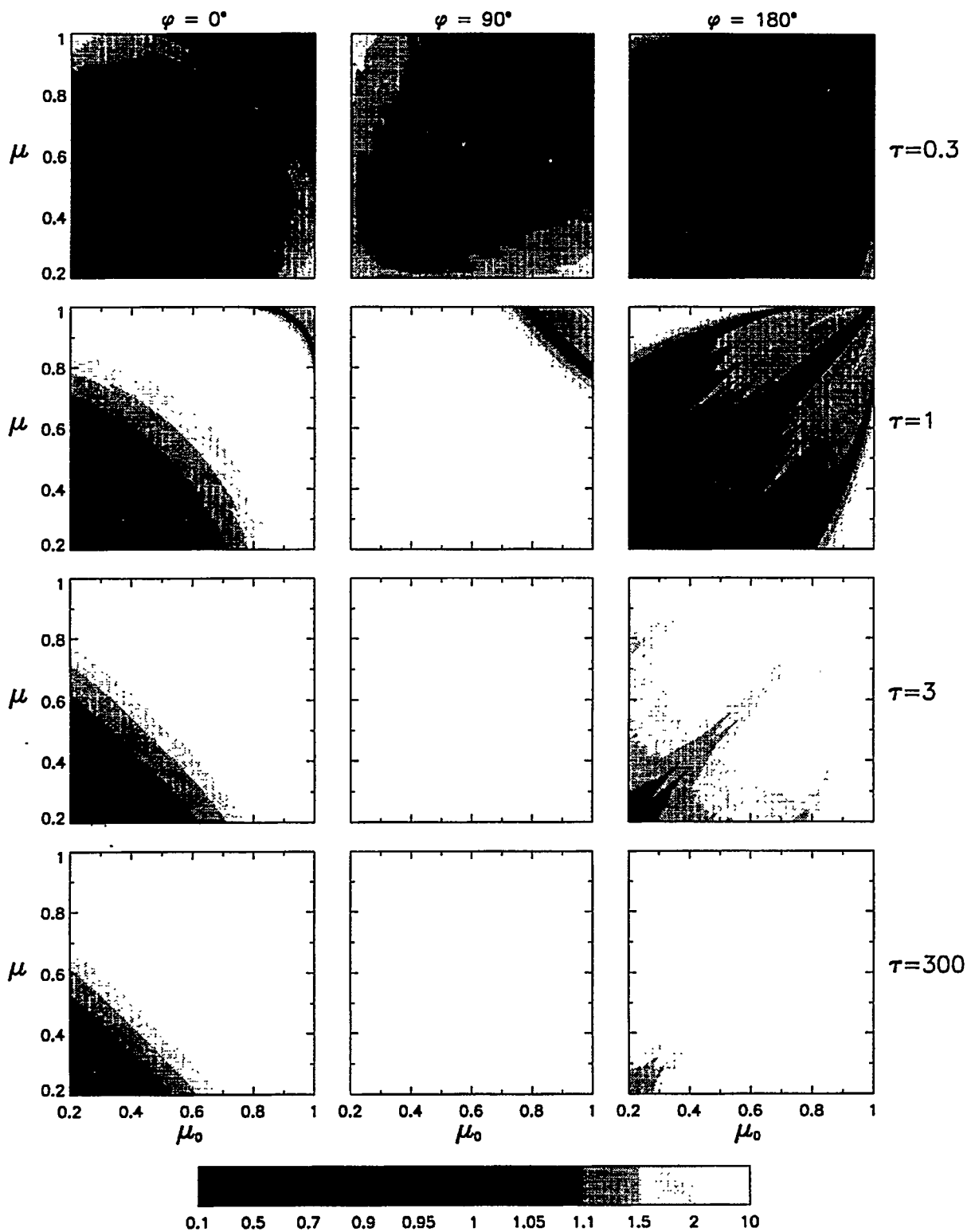


Figure 2. Ratio $\rho(\tau_i)$ of the reflectivity computed for the ice cloud with optical thicknesses $\tau_i = 0.3, 0.1, 3,$ and 300 relative to that of the liquid water cloud with optical thickness $\tau_w = 1$. Solid lines show contours at the level $\rho = 1$.

Computations for $\tau_i = 1$ show that nonspherical/spherical differences in reflectance exactly follow those in the single-scattering phase functions (Figure 1). Specifically, the ice cloud reflectivity is larger at side-scattering angles and smaller at near-forward- and back-scattering angles. Furthermore, the large single-scattering differences cause values of ρ larger than two for $\tau_i \ll \tau_w$ and values of ρ smaller than 0.5 for $\tau_i \gg \tau_w$. Remarkably, at $\varphi = 0$ and μ and μ_0 smaller than about 0.5, the reflectivity of the water cloud with $\tau_w = 1$ cannot be reproduced by the ice cloud with optical thickness as large as 300. Our computations show that the region of $\rho < 1$ at $\varphi = 0$ and small μ and μ_0 survives even if the ice cloud is semi-infinite. This means that at near-forward-scattering geometries water clouds with optical thickness of order one can produce reflectivities that cannot be matched by ice clouds with an arbitrarily large optical thickness.

Figure 2 can be used to examine the errors in the retrieved cloud optical thickness introduced by the use of the wrong cloud particle model. Specifically, the contours at the level $\rho = 1$ indicate the retrieved optical thickness as a function of μ , μ_0 , and φ if the ice cloud model is used to analyze reflectance measurements for a liquid water cloud with optical thickness $\tau_w = 1$. Figure 2 shows that the retrieved optical thickness is strongly scattering-geometry-dependent and can substantially differ from the actual value 1 (Mishchenko et al. 1995). Following the differences in the single-scattering phase functions (Figure 1), the ice cloud model overestimates the retrieved optical thickness at near-forward- and back-scattering geometries and underestimates it at side-scattering geometries. The errors in the retrieved optical thickness are especially large at near-forward-scattering geometries (i.e., at $\varphi = 0$ and small μ and μ_0). Furthermore, at μ and μ_0 smaller than 0.5, the retrieval scheme using the ice particle model produces no solution at all.

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Radiative Properties of Ice Clouds

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Introduction

A new treatment of cirrus cloud radiative properties has been developed, based on anomalous diffraction theory (ADT), which does not parameterize size distributions in terms of an effective radius (partially published in Mitchell and Arnott 1994). Rather, it uses the size distribution parameters directly, and explicitly considers the ice particle shapes. There are three fundamental features which characterize this treatment: 1) the ice path radiation experiences as it travels through an ice crystal is parameterized, 2) only the physical cross-section or projected area of the particle determines the amount of radiation scattered and absorbed, and 3) as in other treatments, the projected area of the size distribution is conserved. The first two features are unique to this treatment, since it does not convert the ice particles into equivalent volume or area spheres in order to apply Mie theory.

It may not be obvious that the second feature differs from Mie theory. However, Mie theory predicts that the absorption and extinction cross-section of a sphere can be (and often is) substantially greater than would be predicted from its physical cross-section. This involves the capture of photons which have approximately tangential trajectories to the sphere. Once captured, they can either be scattered or absorbed into the sphere. By capturing photons which do not actually collide with the sphere, the absorption cross-section for a sphere becomes greater than the sphere's physical cross-section. Now, the question is, does this same physics also apply to ice?

Testing the Cirrus Radiation Scheme with Laboratory Measurements

Measurements of extinction efficiency (Q_{ext}) were made in a laboratory ice cloud over wavelengths in the thermal infrared. Mean size parameters ($x = \pi \bar{D}/\lambda$, $\bar{D}$ = mean maximum dimension of size distribution, λ wavelength) corresponding to these measurements ranged from 1.2 to 11.0. Details can be found in Arnott et al. (1995). Figure 1

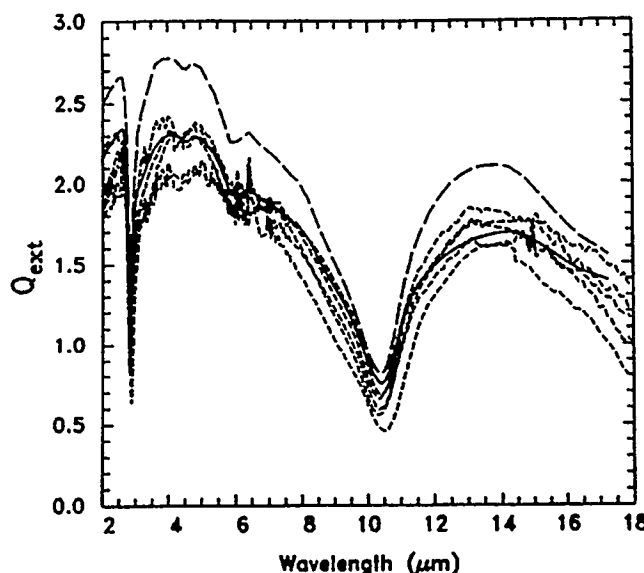


Figure 1. Extinction efficiency measurements (short-dashed curves) for an ice cloud of hexagonal columns, having a mean size ($\bar{D}$) of 7-8 μm . Using $\bar{D} = 8 \mu\text{m}$, the solid curve was predicted by the new cirrus radiation treatment, while the long-dashed curve was predicted from Mie theory.

gives the results of this study for an ice cloud of hexagonal columns. The sizes and concentrations of the ice crystals were measured, and the estimated $\bar{D}$ was 7 to 8 μm . The short-dashed curves depict the laboratory measurements. The solid curve was generated by the new ADT treatment, assuming $\bar{D} = 8 \mu\text{m}$. The minima shown are extremely sensitive to ice crystal size and only occur when sizes are relatively small and the real index of refraction, n_r , approaches 1. The agreement of the measurements with the ADT curve indicates the ice crystal path of the radiation was well represented (feature 1 above). The long-dashed curve was predicted by Mie theory, based on $\bar{D} = 8 \mu\text{m}$. Agreement between Mie theory and observations becomes poorer away from the minima where $n_r > 1$.

To apply Mie theory, column ice crystals were converted to equivalent distance (D_e) spheres, where the distance radiation travels through a crystal and corresponding ice sphere

sphere are equal. The size distribution projected area was conserved. These operations allow us to directly compare our treatment with Mie theory.

A similar comparison between lab measurements and theory was made when the ice cloud was composed primarily of hexagonal plates, and $\bar{D}$ was $14\text{ }\mu\text{m}$. Agreement between ADT and experiment in this case was equally as good as the case with columnar crystals.

Physical Explanation for the Behavior of Q_{ext}

The cause of the extinction minima near $2.8\text{ }\mu\text{m}$ and $10.3\text{ }\mu\text{m}$ is the result of the real part of the refractive index, n_r , approaching a value of 1.00. When $n_r = 1$, scattering due to refraction and reflection is minimized. Thus, the alteration of radiation transmitted through an ice crystal will be reduced, contributing less to extinction. Also, when $n_r = 1$, there is no phase lag for waves passing through the crystal, so there is no wave interference contribution to extinction. This leaves diffraction and absorption to contribute to Q_{ext} at the extinction minima. With just these two processes, $Q_{\text{ext}} \approx 2$ when absorption is strong and $D \gg \lambda$. But as the particle size becomes comparable to or smaller than the wavelength, the ability of the particle to scatter radiation via diffraction decreases, and absorption becomes the dominant contributor to Q_{ext} . As ice crystal size decreases, absorption changes from an area dependence to a mass dependence (i.e., Beer's Law), and the absorption efficiency, Q_{abs} , decreases. This is why the minima for Q_{ext} can be < 1 , even though the *imaginary* index of refraction is substantial at 2.8 and $10.3\text{ }\mu\text{m}$. This complex dependence of the extinction minima on the above phenomena puts theory to a rigorous test.

Testing the Cirrus Radiation Scheme with Field Measurements

This cirrus radiation treatment was compared against data from a cirrus cloud field study (Paltridge and Platt 1981), where broad-band albedos and emittances from a cirrus deck were measured from an aircraft. These measurements are shown in Figure 2 by the "+" signs. The solar zenith angle was 33 degrees. The mean ice crystal size, based on measurements from a 1-D Knollenberg cloud probe, was estimated as $\bar{D} = 6\text{ }\mu\text{m}$. This size was obtained by extrapolating the measured size distribution to zero size.

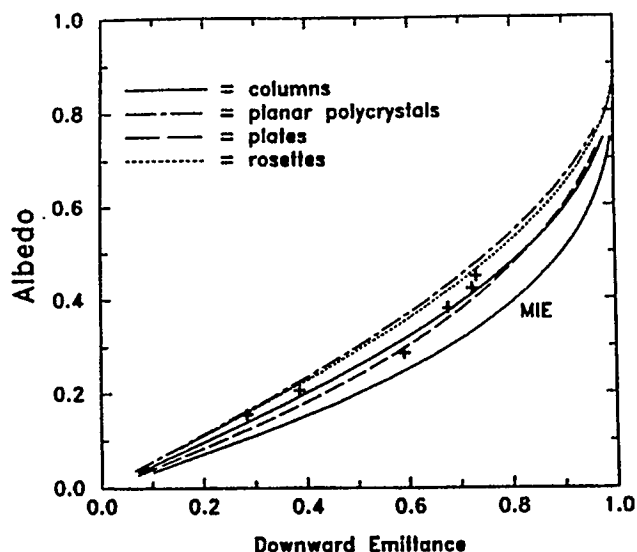


Figure 2. Albedo-emittance data (+) from the cirrus cloud sampled by Paltridge and Platt during Flight 7. Based on measurements, $\bar{D} \approx 6\text{ }\mu\text{m}$. This size was used to produce the theoretical curves. The lower curve for hexagonal columns is based on Mie theory.

Theoretical curves are shown for $\bar{D} = 6\text{ }\mu\text{m}$, for various ice crystal shapes. These were predicted from the new ADT treatment. This is the first time these observations have been explained theoretically from the observed microphysics.

Different Physics for Water and Ice?

A method was developed to make ADT "act" like Mie theory. This Mie modified ADT treatment is like the ADT treatment except that the effects of internal reflection/refraction and grazing photon capture have been parameterized to make ADT "act" like Mie theory. This parameterization generally matched the Mie theory result to within 10% for $x \geq 1$, where $x = \pi D/\lambda$. Grazing photon capture causes the absorption cross-section of an ice crystal to be larger than would be predicted from the physical cross-section. This results in greater emittances than evidently occur.

This parameterization was used to help interpret the above results. This modified ADT (short-dashed curve) is compared against Mie theory (solid curve) in Figure 3 for a size distribution of ice spheres having $\bar{D} = 15\text{ }\mu\text{m}$, where Q_{abs} is predicted over a wavelength range from $1.4\text{ }\mu\text{m}$ to $50\text{ }\mu\text{m}$.

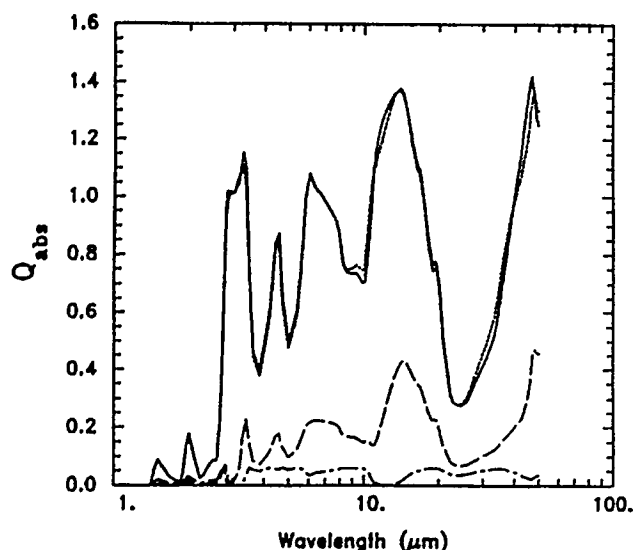


Figure 3. Absorption efficiencies predicted by Mie theory (solid curve) and modified ADT (short-dashed curve), for wavelengths from 1.4 μm to 50 μm and an ice sphere size distribution having $\bar{D}=15 \mu\text{m}$. The long dashed curve is the contribution from grazing photon capture, as predicted by the modified ADT, while the dash-dot curve gives the contribution from internal reflection/refraction.

Also shown is the contribution of grazing photon capture (long-dashed curve) and internal reflection/refraction (dash-dot curve) to Q_{abs} . Grazing photon capture is the dominant of the two processes in the thermal infrared and contributes significantly to Q_{abs} .

The capture of grazing photons contributes to wave resonance phenomena inside a sphere, as discussed in Guimaraes and Nussenzweig (1992). Resonance phenomena may be unique to spheres, although it may occur to a lesser extent in cylinders and some spheroids (Barber and Hill 1990). The above results indicate that grazing photons do not contribute to resonance phenomena in ice crystals, perhaps because the

different geometry of ice crystals does not support wave resonance. More laboratory measurements and analytical work are needed to test this hypothesis.

Acknowledgment

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RRTM: A Rapid Radiative Transfer Model

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Introduction

A rapid radiative transfer model (RRTM) for the calculation of longwave clear-sky fluxes and cooling rates has been developed. The model, which uses the correlated-k method, is both accurate and computationally fast. The foundation for RRTM is the line-by-line radiative transfer model (LBLRTM) from which the relevant k-distributions are obtained. LBLRTM, which has been extensively validated against spectral observations (e.g., the high-resolution sounder and the Atmospheric Emitted Radiance Interferometer [Brown et al. 1995]), is used to validate the flux and cooling rate results from RRTM. Validations of RRTM's results have been performed for the tropical, midlatitude summer, and midlatitude winter atmospheres, as well as for the four Intercomparison of Radiation Codes in Climate Models (ICRCCM) (Ellingson and Fouquart 1991) cases from the Spectral Radiance Experiment (SPECTRE). Details of some of these validations are presented below. RRTM has the identical atmospheric input module as LBLRTM, facilitating intercomparisons with LBLRTM and application of the model at the Atmospheric Radiation Measurement Cloud and Radiation Testbed sites.

Optical Depths

RRTM divides the longwave spectral region into 16 bands chosen for their homogeneity of contributing species and radiative transfer properties. A list of these bands and their respective key species is presented in Table 1. The selection of the spectral regions is facilitated by consideration of the Plates of spectral cooling rate as a function of log pressure provided by Clough and Iacono (1995).

In RRTM, all the optical depths in a given spectral band are represented by a small set of characteristic values that are used to perform the radiative transfer. The procedure that determines these characteristic values begins with the creation of a k-distribution, which involves assigning to

Table 1. RRTM Bands.

Wavenumber Range (cm ⁻¹)	Key Species in RRTM	
	1050-96 mb	96-0.01 mb
10-250	H ₂ O	H ₂ O
250-500	H ₂ O	H ₂ O
500-630	H ₂ O, CO ₂	H ₂ O, CO ₂
630-700	H ₂ O, CO ₂	CO ₂ , O ₃
700-820	H ₂ O, CO ₂	CO ₂ , O ₃
820-980	H ₂ O	--
980-1080	H ₂ O, O ₃	O ₃
1080-1180	H ₂ O	O ₃
1180-1390	H ₂ O, CH ₄	CH ₄
1390-1480	H ₂ O	H ₂ O
1480-1800	H ₂ O	H ₂ O
1800-2080	H ₂ O, CO ₂	--
2080-2250	H ₂ O, N ₂ O	--
2250-2380	CO ₂	CO ₂
2380-2600	N ₂ O, CO ₂	--
2600-3000	H ₂ O, CH ₄	--

each absorption coefficient $k(\nu)$ a value g ($0 \leq g \leq 1$) that represents the fraction of the absorption coefficients in the band smaller than $k(\nu)$. This procedure arranges the k -values in ascending order and defines a mapping $\nu \rightarrow g$ that provides an important perspective in the development of the model (West et al. 1990). The function $k(g)$ is then divided into a small number of intervals, each having a limited range

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of $k(g)$ values, and a representative value $k(g_i)$ is obtained for each interval. Since the equations used to calculate radiances have no explicit dependence on wavenumber, $k(g_i)$ can be used to estimate the radiance for the entire interval. The resulting radiances, weighted by the sizes of their respective intervals, can then be summed to yield the total radiance for the spectral band. This method for calculating radiances for inhomogeneous atmospheres is referred to as the correlated- k method.

Each spectral band in the RRTM is broken into 16 intervals with half-gauss quadrature spacing modified such that 7 intervals lie between $g = 0.98$ and $g = 1.0$. This modified quadrature spacing is done to better represent the high values of $k(g)$ associated with the centers of the spectral lines. For RRTM we have chosen to calculate each $k(g_i)$ by averaging the $k(g)$ values in each interval so as to include the contribution from every k -value in the interval. The full set of k -values comes from LBLRTM, which uses the water vapor continuum model CKD_2.1 (Clough et al. 1989) and the 1992 HITRAN line parameter database (Rothman et al. 1992).

For an arbitrary atmosphere, $k(g_i)$ values are obtained by linear interpolating in log pressure and temperature stored values of $k(g_i)$. In bands with more than one key contributing species, an additional interpolation is performed in a parameter, η , that represents the relative (strength-weighted) abundance of the two most critical species, which greatly affects the mapping $v \rightarrow g$. This parameter, called the binary species parameter, is defined as

$$\eta = \frac{S_1 W_1}{S_1 W_1 + S_2 W_2}, \quad (1)$$

where S_i is the integrated line strength of species over the spectral band and W_i is the layer column amount of that species. For these overlap bands, values of $k(g_i)$ are stored for reference atmospheres with values of η equal to 0, 1/8, 2/8, ..., 1 for a given pressure and temperature in the lower atmosphere and for values of η equal to 0, 1/4, 1/2, 3/4, and 1 in the upper atmosphere. For certain bands the species considered most critical are changed with altitude regime (see Table 1). In all bands, reference $k(g_i)$ values are stored for 59 pressure levels from 1050 - 0.01 mb, with a fixed ratio of ~ 1.2 between successive pressure values. At each reference pressure level, $k(g_i)$ values are stored at 5 temperatures T_{ref} , $T_{ref} \pm 15K$, $T_{ref} \pm 30K$, where T_{ref} is

obtained from the midlatitude summer temperature profile. The $k(g_i)$ values that are the result of this interpolation procedure include the spectral line contributions from the major species and the water vapor foreign continuum.

The contribution of the water vapor self-continuum to the optical depth is included by a separate procedure due to its quadratic dependence on water vapor abundance. First, the mapping $v \rightarrow g$ for all absorbers in each band is determined for two midlatitude summer atmospheric layers at 296 K and 260 K. The two mappings are then applied to the self-continuum absorption coefficients in the band at the respective temperatures. This procedure yields, at each temperature, a set of average self-continuum absorption coefficients $k_s(g_i)$, which are used to obtain the values of $k_s(g_i)$ at any temperature by appropriately interpolating. Finally, the contribution of the self-continuum to the optical depth is computed by multiplying $k_s(g_i)$ by a scaling factor that is quadratic in water vapor density.

Radiative Transfer Algorithm

RRTM performs a separate radiative transfer through the specified altitude range for each interval in each band. This, in effect, treats each interval, which "represents" a large number of frequencies in the band, in the same manner as a single spectral point is treated in a line-by-line model. As does LBLRTM, RRTM makes use of the "linear in τ " approximation for vertically inhomogeneous atmospheres and uses a Pade approximation to calculate an effective Planck function for each layer (applicable for any τ in the layer) by creating an appropriately weighted combination of the Planck function at the layer boundary temperature and the Planck function at the mean layer temperature (Clough et al. 1992). However, for a model using the correlated- k method, it is non-trivial to determine the appropriate (layer and level) Planck functions for any interval in a band since the Planck functions evaluated at all frequencies that are associated with that interval must be taken into account. Moreover, the mapping $v \rightarrow g$ can vary dramatically with atmospheric properties, especially the abundance of the band's key species, thereby limiting the accuracy of approximating the Planck function for the interval by the value of the Planck function at a single "dominant" frequency. This issue is addressed in RRTM by a two-part procedure. First, the integrated Planck energy for a band is a function only of temperature and is easily calculated and stored. The fraction of this energy that is due to the frequencies associated with an interval, defined by

$$f_g \equiv \frac{B_g W_g}{\frac{1}{v_2 - v_1} \int_{v_1}^{v_2} B_v(T) dv}, \quad (2)$$

where B_g is the average Planck function of the frequencies in the interval, W_g is the weight for the interval, and v_1 and v_2 are the frequency boundaries of the band, depends primarily on the mapping $v \rightarrow g$, which, in turn, is mainly dependent on the abundance of the band's key species. Therefore, for bands with one key species, f_g is considered constant. For bands with two key species, f_g is treated as a function of η and is computed by linear interpolation from stored f_g values calculated for reference values of η . All calculations performed to generate reference f_g values for a band are done at a pressure level appropriate for that band and use a midlatitude summer atmospheric profile. At execution time, RRTM computes the Planck function for an interval by multiplying the calculated f_g by the integrated Planck function for the band. The use of this procedure to calculate the Planck function assures that the total Planck energy is correct at every level in each band. The radiances calculated by the methods described above are then integrated (which is approximated by a first-moment Gaussian quadrature) over a hemisphere to yield fluxes.

Timing

Results of preliminary timing tests for RRTM indicate that computing the fluxes and cooling rates for a 51-layer atmosphere, which includes the performance of 256 (16 intervals $\times$ 16 bands) upward and downward radiative transfer calculations and, therefore, the computation of 256 optical depths per layer, takes 0.05 s on a SPARCserver 1000. This compares favorably to other rapid radiative transfer models. A further indication of the computational efficiency of the model is that these radiative transfer operations in RRTM take 1.8 times the amount of time needed to perform $51 \times 16 \times 16$ exponentials. Further evaluation of RRTM's timing is in progress.

Validations

The results of flux and cooling rate calculations performed by RRTM have been compared to those done by the LBLRTM for the midlatitude summer, midlatitude winter, and tropical atmospheres. Figures 1-4 show the results of these validations for the midlatitude summer atmosphere for the spectral range 10-980 cm^{-1} . RRTM and LBLRTM upward flux calculations at all altitudes differ by no more than 0.25 W/m^2 (Figure 2). For downward fluxes, errors of

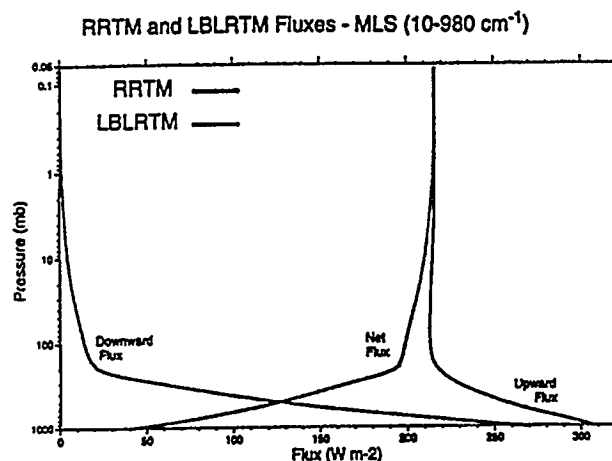


Figure 1. Fluxes as a function of log pressure from the LBLRTM and from the RRTM.

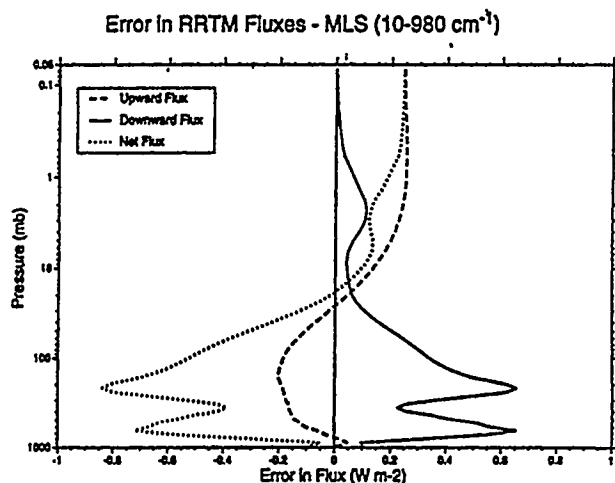


Figure 2. Difference between RRTM and LBLRTM fluxes as a function of log pressure.

a few percent for the entire spectral range are typical in the upper atmosphere, with an error at the tropopause of 0.4 W/m^2 . The errors in downward flux in the lower atmosphere are smaller, including an error in surface downward flux of 0.1 W/m^2 . The errors in the heating rate calculations (Figure 4) also are smaller near the surface, with typical errors in the 1000-100 mb range of less than 0.05 K/day. Errors in cooling rates between 1.0 and 0.1 mb can be as high as 0.5 K/day. Similar results to these have been found for the midlatitude winter and tropical atmospheres. In addition, very satisfactory results are obtained relative to both LBLRTM and observations for the calculation of downward surface radiance for the four SPECTRE atmospheres used in ICRCCM.

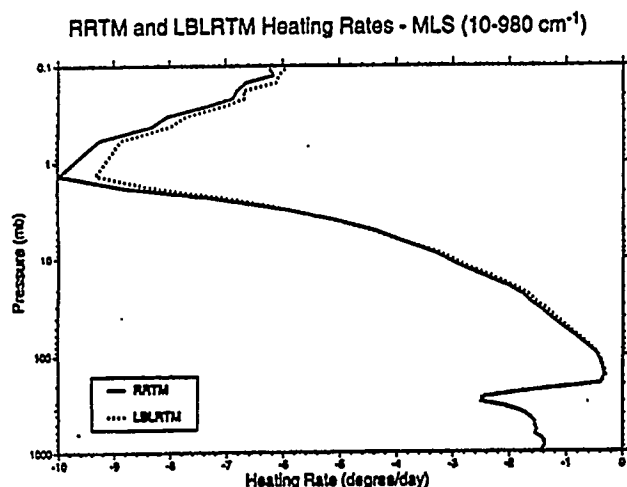


Figure 3. Heating rates as a function of log pressure from the LBLRTM and from the RRTM.

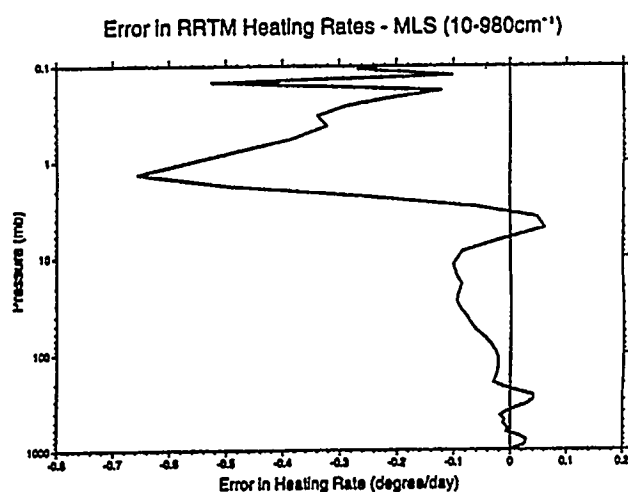


Figure 4. Difference between RRTM and LBLRTM heating rates as a function of log pressure.

Summary

RRTM currently provides an accurate and rapid method for computing longwave fluxes and cooling rates for an arbitrary clear atmosphere. RRTM, due to its use of linear interpolation to determine all necessary physical quantities, does not suffer from discreteness problems, enabling the model to accommodate microlayering.

The next phase in the development of RRTM will involve the extension of the algorithm to include the shortwave region of the spectrum and will allow the calculation of fluxes and cooling rates for both clear and cloudy skies.

Acknowledgments

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Instrument Development for Atmospheric Radiation Measurement (ARM): Status of the Atmospheric Emitted Radiance Interferometer-Extended Resolution (AERI-X), the Solar Radiance Transmission Interferometer (SORTI), and the Absolute Solar Transmission Interferometer (ASTI)

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Three instruments are currently under development for Atmospheric Radiation Measurement (ARM) at the University of Denver. The AERI-X (Atmospheric Emitted Radiance Interferometer - Extended Resolution) and the SORTI (Solar Radiance Transmission Interferometer) are joint projects with the University of Wisconsin (Hank Revercomb). The prototype SORTI has been operational at the Southern Great Plains (SGP) Cloud and Radiation Testbed (CART) site since January 1994. An AERI-X pre-prototype was installed in Eureka, Northwest Territories (NWT), Canada, in September 1994. The ARM AERI-X prototype has been running in Denver since December and will be tested at the SGP CART during the April intensive observation period (IOP). A demonstration ASTI (Absolute Solar Transmission Interferometer) has collected data from Denver and a prototype for ARM is under construction.

The SORTI prototype was installed at the SGP CART site in November 1993 and has been routinely operated by site staff since January 1994. Approximately 1000 spectra, occupying 2.5 Gbytes has been accumulated and stored on optical disk (the data are available from the authors). Figure 1 shows an example of a small region of the spectra obtained with the SORTI prototype, along with line-by-line, layer-by-layer simulations. Although the basic agreement is excellent, significant differences remain. For example the water vapor line at about 1106.7 cm^{-1} does not have correct wings, even though the simulation was done using the closest radiosonde ascent. At the level of fine detail, the ozone lines do not have quite the right depth, because a generic ozone profile has been used.

Two AERI-X prototypes have been constructed. These instruments use a different interferometer than the standard AERI, but the same calibration fixtures. The second instrument was built for the National Science Foundation and was

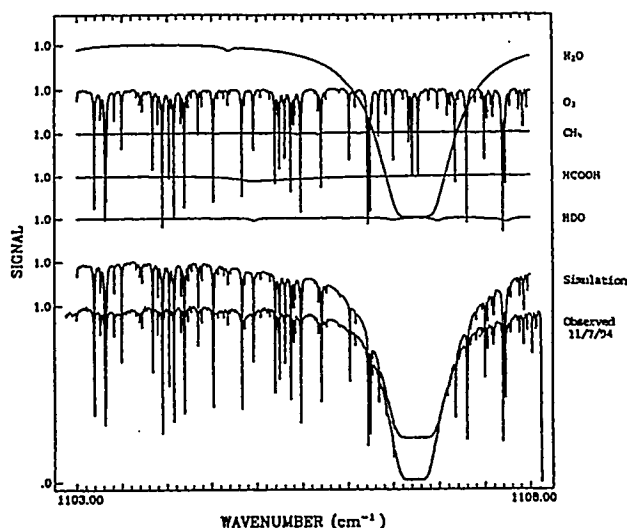


Figure 1. Line-by-line, layer-by-layer simulation of prototype SORTI spectrum from November 7, 1994, at the SGP CART site. The simulation used the closest radiosonde for temperature and water vapor, and a generic mid-latitude atmosphere for all of the constituents. The observed and calculated spectra have been normalized to 1 at the largest value in the frame. The contributions of the most significant absorbers in this region are shown in the upper section. Note that the strong water line near 1106.7 cm^{-1} does not match very well. Some of the smaller features do not have the right depth, because a generic atmosphere has been used.

finished first. It was installed at the Canadian Atmospheric Environment Service NDSC station at Eureka, Northwest Territories (80°N , 86°W) last September. Figure 2 illustrates the difference between a mid-latitude location

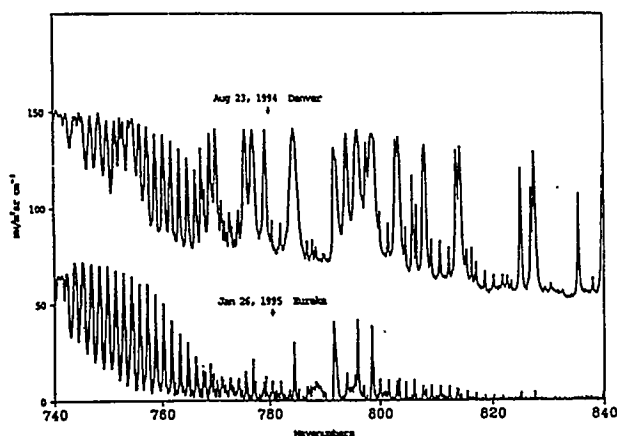


Figure 2. Atmospheric emission spectra obtained with an AERI-X prototype. Both are 30-minute observing sequences, with a 30° elevation angle. Spectra are on the same scale.

(Denver) in the summer and an arctic location in winter. The largest difference is in the amount of water vapor, while the temperature difference shows as well, particularly on the edge of the CO₂ band near 750 cm⁻¹. The ARM prototype will be tested at the SGP CART site during the April IOP.

The ASTI development instrument has successfully produced radiometrically calibrated spectra of the solar

spectrum arriving at the ground in Denver. An example of these spectra from a noon run on March 14, 1995, is shown in Figure 3.

An ASTI prototype is under construction. The prototype will have automatic filter changing and calibration, among other improvements.

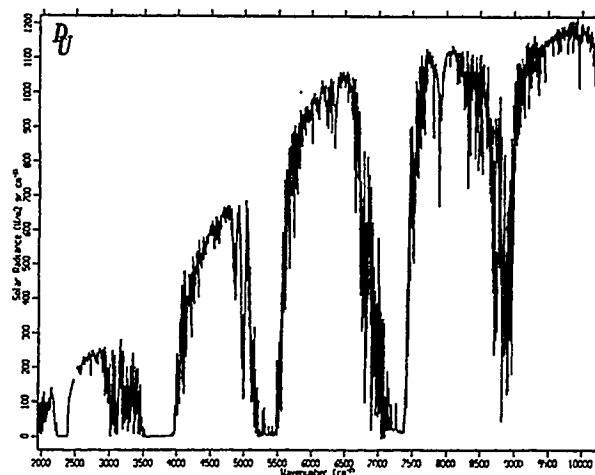


Figure 3. Solar spectrum on March 14, 1995. Resolution 4 cm⁻¹; approximate airmass 1.35.

Fractional Integration and Radiative Transfer in a Multifractal Atmosphere

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Recently, Cess et al. (1995) and Ramathan et al. (1995) cited observations which exhibit an anomalous absorption of cloudy skies in comparison with the values predicted by usual models (homogeneous atmosphere) and which thus introduce large uncertainties for climate change assessments.

These observations raise questions concerning the way general circulation models (GCMs) have been tuned for decades, relying on classical methods, of both radiative transfer and dynamical modeling, limited to studying the radiation/dynamics relationship on an arbitrary scale (often considered as "characteristic"), although the interactions between clouds and radiations occur over a wide range of scales.

These observations also tend to demonstrate that homogeneous models are simply not relevant in relating the highly variable properties of clouds and radiation fields. However smoothed, the intensity of clouds' multi-scattered radiation fields reflect this extreme variability.

Byrne et al. (1995), in order to explain this anomalous absorption, proposed a simple model of broken clouds and measured an increase in the value of the photon mean free path in comparison with the value calculated for a homogeneous atmosphere. The modeled media is a layer filled only with a mixture of clouds and portions of clear skies. They argue that photons, diffused by a first cloud, can circulate horizontally in a "clear sky" region and be reflected in the opposite direction by another cloud and could thus be "trapped" between two clouds (Figure 1). This phenomenon increases significantly the mean free path of photons and thus increases the total absorption of the layer.

However, Byrne et al. (1995) admit that the extreme variability of the radiation fields can be best understood in a multifractal framework (see for instance the different

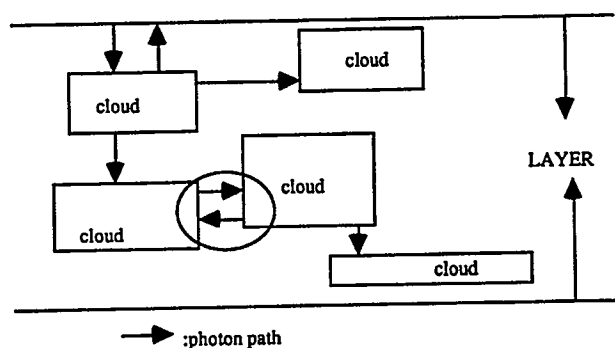


Figure 1. Illustration of the argument of Byrne et al. (1995): Photons are trapped between two or several clouds and thus it increases the global absorption.

contributions on this subject in Davis et al. 1990, Gabriel et al. 1990, Lovejoy et al. 1990, Schertzer and Lovejoy 1991). Indeed, the (scalar) multifractal model of cloud fields, as discussed below, respects the clouds texture, clustering, bands and intermittency, and the non-linear nature of the true dynamical processes at all scales. This model corresponds to a stochastic model of passive clouds, passively advected by a turbulent velocity field, using coupled cascade processes, and non-linearly conserving the fluxes of energy and concentration variance (Schertzer and Lovejoy 1987). A (scale invariant) multifractal field ϵ_λ is characterized by an infinite hierarchy of singularities γ with associated codimensions $c(\gamma)$; that is, at a scale ratio λ , the probability (Pr) of the fluctuations of the field diverging faster than λ^γ scales as $\lambda^{-c(\gamma)}$ (Schertzer and Lovejoy 1987, 1991).

$$\Pr(\epsilon_\lambda \geq \lambda^\gamma) \approx \lambda^{-c(\gamma)} \quad (1)$$

and the statistical moments scale as

$$\langle \epsilon_\lambda^q \rangle = \lambda^{K(q)} \quad (2)$$

where $K(q)\ln\lambda$ is the second Laplacian characteristic function (or cumulant generating function) of $\Gamma_\lambda = \ln \epsilon_\lambda$ which is the generator of the process.

The functions $c(\gamma)$ and $K(q)$ are related to each other, using a Legendre transform (Parisi and Frisch 1985) and for "universal multifractals" depend on only three parameters: α (the degree of multifractality), c_1 (the mean homogeneity), and H (the deviation from conservation).

The light propagates through this multifractal cloud density field which respects the Radiative Transfer Equation (Chandrasekhar 1950) and we try to exhibit the relationship between the singularities of the cloud density field and the radiative field with respect to the cloud density. The radiative transfer equation is highly nonlinear and it is not possible to find analytically a straightforward solution.

However it becomes rather simple when applied to a 1D medium. The variation of the radiative field is then proportional to the variation of the optical depth of the cloud (i.e., corresponds to a 1D-integration of the cloud density field). For a plane parallel medium the relationship appears to be a 1D-integration as well. We argue that this could be similar for more general media (i.e., which have inhomogeneities along all directions).

We want then to compare the Radiative Transfer Equation with a fractional integration and therefore we go back to a more general study and determine what happens to the statistics of a fractional-integrated multifractal field.

At first sight, a fractional integration of order H (i.e., a power filter, order H) should only shift the singularities of the fractional integrated field ($\lambda^\gamma \rightarrow \lambda^{\gamma-H}$).

However it appears that it introduces also a critical singularity below which it is not possible to get any information about the statistics.

We compared two fields, the first one being a multifractal field observed on a given dimension D and its D' -integrated counterpart ($D' < D$). We studied the relation between their respective moments through the function $K(q)$ and between their codimensions $c(\gamma)$. Then introducing the critical moment $q_{D'}$, due to divergence of high order moments, defined by $K(q_{D'}) = D' (q_{D'} - 1)$, we get

$$\begin{aligned} q < q_{D'} &\rightarrow K_{D'}(q) = 0 \\ q > q_{D'} &\rightarrow K_{D'}(q) = K_D(q) - (D - D')(q - 1) \end{aligned} \quad (3)$$

For codimensions we have

$$\begin{aligned} \gamma < \gamma_{D'} &\rightarrow c_{D'}(\gamma) = 0 \\ \gamma > \gamma_{D'} &\rightarrow c_{D'}(\gamma) = c_D(\gamma - D + D') - (D - D') \end{aligned} \quad (4)$$

In order to illustrate this point, we made 100 simulations of a multifractal field (Wilson et al. 1991; Pecknold et al. 1993) with $\alpha=1.35$ (measured for clouds by Tessier et al. 1993), $H=0$, and $C_1=0.75$. Then we made a fractional integration of order $D'=1$ for each realization, and we calculated the moments and codimensions. We represent these on the following graphs which appear to confirm the behavior predicted previously (Figures 2 and 3 where $D=2$ and $D'=1$).

Integration smoothes the low order singularities of a multifractal field, which means that it should be also the case for the Radiative Transfer Equation. Thus, the radiance field reflects only the high concentrations of water in atmosphere. However, the bad news is that we have to be very careful with analysis of these fields; they must be fractionally differentiated, otherwise the relevant range of singularities that exhibits the universal exponents C_1 , α , and H becomes rather small.

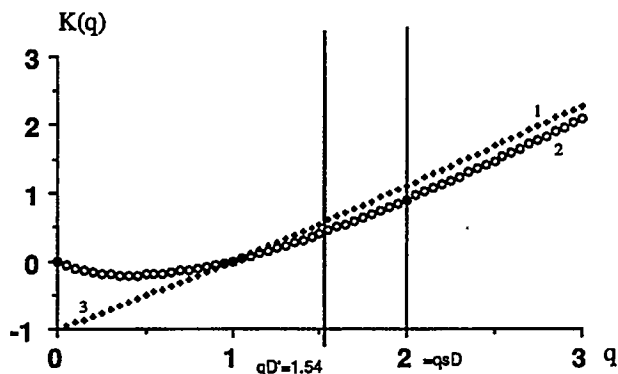


Figure 2. $K(q)$ for D' -integrated and dimension D fields with $\alpha = 1.35$ and $C_1 = 0.75$ ($H=0$) for $D=2$ and $D'=1$. Captions: + Integrated field on a two dimensional field $D' = 1$; o $D=2$.

Region 1: $K_{D'}(q) + q - 1 \approx -1.24 + 1.17q$

Region 2: $K(q) \approx -1.47 + 1.18q$

Region 3: $K_{D'}(q) + q - 1 \approx -1.01 + 0.99q$

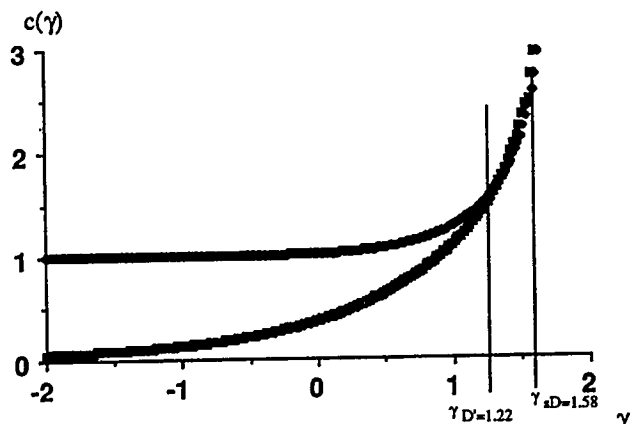


Figure 3. Codimensions of a D dimensional field and the corresponding D' -integrated field with $\alpha=1.35$ and $C_i=0.75$ ($H=0$) for $D=2$ and $D'=1$. Captions: ■ $D=2$; ♦ one dimensional integrated of previous, $D'=1$.

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Direct Shortwave Forcing of Climate by Anthropogenic Sulfate Aerosol: Sensitivity to Particle Size, Composition, and Relative Humidity

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Recent estimates of global- or hemispheric-average forcing of climate by anthropogenic sulfate aerosol due to scattering of shortwave radiation ("direct" effect) are uncertain by somewhat more than a factor of 2. The principal sources of this uncertainty are atmospheric chemistry properties (yield, residence time), and microphysical properties [scattering efficiency, upscatter fraction, and the dependence of these properties on particle size, composition, and relative humidity (RH)]. Among anthropogenic aerosols, much attention has been focused in sulfates because of their importance and relative ease in quantitative evaluation.

Charlson et al. (1992) and Penner et al. (1994) have estimated the magnitude of the direct forcing using a box model as a first approximation. This approximation is reasonable on a global average given that the forcing is linear in aerosol loading in the optically thin limit. The box model readily allows for the examination of uncertainties but cannot account for geographic distribution for which a chemical transport model is required (Charlson et al. 1991). The box model also does not take into account the wavelength dependence on α^* [in the present study, the scattering efficiency is evaluated relative to the mass of sulfate in the aerosol and is expressed as $\alpha^* = \alpha \times F(\text{RH})$] and Rayleigh scattering (Kiehl and Briegleb 1993). The area-averaged shortwave forcing by this model for a specific source strength of SO_2 (precursor of sulfate formed by atmospheric oxidation of SO_2) is

$$\Delta F_{\text{SO}_4^{2-}} = -0.5 F_T (1 - A_c) T^2 (1 - R_s)^2 \bar{\beta} \alpha_{\text{SO}_4^{2-}}^{\text{RH}_r} F(\text{RH}) \beta_{\text{SO}_4^{2-}} \quad (1)$$

The minus sign indicates that this is representative of a cooling effect and the 0.5 is due to the fact that only half the planet is illuminated at any given period of time. The symbols in the above equation are defined as follows:

F_T = the solar constant, the solar radiative flux at a mean distance from the earth to the sun

A_c = the fractional cloud cover in the area of concern. The factor $(1 - A_c)$ is introduced because the albedo enhancement is applicable only for cloud free regions.

T = the fraction of incident or scattered light transmitted through the atmosphere above the aerosol layer

R_s = the albedo of the underlying surface. The factor $(1 - R_s)^2$ takes into account multiple reflection between the surface and the aerosol layer.

$\bar{\beta}$ = the fraction of the radiation scattered upward by the aerosol, averaged over the sunward hemisphere.

$\alpha_{\text{SO}_4^{2-}}^{\text{RH}_r}$ = the light-scattering efficiency of sulfate aerosol, i.e., scattering coefficient per sulfate mass, at a reference low relative humidity ($\text{RH}_r = 30\%$)

$F(\text{RH})$ = the relative increase in scattering cross-section at an ambient RH

$\beta_{\text{SO}_4^{2-}}$ = the average SO_2 loading (sulfate column mass burden) that is a product of the source strength of anthropogenic SO_2 , the fractional yield of emitted SO_2 that reacts to produce sulfate aerosol, and the mean residence time of SO_2 aerosol in the atmosphere divided by the area of the geographical region to which the calculation is applied.

This paper examines the sensitivity of forcing to these microphysical properties for the purpose of obtaining a better understanding of the properties required to reduce the uncertainty in the forcing. To focus on the radiative terms in Equation 1 it is useful to define a normalized forcing for a cloud-free region as

$$\Delta G_{\text{SO}_4^{2-}} = \frac{\Delta F_{\text{SO}_4^{2-}}}{\beta_{\text{SO}_4^{2-}}} \quad (2)$$

which has units of W m^{-2} per $\text{g}(\text{SO}_4^{2-})\text{m}^{-2}$, or $\text{W g}(\text{SO}_4^{2-})^{-1}$. The relation (ΔG) between aerosol loading and forcing developed here is suitable for comparing modeled and measured aerosol forcing at specific locations and for use in climate models incorporating aerosol forcing provided aerosol microphysical properties are known or assumed.

The scattering efficiency of an ammonium sulfate aerosol at low RH (40%) varies by about a factor of 2 for particle radius 0.1 to 1 μm , with a maximum of $\approx 8 \text{ m}^2 \text{g}^{-1}$ (sulfate) for particle radius 0.25 μm (see Figure 1). The maximum particle mass scattering efficiency increases with increasing RH to as great as $70 \text{ m}^2 \text{g}^{-1}$ (Sulfate) at RH=97% because of accretion of water with increasing RH. The upscatter fraction likewise depends markedly on particle radius, with a hemispheric average in the range of 17% to 50%; upscatter fraction also depends strongly on solar zenith angle (SZA),

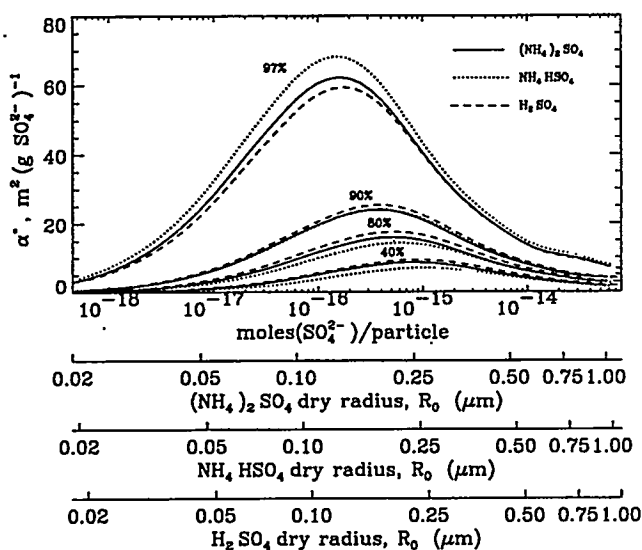


Figure 1. Dependence of scattering efficiency, integrated over the solar spectrum, for aqueous $(\text{NH}_4)_2\text{SO}_4$, NH_4HSO_4 , and H_2SO_4 aerosol on moles per particle of substance for indicated values of relative humidity. Particle dry "radius" (evaluated as the radius of the sphere of equal volume) is also shown as secondary axes.

ranging from a small fraction (10% to 30% depending on radius) near zenith to 50% at the limb. Based on the global and annual values given by Charlson et al. (1992), $\Delta G = -715 \text{ W g}^{-1}$ for a cloud free planet. Charlson et al. overestimate the normalized forcing by a factor of two (see Figure 2). This factor comes from the fact that a constant $\alpha \times F(\text{RH}) = 8.5 \text{ m}^2 \text{g}^{-1}$ over the solar spectrum is used. This average value is chosen based on the assumption that Rayleigh scattering is constant throughout the solar spectrum. However, Rayleigh scattering reduces ΔG more at shorter wavelengths. Thus, the average value should actually be reduced by a factor of about two, placing it in better agreement with the present study. Differences between the results from Kiehl and Briegleb (1993) and this study are primarily due to the fact that although they use the European Centre for Medium-Range Weather Forecasts RH to compute α^* , they obtain the optical properties of anthropogenic aerosol from the dry particle size distribution that ignores the dependence of RH.

The role of RH is of major importance, especially because of its influence on the scattering efficiency. Seasonal and latitudinal variations in $\langle \beta \rangle$ can influence ΔG by as much as a factor of two. However, influences by different particle compositions such as ammonium sulfate, ammonium bisulfate, and sulfuric acid appear to play a more secondary role on the changes in the magnitude of ΔG .

The normalized forcing, ΔG , is a useful quantity for the purpose of focusing solely on the radiative terms that are needed to compute the radiative forcing. ΔG computed as a

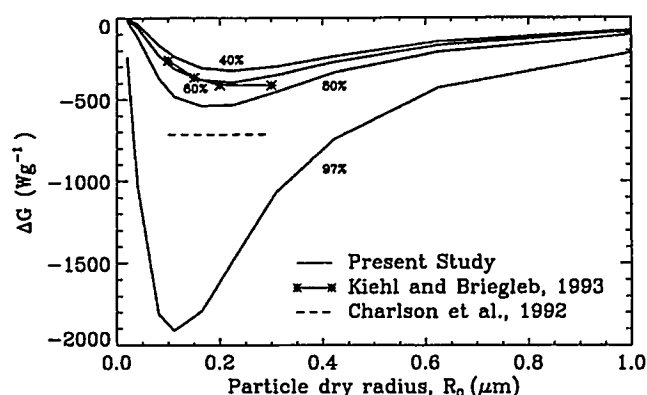


Figure 2. Normalized ammonium sulfate aerosol forcing for a hypothetical clear-sky planet as a function of particle dry radius for several RH.

function of particle size makes it possible to apply real particle size distributions for the purpose of determining the forcing as a function of variables such as RH. By isolating the influences of the microphysical properties of the aerosol particles from factors such as cloud cover and column mass burden, ΔG can be used in realistic climate models that use modeled or observed RH.

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Development and Testing of an Aerosol-Stratus Cloud Parameterization Scheme for Middle and High Latitudes

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Introduction

The aim of this new project is to develop an aerosol/cloud microphysics parameterization of mixed-phase stratus and boundary layer clouds. Our approach is to create, test, and implement a bulk-microphysics/aerosol model using data from Atmospheric Radiation Measurement (ARM) Cloud and Radiation Testbed (CART) sites and large-eddy simulation (LES) explicit bin-resolving aerosol/microphysics models. The primary objectives of this work are twofold. First, we need the prediction of number concentrations of activated aerosol which are transferred to the droplet spectrum, so that the aerosol population directly affects the cloud formation and microphysics. Second, we plan to couple the aerosol model to the gas and aqueous-chemistry module that will drive the aerosol formation and growth.

We begin by exploring the feasibility of performing cloud-resolving simulations of Arctic stratus clouds over the North Slope CART site. These simulations using Colorado State University's regional atmospheric modeling system (RAMS) will be useful in designing the structure of the cloud-resolving model and in interpreting data acquired at the North Slope site.

The RAMS Model

RAMS is a completely new code that was derived from the merger of two modeling programs: the cloud modeling program under the direction of William R. Cotton and the mesoscale modeling program directed by Roger A. Pielke (Pielke et al. 1992). RAMS can be run either hydrostatically or non-hydrostatically and contains an interactive grid-nesting capability allowing a large number of

telescoping grids or a number of simultaneous non-telescoping grids. Furthermore, RAMS is designed to be initialized from the objective analysis of standard atmospheric soundings and be driven by time-varying boundary conditions provided by standard 12-hourly soundings or from larger scale model grid point data such as the National Meteorological Center Nested Grid Model. This mechanism of data assimilation could be readily modified to allow interfacing with general circulation model grid-point data. Particularly useful features of RAMS for examining aerosol chemistry and aerosol impacts on clouds on the regional-scale are its ability to explicitly represent clouds of all sorts ranging from cumulus clouds to stratus and cirrus clouds including their interactions with atmospheric radiative transfer. Additionally, RAMS has the ability to represent surface properties including soil/vegetation effects on the surface energy budget, sea ice, and snow-covered land surfaces.

Cloud Microphysics

A new microphysical scheme has been developed which allows two-moment prediction of the distribution function for each hydrometeor category, making the determination of each hydrometeor size spectra less arbitrary (Meyers 1995). This scheme predicts the mixing ratio and number concentration of rain, pristine ice crystals, snow, aggregates, graupel, and hail categories. The cloud microphysical scheme includes the use of a generalized gamma size-spectrum, for all hydrometeor categories (Walko et al. 1995). Some other relevant features of the new two-moment scheme include

- approximate solutions to the stochastic collection equation rather than the continuous-accretion model approach

- predictive equations for ice nuclei (IN)
- new heterogeneous and homogeneous ice nucleation parameterizations with analytical flux equations to predict mixing ratio and number concentration conversion from pristine ice crystals to snow due to deposition (no riming)
- diagnosed crystal habit dependent on temperature and saturation.

Model Development

RAMS has traditionally been used mainly for mesoscale simulations at middle latitudes, particularly over the lower 48 states. In order to perform simulations in the high latitude and arctic regions, several modifications of the standard RAMS version 3b are necessary. For example, the RAMS Isentropic Analysis package (data assimilation, objective analysis, and model initialization routines) has been extended to operate correctly in the polar region and across the International Date Line. The predictive model as well requires additional features to achieve the desired results.

Surface Parameterizations

An adequate representation sea ice is a very important requirement in any atmospheric model simulation of the marine arctic environment. Sea ice affects the longwave radiation in the winter months, while in the summer months the shortwave radiation budget is affected as well. Of more significance, the surface heat and moisture flux properties of sea ice and open water differ dramatically and can have a profound effect on the short- and long-term climate (Chapman and Walsh 1993). Furthermore, the southward extent of sea ice varies as much as 15°C seasonally (Walsh et al. 1993).

Monthly gridded data sets of northern hemisphere sea-ice percentage (Walsh and Johnson 1979) have recently become available through the National Center for Atmospheric Research and efforts are under way to incorporate this data into RAMS. The rather coarse nature of this data set (grid boxes of 111 km), and low temporal resolution (essentially "snapshots" for the last day of each month), may prove inadequate for verifiable simulations at the LES scale. Unfortunately, more resolved regional sea-ice data sets are not readily available.

Ice as a surface-cover type is already available in RAMS, and a snow-surface parameterization is currently in the development and testing stages.

New Microphysical Features

A major objective of this research project is to predict the effects of changes in the aerosol spectrum upon evolving cloud properties, for both cloud condensation nuclei and IN. Therefore, schemes for representing the activation of aerosol are an important focus. Since activation is sensitive to the predicted value of supersaturation, we use prognostic equations for the concentrations and mixing ratios of cloud water, rainwater, and all ice hydrometeor categories.

Currently RAMS is being updated to include prognostic equations for two aerosol species: one a dry aerosol species represented by a single mode, and the other a hygroscopic aerosol species represented by up to three modes. These four modes will allow for representation of aerosol size and chemistry under both "clear" and arctic haze conditions.

Assuming that the aerosol is distributed in a log-normal or gamma distribution, the number concentrations and mixing ratios of these species are explicitly predicted. Advection, turbulent diffusion, and removal by nucleation are considered. Moreover, cloud droplets are assumed to follow a gamma distribution, with concentration and mixing ratio of cloud droplets being predicted as well.

Tests are currently under way to ensure that predicted vapor concentrations, along with predictions for numbers of aerosol activated and numbers of drops and crystals, are reasonable. Furthermore the scheme is designed to be used with the relatively long time steps typically used in mesoscale models. As discussed below, an explicit bin-resolving microphysical model will also be used in RAMS LES experiments to help determine the appropriateness of the treatment in the two-moment model.

In addition to accurately representing the supersaturation, and relating this to the numbers of aerosol particles activated, the moment treatment should reflect the removal of larger-sized particles by activation, so that new aerosol mass is not activated in subsequent time steps. For simpler schemes where breadth parameters are fixed, updated number and mass concentrations are reallocated into larger-sized particles, creating "tails" from which mass is continuously activated. We are exploring a number of schemes designed to mitigate this problem.

We are also developing methods to “track” the aerosol mass through the cloud spectra so that evaporating drops or crystals will release particles that become re-suspended (Richardson 1995), as some of these particles may be altered by physical and chemical processes within the cloud.

Synoptic-Scale Simulations

The synoptic-scale flow is an important factor in the formation of arctic clouds, both in the transport of heat and moisture and in the long-range transport of trace gases and aerosols (Iversen 1989). High latitude synoptic-scale disturbances such as extra-tropical cyclones and baroclinic waves produce broad regions of vertical motion enhancing the production of hydrometeors and perturbing the ambient aerosol spectrum. Therefore it is important that the meso-scale model accurately simulate the larger-scale dynamics and the transfer of energy and enstrophy between scales.

To validate the model physics in the arctic environment, we are performing relatively coarse-grid simulations (100- and 50-km Δx) of a major winter storm which occurred over the arctic slope of Alaska on February 28, 1989 (see discussion below). Some of the important issues to be addressed by this and other sensitivity studies are

- the vertical grid-spacing required to adequately resolve the extremely stable wintertime arctic boundary layer dynamics
- relative merits of data assimilation and objective analysis on isentropic surfaces versus σ_z (terrain-following) surfaces
- the impact of long-range aerosol transport on arctic haze production and its consequent feedback on the radiation budget
- the effect of model domain height on the evolution of synoptic-scale dynamics.

Large Eddy Resolving Simulations

Parallel to the model development and testing at the large scale, LES experiments are planned to test the microphysical schemes.

To refine and test the two-moment, multiple-mode micro-physics scheme, LES experiments will be run with explicit bin-resolving microphysics, applied to the aerosols

(Richardson 1995) and cloud droplets (Feingold et al. 1994). The aerosol scheme uses assumed particle chemical composition for the activation process as well as for determining haze size under various relative humidities. The explicit-microphysics LES runs employ small time steps with accurate representation of fluctuations in supersaturation to capture the details of the aerosol/cloud interactions. We will also extend the bin model to include ice-phase hydrometeors (Reisin 1995).

A perhaps more rigorous test of the cloud model will be performed by implementing a “floating LES” within a mesoscale flow field. With current computer resources, it is not feasible to perform simulations with nested grids ranging from the synoptic-scale (~ 100 km Δx) down to the LES scale (~ 50 -m Δx). To circumvent this restriction, we intend to perform LES simulations with time-varying boundary conditions, such boundary conditions being determined by the output of a multiple nested-grid mesoscale simulation. In this manner we may evaluate the time-dependent effects of long-range transport of aerosols in a horizontally inhomogeneous LES domain “advecting” in the mean mesoscale flow.

North Slope Case Study (February 28, 1989)

As an example of a mesoscale weather system simulation in the arctic, we here briefly describe a single grid RAMS simulation of a heavily precipitating winter storm occurring on February 28, 1989. This storm produced over 2 cm of precipitation (liquid-equivalent) at Point Barrow on Alaska’s North Slope, more than 10 times the typical observed total February precipitation for this location. This rather unusual event occurred under a strong lower-and mid-tropospheric ridge of high pressure.

At the model initialization time of 0000 UTC February 28, a tongue of strong southerly surface winds extended from 50°N northward over the Bering Sea and through the Bering Strait. This low-level jet was advecting high θ_e air northward over a quasi-stationary polar front oriented zonally at about 75°N over the Arctic Ocean (Figure 1a). The resultant over-running produced a region of moderate to heavy snowfall south of the front, extending to the northern coastal areas of Alaska. Additionally, the moist flow over the Brooks Range in northern continental Alaska had a significant upslope component in the lowest 5 km of the atmosphere.

In less than 1 hour of simulation time, precipitation was occurring over the northern coast and Arctic Ocean. Figure 1b shows the cloud at 3000 m above sea level at 0800 UTC, after 8 hours of simulation time. Figure 2 shows vertical cross-sections of several ice mixing ratio fields at this time. The precipitation appears to be forced by both topographical and mesoscale baroclinic mechanisms. The pristine ice and snow mixing ratio maxima are both located north of the Brooks Range with peak values (0.01 and 0.02 g kg^{-1} , respectively) confined primarily above 3 km .

Aggregation is quite active in this region as shown by the aggregate mixing ratio field (Figure 2d). The aggregate field extends to lower altitudes with peak values (0.11 g kg^{-1}) much greater than the pristine ice and snow categories. Graupel is also simulated (not shown) in this region, indicating the presence of some supercooled cloud water north of the Brooks Range.

The results from this sensitivity simulation are encouraging in that they demonstrate the ability of the standard RAMS single-moment microphysics to perform credibly in the Arctic.

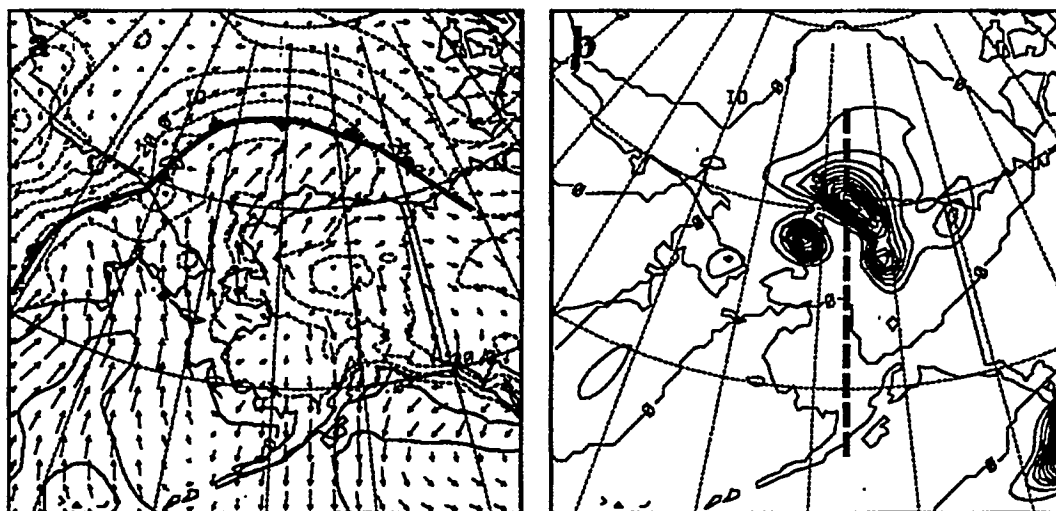


Figure 1. a) The surface winds, temperature (5°C contours) and frontal position for Feb. 28, 1989, at 0000 UTC, the initial time of the simulation. The strong southerly surface jet over the Bering Sea and Bering Strait with winds exceeding 25 ms^{-1} is providing significant positive low-level θ_e advection and upslope conditions for the Brooks Range in N. Alaska. b) Total ice mixing ratio at 0800 UTC and 3000 m above sea level, after 8 hours of simulation. The contour interval is 0.01 g kg^{-1} and the maximum value is 0.3 g kg^{-1} . The heavy dashed line shows the location of the N-S vertical cross sections shown in Figure 2.

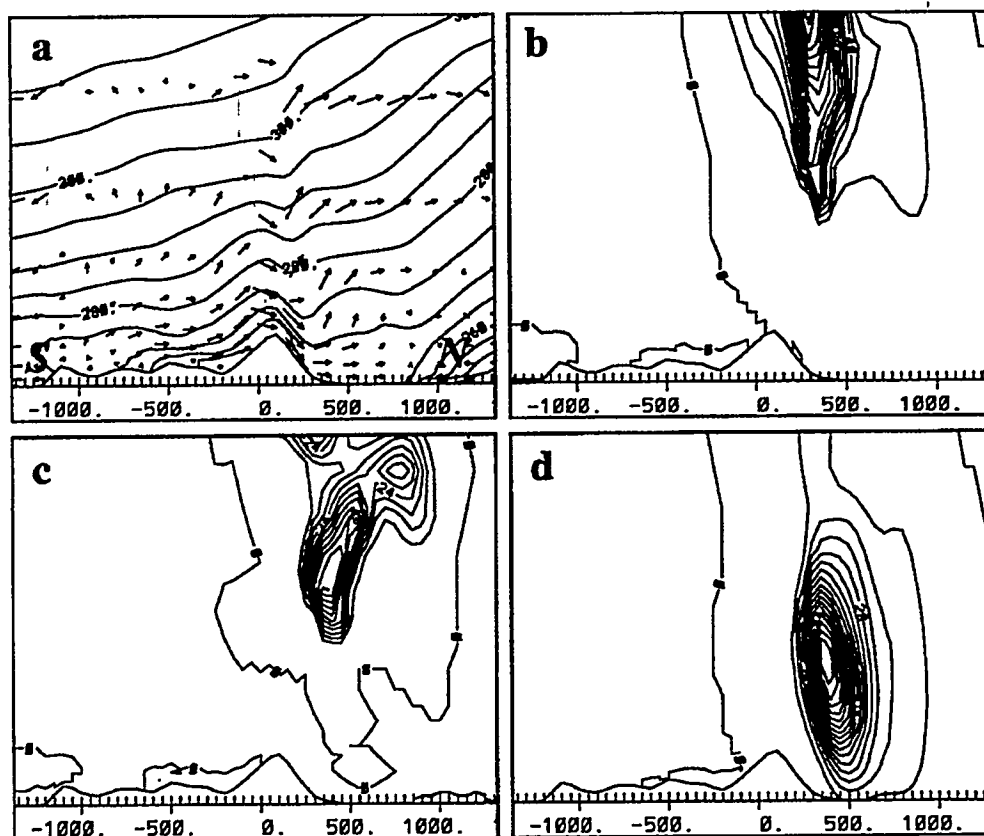


Figure 2. All cross-sections at time 0800 UTC, located along dashed line in Figure 1b. Figures are oriented with south to the left as denoted in a. Only the lowest 6 km of the domain is shown in the figures. a) Wind vectors and potential temperature contours (5°C increments), maximum winds are 32 m s^{-1} ; b) snow mixing ratio with a contour increment of 10^{-3} g kg^{-1} and a maximum value of 0.02 g kg^{-1} ; c) pristine ice mixing ratio, contour interval of $6 \times 10^{-4}\text{ g kg}^{-1}$; d) aggregate mixing ratio, contour interval of $7 \times 10^{-3}\text{ kg}^{-1}$ and a maximum of 0.11 g kg^{-1} .

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Airborne Studies of Submicron Aerosol in the Troposphere Over West Siberia

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Introduction

Submicron fraction particles that have the longest lifespan and are included in almost all atmospheric processes (Rosenberg et al. 1980; Rosenberg 1983) are of special importance among the great variety of sizes of particles present in the atmosphere. Submicron particles mainly determine the optical state of the atmosphere in the visible spectral range, essentially cause the absorption of infrared radiation (Rosenberg et al. 1980; Panchenko 1988) and, since they are the products and participants in all aerosol-to-gas transformations, accumulate a lot of various chemical compounds and transfer them to large distances.

Investigation of the processes of the spatial-temporal variability of aerosol particles for different climatic zones of the earth is the experimental base for studying their effect on climatically and ecologically significant factors and estimating their unfavorable tendencies. The increasing anthropogenic loading of the earth's atmosphere is creating an urgency for aerosol research. Regardless of how perfect the analytical and numerical methods of solving radiation problems may be, success in forecasting climatic change is mainly determined by the reliability of the experimental data on optical parameters of the atmosphere and of the description of their variability under the effect of external factors.

Experiment

Complex airborne investigations carried out during a comprehensive geophysical experiment in the characteristic climatic zones of the northern hemisphere made it possible to obtain new information on the optical properties of atmospheric aerosol (Belan et al. 1995).

Taking into account the accumulated experience in investigating aerosol processes and using the characteristic lifetime of submicron aerosol particles in the troposphere as a guide, analyzing the factors that determine the variability of aerosol characteristic in the regional scale should be our primary

concern. We chose West Siberia as the region for investigations because we performed monitoring observations there from onboard an aircraft-laboratory (total amount of the data obtained is 602 vertical profiles of aerosol scattering coefficient) during 1986-1988. The applied instrumentation and measurement technique is described by Panchenko et al. (1994).

The most significant examination of the bulk of data obtained was based on the comparison of meteorological parameters and synoptic attributes observed during the experiment with analogous long-term data obtained by the Russian Meteorological Network (Guterman 1980). Average values of meteorological parameters (temperature and specific humidity of air) in our bulk of data are close to the average climatic data for our region (Panchenko et al. 1994). This fact provides reason enough to analyze in detail the hierarchy of external processes determining the transformation of the scattering coefficient stratification $\sigma(H)$ and the variations of aerosol content in the lower troposphere. The analysis was carried out in order to develop a model that would adequately take into account the role of any factor in general variability of aerosol characteristics.

Results and Discussion

The monthly average values of the optical thickness τ ($\lambda = 0.52 \mu\text{m}$) and isolines of the scattering coefficient at the same wavelength are shown in Figure 1.

It follows from the analysis of these data that near the ground the greatest aerosol content is observed in winter and is caused by intensification of anthropogenic aerosol sources and by the fact that the inverse temperature profile in the near-ground atmospheric layer prevents penetration of aerosol emission into upper layers of the atmosphere.

The elevated aerosol content in spring at heights above 2.5 km is connected with the movement of the snow melting boundary and is caused by the aerosol coming to the atmosphere of West Siberia from the remote sources under the effect of the westward transfer.

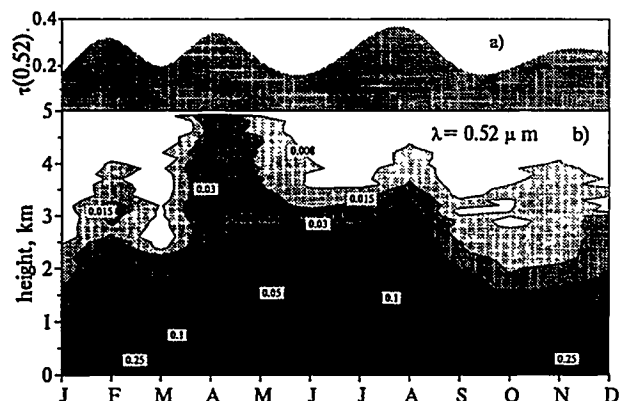


Figure 1. Annual behavior of optical thickness (a) and isolines of aerosol scattering coefficient, km^{-1} (b).

It is noted that in the majority of realizations the vertical profile $\sigma(H)$ has the layer structure, and the layers are situated near the levels where the alternation of stable and nonstable microareas ($\Delta H \sim 100$ m) is observed in the temperature stratification.

Season peculiarities of formation of σ are caused mainly by the temperature regime of the atmosphere, and one can use the three-layer concept of aerosol height distribution as the first approximation for their description. Within the frameworks of this approach, one can select the following height ranges: up to H_1 is the ground layer, H_2 is the height of the layer of intense turbulent exchange, above H_2 is the layer of free atmosphere (Figure 2).

Aerosol characteristics in different air masses coming to West Siberia in different seasons have reliable differences. In particular, we revealed the differences in vertical profiles of aerosol scattering coefficient in Arctic and polar air masses (Figure 3).

In the winter, above ~ 500 m, Arctic air contains, on average, more aerosol particles than observed in polar air. In the spring, above 2.5 km, there are observed differences in the Arctic air from the characteristic of the winter conditions, but in the lower 1-km layer the air becomes similar to the summer air. In the summer, Arctic air masses bring air to the region under investigation that is clearer in all height ranges in comparison with polar air masses.

The estimate of season peculiarities of aerosol accumulation in the Arctic atmosphere assessed from the data of airborne nephelometric measurements over West Siberia showed that

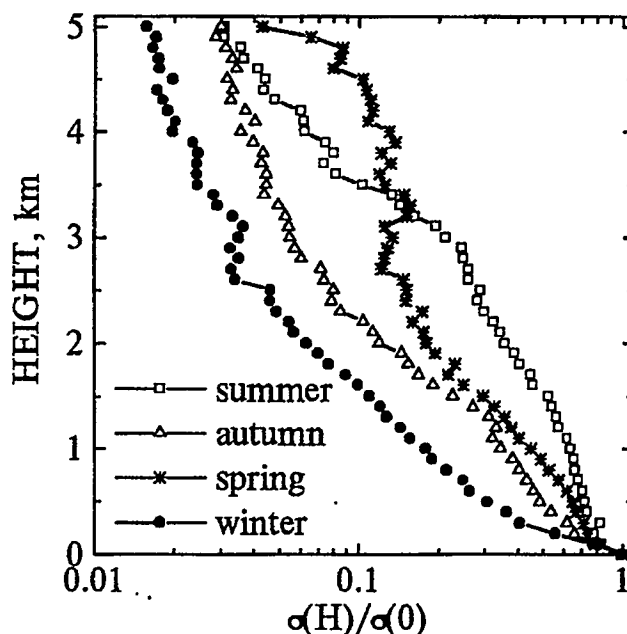


Figure 2. Normalized vertical profiles of aerosol scattering coefficient.

the estimated values of the mass content are close to the values that are observed immediately in the Arctic area. The analysis showed that in winter there is no daily behavior of the aerosol vertical profile above the ground layer. This makes it possible to take into account only variations of the relative humidity when modeling optical characteristics.

Principal features of daily transformation of aerosol vertical profile in the spring and autumn are similar to the summer daily behavior, but the amplitude of variations is much less. In summer, as a result of the heating of the underlying surface and the atmosphere during the day, the height of the mixing layer increases and the layers are filled by aerosol. In the evening, the aerosol emission from the near ground layer stops, and the lower atmospheric layers (below $H \sim 1.5$ km in average) start to empty. Above this layer up to $H \sim 3.5$ km, some increase of the height of the mixing layer continues.

Conclusion

The data obtained provide evidence of the fact that taking into account all these factors is favorable for increasing the accuracy of the developed regional models of the optical characteristics of atmospheric aerosol.

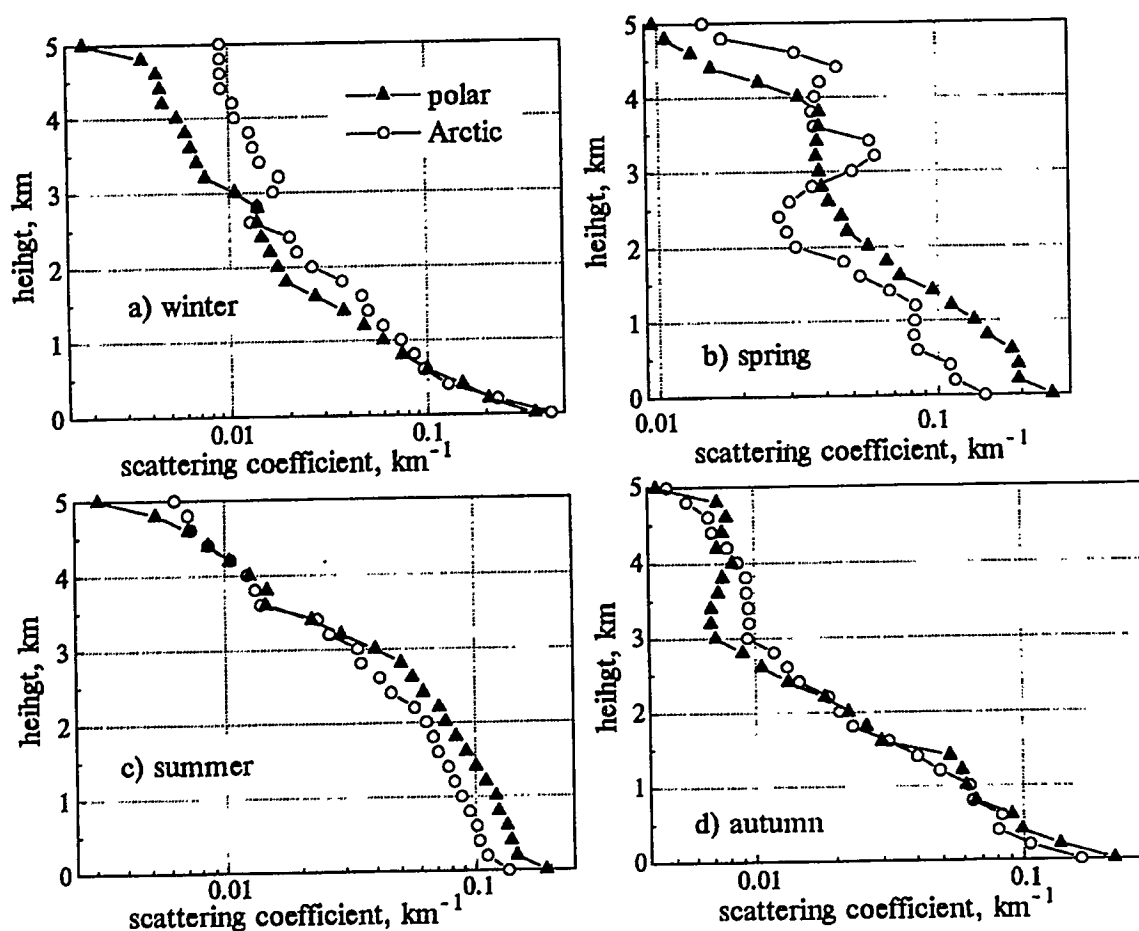
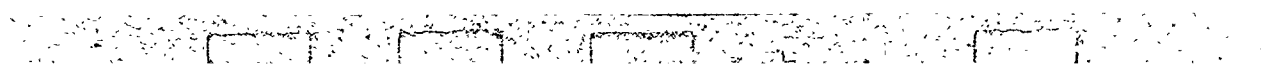


Figure 3. Vertical profiles of aerosol scattering coefficient in different air masses.

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The Optical Properties of Equatorial Cirrus in the Pilot Radiation Observation Experiment

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Introduction

The development of a sensitive filter radiometer for the Atmospheric Radiation Measurement (ARM) Program has been reported by Platt et al. (1994, 1995). The aim was to develop a reliable and fast instrument that could be used alongside a lidar to obtain near realtime optical properties of clouds, particularly high ice clouds, as they drifted over an ARM Cloud and Radiation Testbed (CART) site allowing calculation of the radiation divergence in the atmosphere over the site. Obtaining cloud optical properties by the lidar/radiometer, or LIRAD, method was described by Platt (1979) and Platt et al. (1987); the latter paper also describes a year's data on mid-latitude cirrus.

The optical properties of equatorial cirrus (i.e., cirrus within a few degrees of the equator) have hardly been studied at all. The same is true of tropical cirrus, although a few observations have been reported by Davis (1971) and Platt et al. (1984).

The ARM Pilot Radiation Observation Experiment (PROBE) took place in the southern summer of 1992-93 as part of the Tropical Ocean Global Atmosphere (TOGA) Coupled Ocean Atmospheric Response Experiment (COARE), and was sited at Kavieng, New Ireland, Papua New Guinea (2.5°S latitude, 152°E longitude). The CSIRO Division of Atmospheric Research (DAR) lidar and radiometer was operated at Kavieng for about three weeks in January-February 1993. During that time, several extensive cirrus clouds were studied, particularly during periods when there was little cumulus or altocumulus developing in the line of sight. This article describes these observations, the analysis method used, and a summary of the results.

Observations

Observations were made on cirrus clouds whenever the opportunity arose during daylight hours and on a few

evenings. Extensive cirrus layers were observed on January 24th, 26th, and 29th and February 3rd and 4th. Cirrus were also studied on some other days for brief periods.

The equipment consisted of the CSIRO three wavelength lidar, operating at 0.53 micron, and two radiometers, the CSIRO DAR MARK II radiometer detecting in a band centered at 10.84 micron wavelength and the ARM cooled HgCdTe radiometer detecting at three wavelengths centered at 8.62, 10.86, and 12.04 microns. At Kavieng, observations were made both at 8.62 and 10.86 microns. The CSIRO radiometer was used on January 24th and February 3rd and 4th when liquid nitrogen was unavailable. It was also operated simultaneously with the ARM radiometer on January 29th. A detailed comparison of the performances of these radiometers was given in Platt et al. (1995). The ARM radiometer was found to be markedly superior.

The lidar and radiometer were both operated vertically and both with fields of view of about 8 mrad, corresponding to a cloud area sampled of about 130 m at 16-km altitude, and 80 m at 10 km. The lidar fired a number of shots at a frequency of 10 hz at intervals of 30 seconds, the number of profiles varied depending on the cloud height and optical depth. The radiometers recorded continuously with time constants of 5 seconds for the CSIRO MARK II and 1 second for the ARM. The Lidar and radiometer clock times were compared closely.

Observations were made when there was cirrus above the site without intervening clouds. Lidar observations were obtained at times also on middle-level altocumulus as well as cumulus, but the large changes in dynamic range in backscatter caused problems and the lidar was frequently turned off when low clouds intervened. A radiance trace for one day is shown in Figure 1. Periods of cumulus in Figure 1 are clearly identifiable as an increase in radiance to about $4.7 \text{ W m}^{-2} \text{ sr}^{-1}$ at 10.84 microns for "black" cumulus.

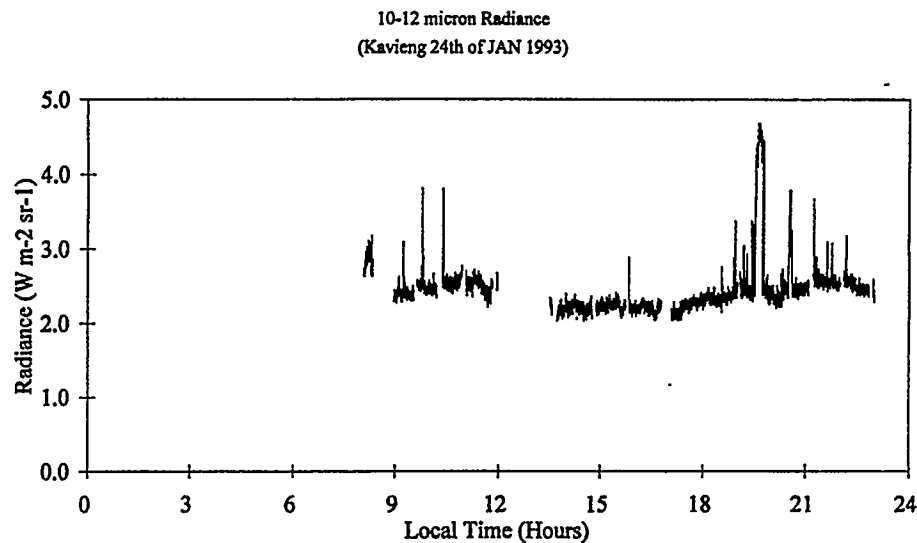


Figure 1. Radiance output for one day using the CSIRO radiometer. High peaks in the radiance represent cumulus episodes.

Calibration

In order to calculate the optical properties of the clouds accurately, both lidar and radiometers needed to be well calibrated. The lidar was calibrated against Rayleigh scatter at an altitude which was judged to be aerosol free by a method developed by Young (1995). Each lidar profile was obtained by averaging a sufficient number of shots to obtain a profile that had enough signal-to-noise ratio both below and above the cloud (when semitransparent). In these regions the range-corrected lidar signal was fitted to a reference molecular backscatter signal that was obtained from suitably interpolated radiosonde profiles. By fitting the lidar profile to the molecular profile above and below the cloud layer the cloud transmittance and hence the visible optical depth could also be calculated. As radiosonde data were only available below about 18 km, climatological or satellite data were used above this height. One difficulty with using this method with high clouds at Kavieng was the strong scattering from the Mt. Pinatubo volcanic aerosol layer above the clouds which made any fitting of the lidar profile to a molecular profile in this region invalid. A solution to this problem was found in these cases by replacing the molecular reference profile with a reference profile produced by averaging several thousand cloud-free lidar signals. Each profile was calibrated against the reference profile, thus allowing an assessment of data quality by studying the magnitude of the means and standard deviations in the fitting parameters.

The infrared radiometers were calibrated against liquid nitrogen, or ice water, when liquid N_2 was not available. Difficulties were experienced in the accurate calibration of the CSIRO radiometer which had to be operated outside the sea-container laboratory. Fluctuating ambient temperatures caused large drifts in the electronics gain and draughts caused the build-up of condensed water or ice above the liquid N_2 , thus elevating the measured radiance. However, careful and frequent calibrations reduced uncertainties to a controllable level. The incoming signals were chopped against two reference blackbodies maintained at 40°C that filled half the apertures, a design common to both radiometers. The radiometers were calibrated against an external blackbody at temperatures around 40°C to check at what temperature the output signal fell to zero in practice. Because of small misalignments, this temperature was usually not exactly 40°C . However, knowledge of the exact "cross-over" temperature was sufficient for a complete calibration. Subsequent closer examination of calibration curves revealed some non-linearities in the ARM radiometer, but only at high calibration temperatures. These were taken into account in the calibrations.

Theory

The lidar equation can be written

$$\gamma(\pi) = (k/2\eta) \left[1 - \exp \left(-(2\alpha\eta) \log \left(\frac{1}{1-\epsilon} \right) \right) \right] \quad (1)$$

where k is the backscatter to extinction ratio, η is a multiple scattering factor that effectively reduces the extinction, ϵ is the infrared emittance, and α is the ratio between visible lidar extinction and infrared absorption optical depths. It can be seen that when $\epsilon = 1$, $\gamma(\pi) = (k/2\eta)$

The infrared emittance can be related to the infrared absorption optical depth δ_a by the relation

$$\epsilon = 1 - \exp(-\delta_a) \quad (2)$$

where scattering terms are neglected.

Analysis

The basis of the analysis of the data was the method reported by Platt et al. (1987)-the standard lidar/radiometer (LIRAD) technique. However, one refinement was the use of water vapor data measured by the microwave radiometer at the Environmental Technology Laboratory of the National Oceanic and Atmospheric Administration which supplied a continuous output of the water vapor path over the site. Profiles of temperature and humidity were also available every six hours from radiosonde data. These data allowed a continuous update of the water vapor emission and transmittance of radiation to the radiometer, instead of relying on the simple linear interpolation between morning and evening radiosondes that had been used in previous experiments. The water vapor transmittance was obtained as follows. The water vapor radiances were calculated from the radiosonde data for the whole period using a radiative transfer equation for a range of water vapor continuum absorption coefficients, allowing also for water vapor line absorption (Platt et al. 1984). The radiances were then compared with clear-sky radiances observed at similar times. The transmittance was calculated from the radiance that was closest to the observed radiance. Apparent changes in water vapor continuum absorption were attributed to changes in aerosol absorption different from the nominal aerosol optical depth of 0.04 (Platt et al. 1984).

In order to utilize the observed microwave water vapor path (w), w was normalized to the observed radiance when the sky overhead was clear. Intermediate radiance's were then related to the measure values of w . If necessary, radiance's were interpolated between observation points. The water vapor transmittance was then related to the corresponding radiance. This was done for the Kavieng data by regressing all theoretically calculated radiance's against corresponding transmittances for the total period of Kavieng radiosonde observations. The resultant plot is shown in Figure 2.

The standard LIRAD technique utilized the retrieved lidar profile of backscatter coefficient to calculate a cloud radiance, with the radiosonde data interpolated between the times of two radio soundings. The radiance calculation used a simple non-scattering radiative transfer equation, with spectral blackbody radiances assigned to different cloud levels and a constant ratio (ξ) assumed between cloud absorption coefficient and lidar backscatter coefficient. The measured surface radiance was corrected for water vapor emission and absorption and then compared with the calculated value. The constant ξ was then adjusted until agreement between the measured and observed radiances was obtained. The cloud infrared emittance and transmittance was then calculated. Corrections were made for first order effects of both scattering within the cloud and reflection of upwelling surface radiation.

The cloud backscatter coefficients were corrected for attenuation by a forward integration technique (Platt 1979) using a cloud backscatter to extinction ratio $k/2\eta$ (see Equation 1) derived from the lidar integrated backscatter $\gamma'(\pi)$ when the cloud became optically thick and the emittance tended to unity. The cloud visible optical depth could thus also be derived. The cloud infrared optical depth is similarly derived from the infrared emittance. The uncertainties in both the visible and infrared optical depth calculations are known to become greater as the cloud becomes optically thick. A relevant error analysis was given by Platt (1979).

Results and Discussion

Preliminary results were reported in Platt et al. (1995), where it was shown that the lidar integrated backscatter $\gamma'(\pi)$ correlated well with the infrared emittance in one limited period. Here we present a more complete set of results showing the behaviors of cloud depth, infrared emittance, and integrated backscatter.

The depths of the cirrus clouds observed over Kavieng are shown in Figure 3. They were consistently large, in the region of 3 or 4 km on average. These depths were considerably greater than those measured either at Darwin or in mid-latitudes (Platt et al. 1984, 1987). The Darwin points are also shown for comparison in Figure 3. The clouds were generally multi-layered although they often consolidated during daylight hours, descended, and became optically denser. These deep, multi-layer clouds would indicate deep layers of moist air below the tropopause, presumably due to water vapor ADVECTED from thunderstorms or mesoscale systems.

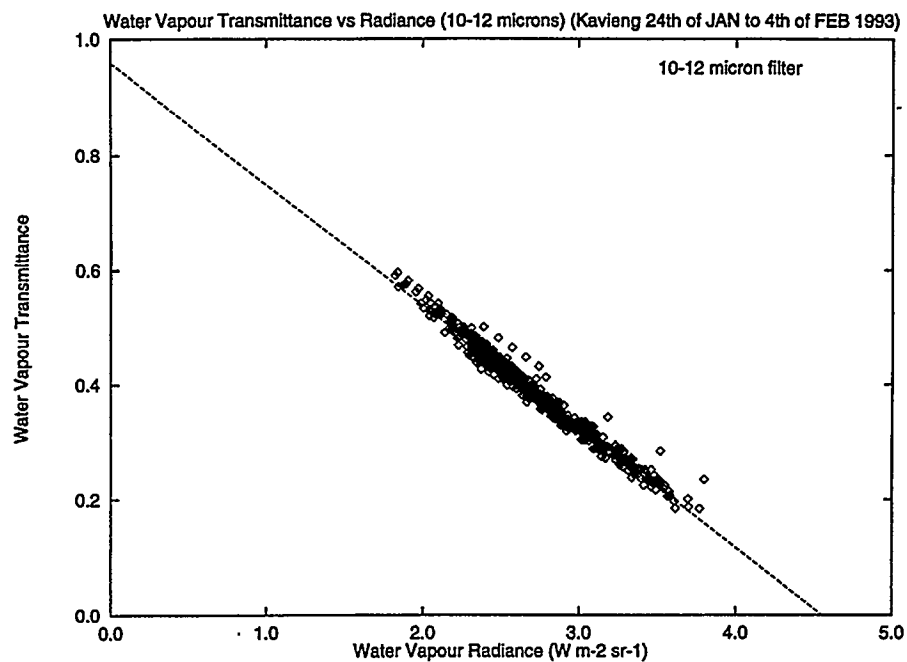


Figure 2. Plot of the water vapor transmittance from the cirrus clouds to the surface against water vapor radiance.

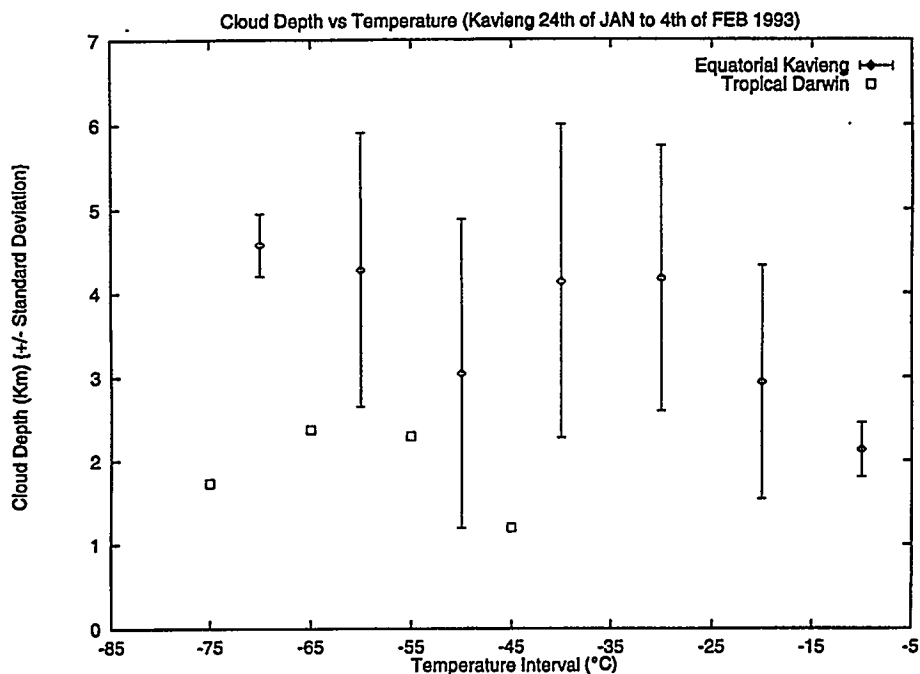


Figure 3. Plot of cirrus cloud depth against mid-cloud temperature. The bars represent the variability in cloud depth.

The beam emittance is shown in Figure 4 where the frequencies of various emittances are displayed. The emittance peaks at 0.2 to 0.3 which again is different from both Darwin and the mid-latitude cases for synoptic cirrus and values were more typical of anvil cirrus (e.g., Platt et al. 1984). In the former cases, the cirrus frequency decreased systematically with emittance.

Plots of integrated backscatter (see Equation 1) versus infrared emittance are shown in Figures 5 and 6 for the temperature regions of -35°C to -75°C and -5°C to -35°C , respectively. These have a characteristic shape typical of the Darwin and mid-latitude clouds. Values of $k/2\eta$ have been deduced for each temperature range and are shown plotted against temperature in Figure 7. In this case, the values are very similar to the Aspendale summer values, but lower than those at Darwin, where all points of $\gamma'(\pi)$ versus ϵ tended to lie on the same line. This behavior in the integrated backscatter in equatorial cirrus is encouraging in that the crystal composition of clouds at various latitudes may be quite similar. There are, of course, exceptions, as shown in Figure 7, that need further investigation.

Conclusions

The data on equatorial cirrus presented here are quite unique. As intended in the original aims of PROBE, the data form an initial resource for climate modelling of cirrus over the warm pool areas of the Tropical West Pacific region. The data will first be used in a study, in collaboration with G. L. Stephens, to determine the radiation budget of and heating in the cirrus clouds. This forms a study that is central to the general ARM thrust. The data will be available from in the PROBE section of the TOGA COARE archive. The visible optical depths and lidar linear depolarization ratios of the clouds for some selected clouds have also been determined and will soon be available.

The results point to the desirability of obtaining further data on equatorial cirrus. Particularly, data is needed with high-power lidar and radiometry and interferometry. It would also be desirable to have millimeter radar. The best mode for such data might be a campaign-type experiment at one of the new tropical Western Pacific CART sites, when they are installed.

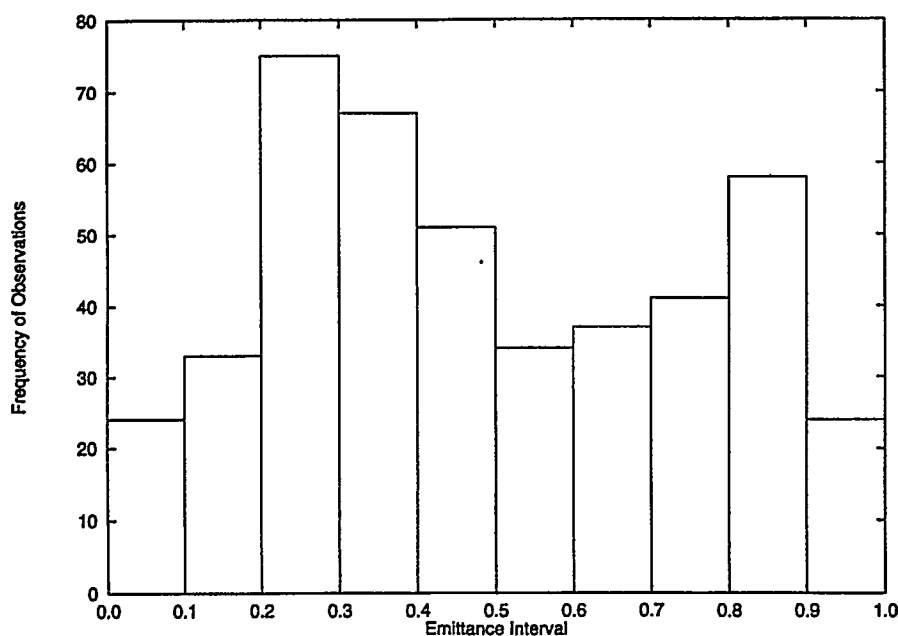


Figure 4. Frequency of observation of beam emittance of a given value.

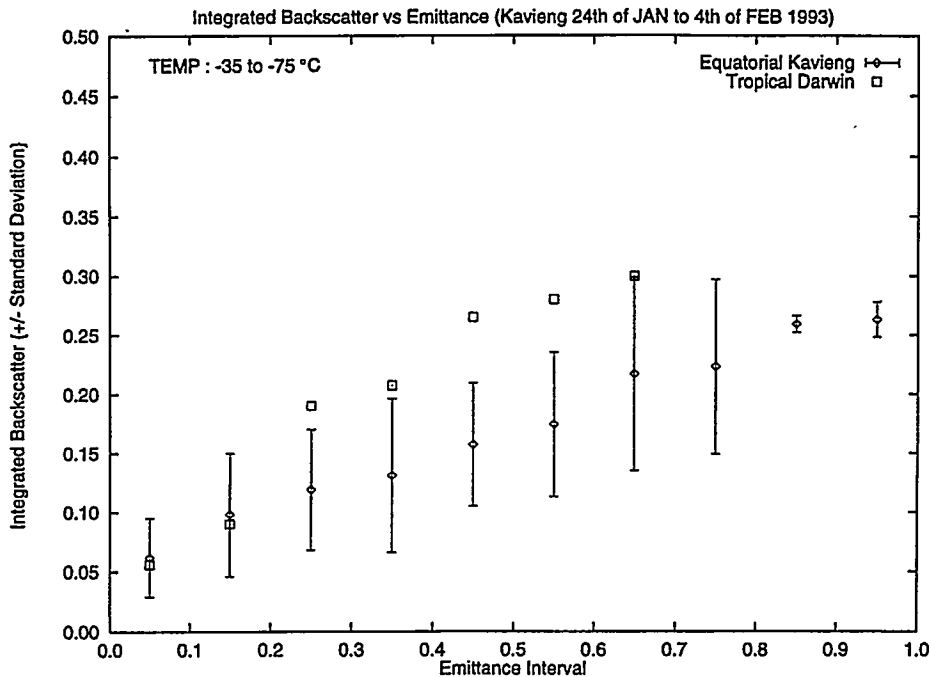


Figure 5. A plot of integrated backscatter against infrared emittance for the temperature range shown. Bars represent variability in observed backscatter, -35 to -75°C.

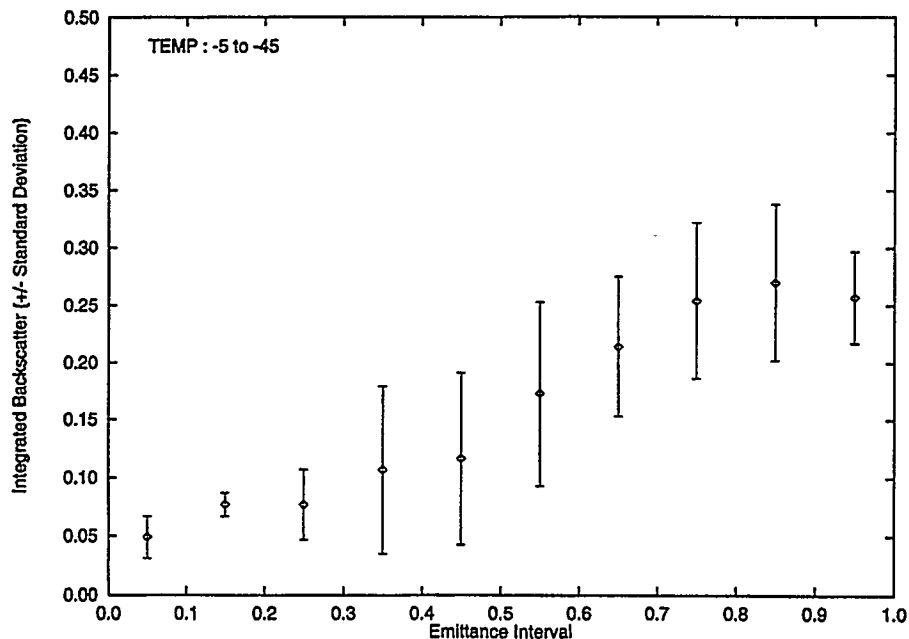


Figure 6. A plot of integrated backscatter against infrared emittance for the temperature range shown. Bars represent variability in observed backscatter, -5 to -45°C.

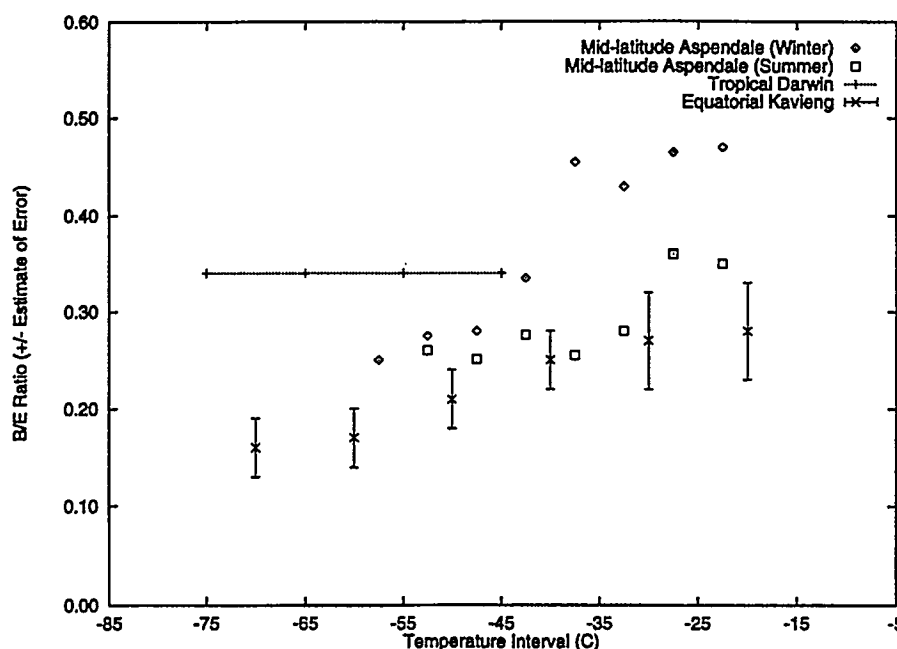


Figure 7. Plot of the isotropic backscatter to extinction ratio modified by multiple scatter of against mid-cloud temperature.

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Atmospheric Emitted Radiance Interferometer (AERI): Status and the Aerosol Explanation for Extra Window Region Emissions

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Introduction

High spectral resolution observations of downwelling emission from 3 to 19 microns have been made by the Atmospheric Emitted Radiance Interferometer (AERI) Prototype at the Southern Great Plains (SGP) Cloud and Radiative Testbed (CART) site for over two years. The spectral data set from AERI provides a basis for improving clear sky radiative transfer; determining the radiative impact of clouds, including the derivation of cloud radiative properties; defining the influences of aerosols in the window regions; and retrieving boundary layer state properties, including temperature, water vapor, and other trace gases. The data stream of radiometrically and spectrally calibrated radiances is routinely provided by Pacific Northwest Laboratory (PNL) to those science teams requesting it, and further information on the instrument and data characteristics is available in the ARM Science Team proceedings for 1993 and 1994 (Revercomb et al.; Smith et al.) and in several conference publications (Revercomb et al. 1993a,b, 1994; Smith et al. 1993).

The emphasis of this science team project is 1) monitoring the AERI performance at CART to define improvements for the design of the operational AERIs and AERI-X systems support; 2) improving our understanding of the radiative transfer of the clear sky, including aerosols; and 3) retrieving trace gas. In this paper we report on the first two areas which have been our major priority to date.

AERI Status

A somewhat diverse selection of topics related to the AERI instrument and its observations are contained in this section.

We begin with the status of the prototype instrument and automated analyses from the SGP CART site, summarize tests that verified the high level of accuracy of the AERI calibration in the CART configuration, briefly describe a very successful recent field experiment conducted with the new operational version of AERI (AERI-01), and finish with the plan for bringing the new AERI to CART.

Prototype at CART

The AERI Prototype was first brought to CART in March 1993 and has operated in its current configuration since December 8, 1993. Sky spectra are observed every 10 minutes for at least 16 hours per day on week days, except during periods of precipitation. When precipitation is detected, a hatch automatically closes to protect the instrument. The operational schedule is currently constrained by the need for manual liquid nitrogen filling of the detector dewar. During Intensive Operating Periods (IOPs) when the CART site is continuously manned, sky data are collected 24 hours a day. Routine 24-hour operation will become the norm when the first operational AERI is delivered to CART with a Stirling cooler, as discussed at the end of this section.

AERI data have been used as the focus of an ARM Instantaneous Radiative Flux (IRF) Quality Measurement Experiment (QME). For every launch of the Balloon-Borne Sounding System, a spectrum is now calculated using the Line By Line Radiative Transfer Model (LBLRTM) constructed for ARM (Clough et al. 1994) and compared to AERI observations. While this quality measurement experiment is still being refined to provide all of the quality control and sky condition information needed for

meaningful comparisons, it will soon be a valuable, automated resource for interpreting AERI data.

Calibration

Early comparisons of AERI minus LBLRTM difference spectra from the SGP indicated an unexpected positive bias in the longwave window region when compared to the mean differences for similar atmospheric conditions from the spectral radiance experiment (SPECTRE) in Coffeyville, Kansas in 1991 (Ellingson et al. 1994). Such a bias could be caused by an unidentified absorber in the atmosphere at the CART site, as discussed later. However, while not considered likely, the possibility of a calibration error needed to be investigated (the bias is only a few percent of ambient radiance but several times larger than expected absolute calibration errors). Therefore, a careful investigation of the AERI calibration was conducted at CART in early April 1994, prior to the April Remote Cloud Sensing (RCS) IOP.

Two general mechanisms that could lead to an erroneously large window region radiance were considered: 1) a small obstruction of the field-of-view (FOV) for sky viewing, and 2) a blackbody temperature error. Changes of instrument transmission or detector sensitivity are not possible explanations, because the AERI is calibrated during every 10-minute sky viewing cycle using views of two reference blackbodies, one at 60°C and one ambient. Also, while the alignment and emissivity of the hot blackbody are especially important, any misalignment or surface degradation would cause errors of the opposite sign.

The FOV for sky viewing was checked by moving the whole instrument with respect to the viewing port and environmental baffle. The test was conducted when the sky was clear, which guarantees a large contrast in window region radiance from the sky and from the room temperature structure. The window region signal level was recorded as the instrument position was sequenced through small incremental steps in two orthogonal directions. A margin of at least three inches was found in every direction before a significant signal from the structure was observed.

A calibration was carried out on both the electronics and the temperature sensors and the associated conditioning and readout electronics. The electronics calibration showed essentially no change since December 1993 (equivalent to less than 0.01°C). A precision thermistor was used as a transfer standard to reference the hot blackbody thermistor readings to a guideline precision radiation thermometer (traceable to NIST) located in Wisconsin. The precision

thermistor sensor calibration, performed on the hot blackbody at ambient temperatures to avoid temperature gradients, found agreement to within 0.03°C. Further, a temperature cycling test from 333 K to 300 K and back to 333 K showed good agreement between the two hot blackbody thermistors themselves. The maximum difference of 0.06°C occurred at the highest temperatures, where the largest cavity temperature gradients would be expected. The ambient blackbody sensors were not compared with the precision thermistor because they are never thermally stressed and the two individual sensors agreed to within 0.03 degrees. We conclude that the AERI calibration blackbody temperature sensors are accurate to about 0.05°C, while a temperature error of 0.5°C would be necessary to explain the observed bias as a calibration error (too high for ambient blackbody or too low for the hot blackbody). Later, we describe our new findings on the impact of aerosols, which we believe is very likely to explain the bias of early CART/SPECTRE comparisons.

As a final note on the AERI calibration, the current best estimates of calibration uncertainties due to uncertainties in the detector non-linearity, blackbody emissivity, and blackbody temperature are summarized in Figure 1. The spectra of these sources of absolute calibration error are compared with that for a change of downwelling radiance which occurs if the water vapor content is changed by 10% for the U.S. Standard Atmosphere.

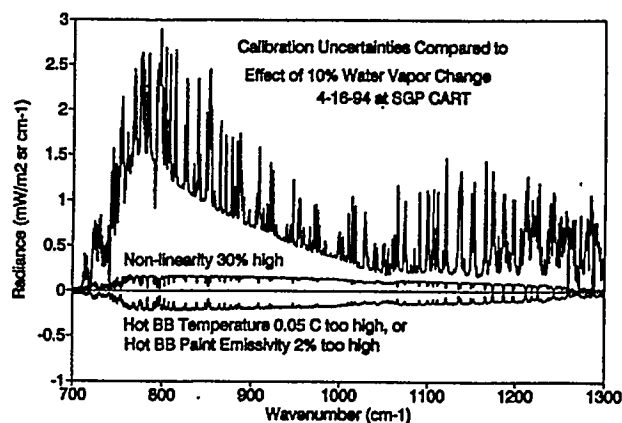


Figure 1. Size of AERI absolute calibration uncertainties from errors in the nonlinearity correction, the hot blackbody emissivity, and the hot blackbody temperature. Ambient blackbody temperature errors cause radiance errors of the opposite sign, and somewhat smaller in the window region. The effect of errors in the ambient blackbody emissivity is essentially negligible compared to that of the hot blackbody.

Field Experiment with the New AERI-01

The AERI-01 was a key component of a field experiment conducted in the Gulf of Mexico during January 1995 with the combined goals of improving sea surface skin temperature observations from shipboard as well as satellite and of evaluating the calibration of the new Geostationary Operational Environmental Satellite-8 imaging and sounding instruments. AERI once again demonstrated the hardness of modern interferometric systems by its successful operation onboard a research vessel, the Research Vessel *Pelican*. The joint National Aeronautics and Space Administration (NASA) and National Oceanic and Atmospheric Administration experiment coordinated NASA ER2 high altitude aircraft flights of the University of Wisconsin (UW) High-resolution Interferometer Sounder (HIS) and the MODIS Airborne Simulator (MAS) with ship-board observations, including *in situ* sea surface temperature observations by the Brookhaven National Laboratory (BNL) "skimmer," more conventional oceanic and meteorological observations, balloon launches with the UW CLASS system, and the AERI viewing the upwelling radiance at three angles as well as the downwelling radiance every 20 minutes.

The AERI produced what we believe are the first shipboard observations of sea surface emissivity at high spectral resolution and at multiple viewing angles. Also it simultaneously measured the sea surface temperature to an accuracy of better than 0.1°C , with very close agreement to the BNL observations sensitive to the top 15 cm of ocean. This experiment is described in detail in Smith et al. 1996. The experiment is a good demonstration of the absolute accuracy of the AERI and of techniques which can be applied to both land surface emissivity observations and to observations from the Unmanned Aerospace Vehicle AERI currently under design. In addition, on January 19, 1995, the ER2 collected data over the SGP CART site, which should provide the best example to date of simultaneous upwelling HIS and downwelling AERI spectral radiance observations for clear sky analyses over the CART site.

AERI-01 Schedule

The delivery of the first operational AERI will take place during the spring of 1995. This milestone has been delayed to make it possible to equip the instrument with a Stirling cooler that will eliminate the need for liquid nitrogen to cool the detectors. The AERI-01 will operate continuously, unconstrained by site manpower schedules. It has been

shown that a Stirling cooler can be operated on the instrument without causing radiometric artifacts from vibration (potentially disturbing the interferometer or inducing detector microphonics) or from electromagnetic interference. The excellent radiometric performance realized with a Litton Stirling cooler is illustrated in Figure 2, showing an AERI longwave band sky spectra observed using the Stirling cooler compared with one using the normal liquid nitrogen cooler. A complete cooler subassembly has been designed and implemented on the AERI-01, including an integral cooling fan and capabilities to continuously monitor the cooler performance (power usage and detector temperature). We have experienced some cooler operational problems, but believe that they are very close to being resolved. The coolers being used are guaranteed for one year of operation and can be removed for refurbishment without removing or replacing the detector/dewar package.

Window Region Emissions: The Aerosol Explanation

One of the key spectroscopic issues to resolve with AERI data is the quantification of sources of opacity in the clear sky window region. Accurate surface energy budget modeling depends on understanding this issue. There are at least three important contributors to uncertainty in the window region opacity that are somewhat difficult to separate: 1) water vapor continuum, 2) aerosols, and 3) atmospheric water vapor profile. We have not included clouds or trace gases, assuming that clouds can be identified

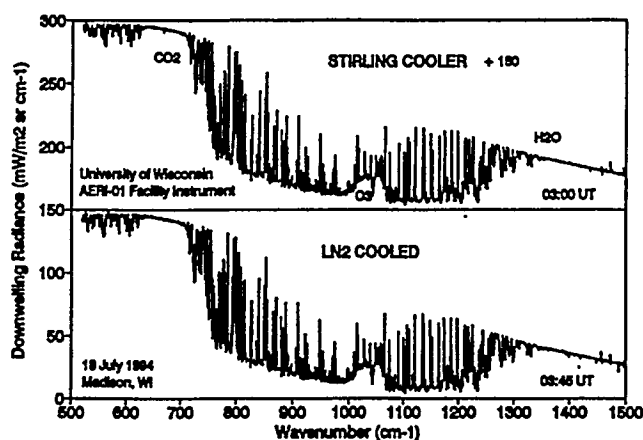


Figure 2. Comparison of a longwave AERI spectrum taken using a Stirling cooled detector to one using a liquid nitrogen cooled detector, while observing clear skies. The agreement is remarkable, considering that the spectra were observed 45 minutes apart.

and avoided by careful use of ancillary CART observations and that the signatures of trace gases can be identified and accounted for. Here we present evidence that the spectral signature of a typical rural aerosol (as defined by Shettle and Fenn [1979], and included in Phillips Laboratory radiative transfer codes LOWTRAN, MODTRAN, and FASCODE; and also in LBLRTM) is present in a large number of the AERI spectra from CART, and that aerosol absorption is the likely explanation for the bias reported at the last ARM Science Team Meeting between SPECTRE and CART AERI observations.

Figure 3 shows an example of a comparison between an AERI observed longwave window region spectrum and an LBLRTM calculation. The difference in the window region is typical of the smaller differences observed during the April 1994 Remote Cloud Sensing IOP studied closely for this analysis. The AERI minus LBLRTM difference or residual, shown in more detail in Figure 4, is somewhat larger than $1 \text{ mW/m}^2 \text{ sr cm}^{-1}$ in the low resolution background region between spectral lines, which is on the order of the discrepancy between SPECTRE and CART radiance differences in the window regions.

It is clear from the comparisons with the radiative impact of a rural aerosol with 20-km visibility calculated with LBLRTM that the residual spectrum can be accounted for in large part with such an aerosol model (Figure 4b). It is equally clear from the comparison with the calculated effect of a 10% increase in water vapor that the residual cannot be matched by adjusting the water vapor amount or by a percent change in the water continuum (Figure 4a). The effects of these water vapor changes are relatively too weak on the shortwave side of the 8–12 micron window region and

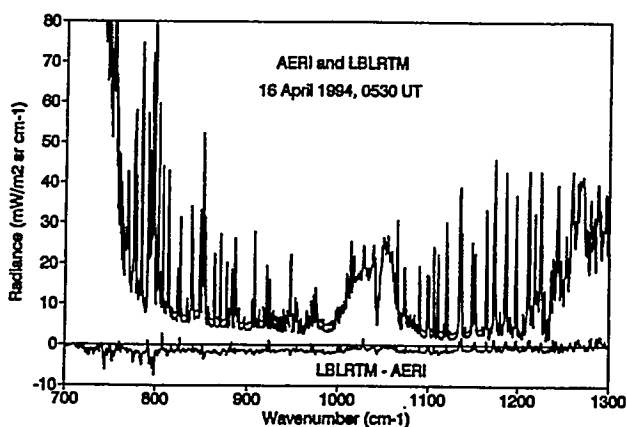


Figure 3. Comparison of an AERI observed longwave window region spectrum and calculated spectrum from LBLRTM.

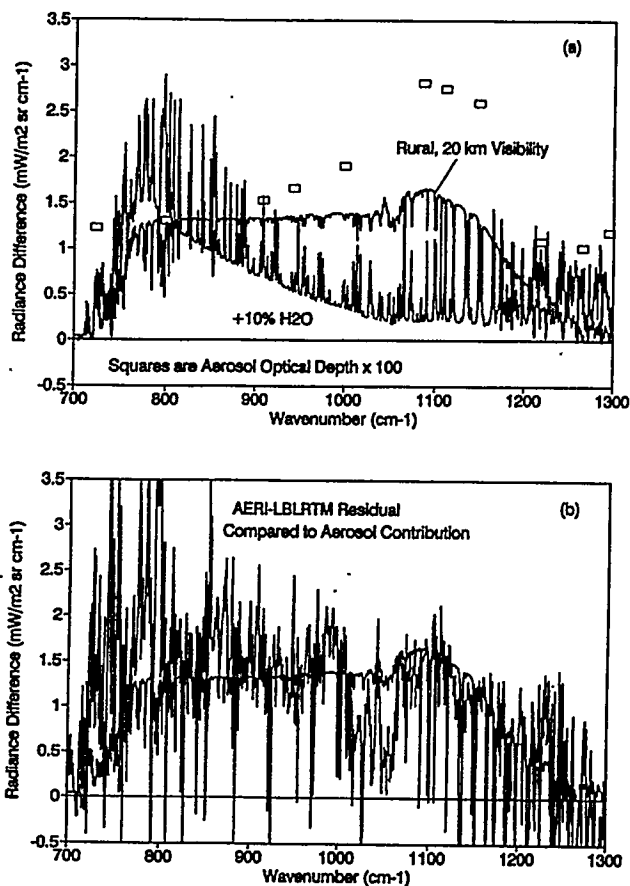


Figure 4. Rural aerosol spectral signature compared to that for (a) a 10% change in water vapor and (b) to one of the smaller AERI minus LBLRTM residuals observed during the RCS Intensive Operating Period in April 1994.

too strong on the longwave side. The relatively large spikes in the residual spectrum from water vapor lines suggest that the AERI data can also be used to determine the correct absolute water vapor amount by evaluating a set of spectral lines for which the spectroscopy is well established. Such adjustments of the absolute water amount will also allow the uncertainty in the water vapor continuum deduced from AERI observations to be reduced.

Aerosol retrievals from AERI data using the rural aerosol model in LBLRTM are summarized in Table 1. The minimum aerosol fit had a 0.53-micron visibility of 23 km and a radiance impact of $1.2 \text{ mW/m}^2 \text{ sr cm}^{-1}$. The lack of a simple correlation with precipitable water or relative humidity suggests that more complex analyses are needed to provide a detailed understanding of the factors determining aerosol amount.

Table 1. Atmospheric properties and window region residuals for selected clear days during the April 1994 Remote Cloud Sensing IOP. The radiance residual listed is the 1000 cm^{-1} radiance of the aerosol fit to the AERI minus LBLRTM difference.

Day/Time April/UTC	T/P °C/mb	Precip. Water cm	R.H. %	Visibility km	Residual $\text{mW/m}^2 \text{sr cm}^{-1}$
16/05:30	14/933	0.9	47	20	1.3
17/02:30	23/985	1.2	44	23	1.2
17/08:30	17/984	1.3	49	18	1.7
23/11:30	15/975	2.5	76	18	1.4
24/05:30	20/972	2.9	69	15	1.9
26/02:30	21/961	1.6	85	1.5	15.5
26/05:30	21/962	0.8	51	8	4.1
27/03:00	18/959	1.3	43	9	3.4

These data show that aerosols are present for extended periods at the SGP CART site, in amounts that can have significant impact on the downwelling radiance. This is a plausible mechanism to explain the relatively higher window region radiances observed by AERI at CART than during the SPECTRE period at Coffeyville in 1991.

Future AERI Plans

The first AERI operational version with no need for liquid nitrogen will be delivered to CART in the Spring of 1995, as mentioned earlier. The implementation of the boundary layer sounding AERIs at each of the SGP boundary sites and AERIs in the Tropical Western Pacific and the North Slope of Alaska is expected to follow as quickly as possible. The contract for these additional AERIs is being written and the systems will be built at a rate of at least two per year as soon as it takes effect.

Note: After this paper was submitted, an obstruction to the AERI-00 Prototype sky view was found by comparing spectra with those from the new AERI-01. The obstruction explains the AERI minus LBLRTM bias potentially explained as ubiquitous aerosols at the SGP CART Site in this paper. The paper is published in its original form to maintain the historical record. Corrected AERI prototype data are now available and some large aerosol signatures are

still present. The detailed impact of the correction will be addressed in our 1996 Science Team paper.

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The Lidar Dark Band: An Oddity of the Radar Bright Band Analogy

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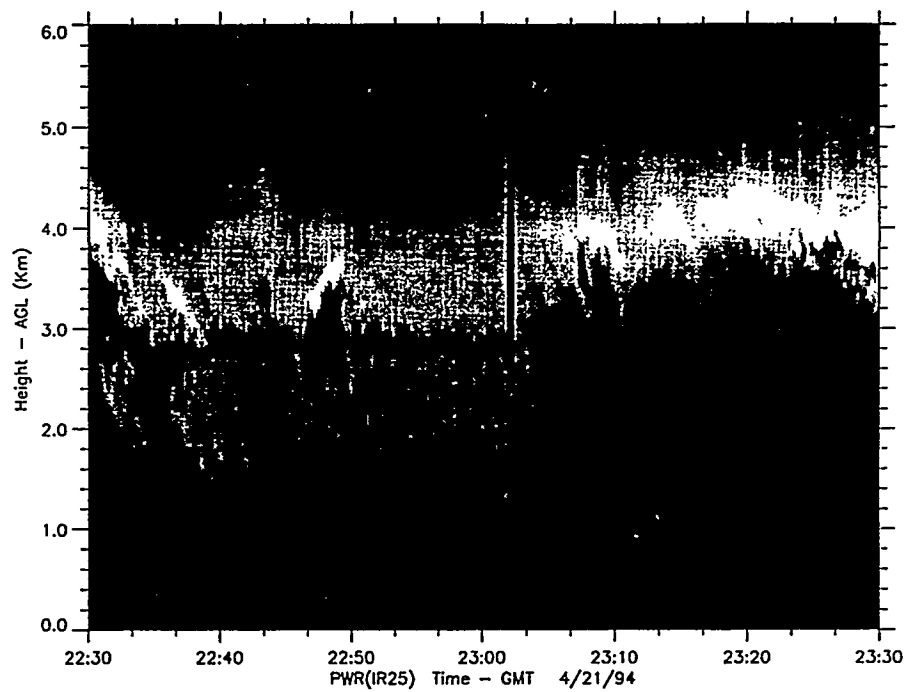
It has long been recognized that rain at the surface often begins as snow in the elevated subfreezing portions of the atmosphere. Attempts to comprehend the details of this process have, over the years, illuminated fundamental properties of mixed phase clouds and the production of precipitation. Similarly, the unique challenges presented by the evolution in the scattering of electromagnetic waves by hydrometeors undergoing the solid-to-liquid phase transition have contributed to advancing our basic understanding of radar and lidar observations of the atmosphere. Since the time of pioneering World War II radar research, anomalous returns from the melting layer have been known as the bright band because of the bright appearance of the layer of melting snow on radar display oscilloscopes (Battan 1973). In addition to the enhanced radar reflectivity factors, increased depolarization and a rapid change in Doppler-derived particle fallspeeds also characterize the melting layer. These attributes result from a combination of factors, including the effects of wet snowflake nonsphericity, the difference in dielectric constants between water and ice, and the large increase in fallspeeds of raindrops.

Surprisingly, since it happens for dissimilar reasons, a lidar bright band phenomenon also occurs at optical frequencies (Sassen 1977a). In this case, it is essentially the increase in particle size encountered while ascending through the melting region that initially causes increased backscattering, followed soon afterward by overwhelming attenuation, that defines the returned energy peak known as the lidar bright band analog. This peak, however, occurs in the snowfall typically above the 0°C isotherm. On the other hand, lidar depolarization appears to maximize near the bottom of the melting layer from wet snowflakes just before they collapse into raindrops (Sassen 1975).

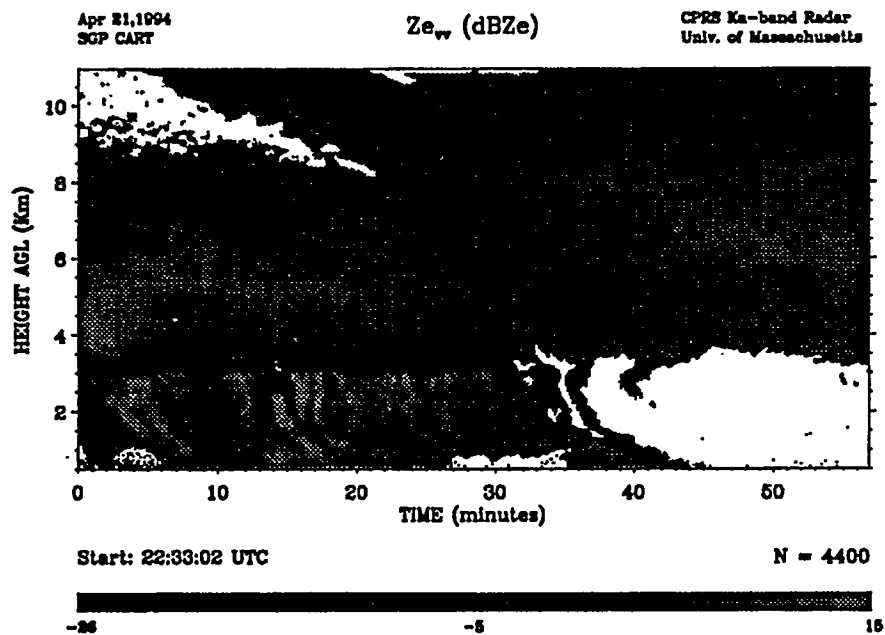
Although much has been learned from independent radar and lidar studies of precipitation, occasionally supported by aircraft profiling, what has been lacking is combined optical, microwave, and in situ observations of the melting layer. Fortunately, the rainshowers on April 21, 1994, during the Remote Cloud Sensing Intensive Observations Period (RCS IOP) at the Southern Great Plains Cloud and Radiation Testbed (CART) site provided an opportunity for coordinated dual-wavelength University of Utah Polarization Diversity Lidar (PDL, at 0.532 and 1.06 μm), University of Massachusetts Cloud Profiling Radar System (CPRS, at 9.1 and 3.2 mm) Doppler radar, and University of North Dakota Citation aircraft measurements.

Vertically-pointing PDL and CPRS data were jointly collected from about 2230-2330 UTC on April 21, 1994, as the first of several light rainshowers ahead of a line of thunderstorms drifted over the site. The sporadic precipitation fell from lowering anvil and embedded stratus clouds. At 2310 UTC the Citation began a vertical spiral ascent above the site, and although little rain was sampled in the dry air layer below the freezing level (at 3.9 km), large snowflakes and some mixed-phase cloud conditions were probed up to a height of 6.5 km before the mission was terminated because of data computer problems.

For this 1-hour period, Figure 1a shows 1.06- μm laser backscattering (in arbitrary units based on a logarithmic grayscale), 1b shows 9.1-mm reflectivity factors in dBZe, 1c shows 3.2-mm vertical Doppler velocities V , and 1d shows 9.1-mm linear depolarization ratios (LDRs) (note inserted radar scales). Together, these displays illustrate several classic features of the radar/lidar bright band. However, what is of most interest here is the presence of the previously unreported lidar dark band between 2.8-3.0 km in Figure 1a.

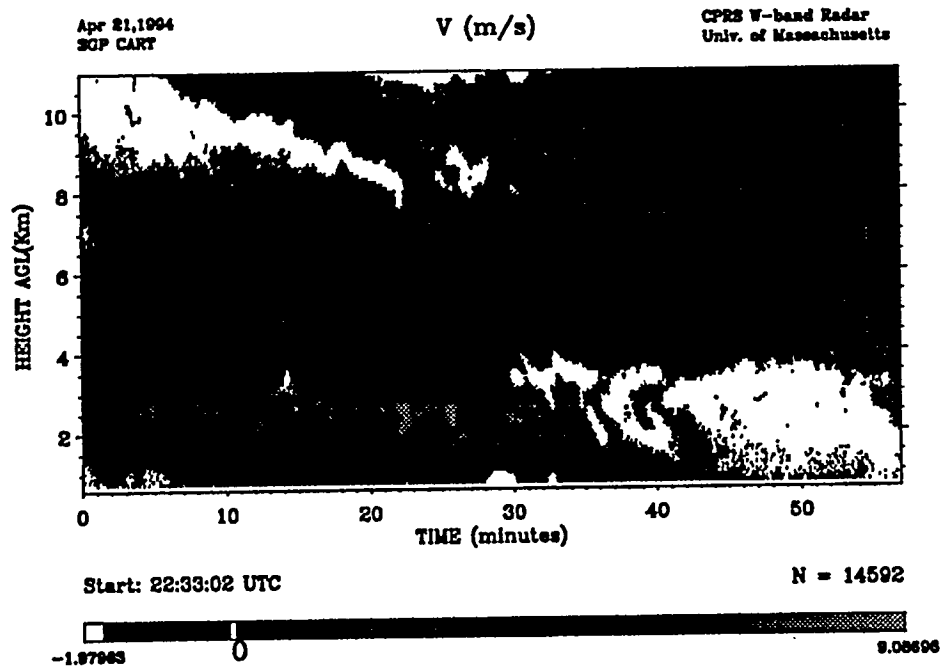


a)

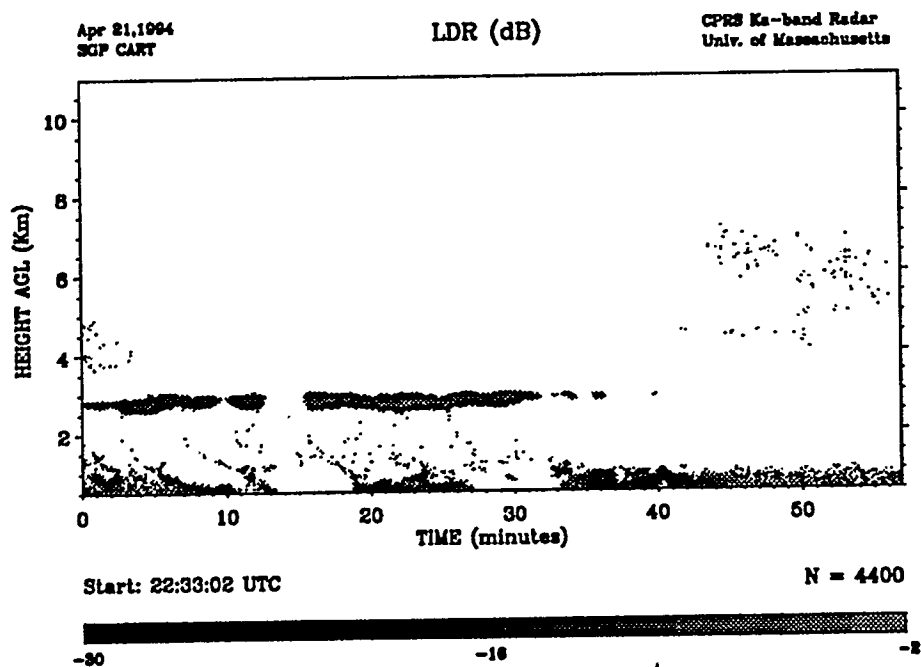


b)

Figure 1. a) Approximately 1-hour height-versus-time plots from the April 21, 1994, RCS IOP case study at the SGP CART site, showing, a $1.06\text{-}\mu\text{m}$ PDL backscattering (note the dark band between 2.8-3.0 km); b) 9.1-mm CPRS radar reflectivity factors in dBZe.



c)



d)

Figure 1 (cont). c) 3.2-mm CPRS vertical Doppler velocity; d) 9.1-mm CPRS linear depolarization ratios.

According to the lidar depolarization data (not shown), the dark band occurs just where the melted snowflake-to-rain-drop transition happens. According to previous laboratory studies of melting ice particles suspended in a polarized laser beam (Sassen 1977b), a minimum in parallel-polarized backscattering occurs when the melting ice mass occupies the center of a raindrop; this blocks the strong paraxial reflection off the far drop face. Interestingly, it is at a height of about 3.1 km that the major changes in microwave scattering are seen, such as sudden increases in V and dBZe and the beginning of a gradual LDR increase. In other words, at least in this case (where hydrometeor evaporation appears to have been important) the radar bright band features are associated with inhomogeneous raindrops, not the wetted snowflakes as is usually assumed.

These illustrations of our findings from this unique dataset shed new light on the scattering and microphysical properties of the melting layer and suggest that revisions to our understanding of the radar bright band are in order.

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Interaction of Clouds, Radiation, and the Tropical Warm Pool Sea Surface Temperatures

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Introduction

The primary focus of this study is the Tropical Western Pacific (TWP). In this study, we combine in-situ observations (Tropical Ocean Global Atmosphere [TOGA]-Coupled Ocean Atmosphere Response Experiment [COARE] and Central Equatorial Pacific Experiment [CEPEX]) with satellite cloud data. Our objectives are fourfold:

- Use surface, aircraft and satellite observations to relate radiation fluxes within the atmosphere and at the surface to water vapor, clouds, cloud microphysical properties, and aerosols
- Use the observations to estimate the warm pool heat budget and the role of deep convection in regulating the evaporation and radiation fluxes at the sea surface
- Use the above results to develop and validate General Circulation Models (GCM) treatment of clouds, water vapor, radiation fluxes, and the surface heat budget over the TWP
- Use the GCM to understand the interactions between clouds, radiation, convection, and the TWP sea surface temperatures.

Recent results are presented on four interrelated issues: treatment of cloud-radiation interactions in GCMs, comparison between a GCM and observations of aspects of the moisture budget, the coupling between convection and surface evaporation, and penetrative radiation and its effect on TWP sea surface temperature (SST). These topics are relevant to the TWP radiation balance (Figure 1): clouds and atmospheric moisture influence the reflection and absorption of radiation; radiation and evaporation are both affected by convection and interact with TWP SST. Short wave radiation penetrates into the upper ocean and, in

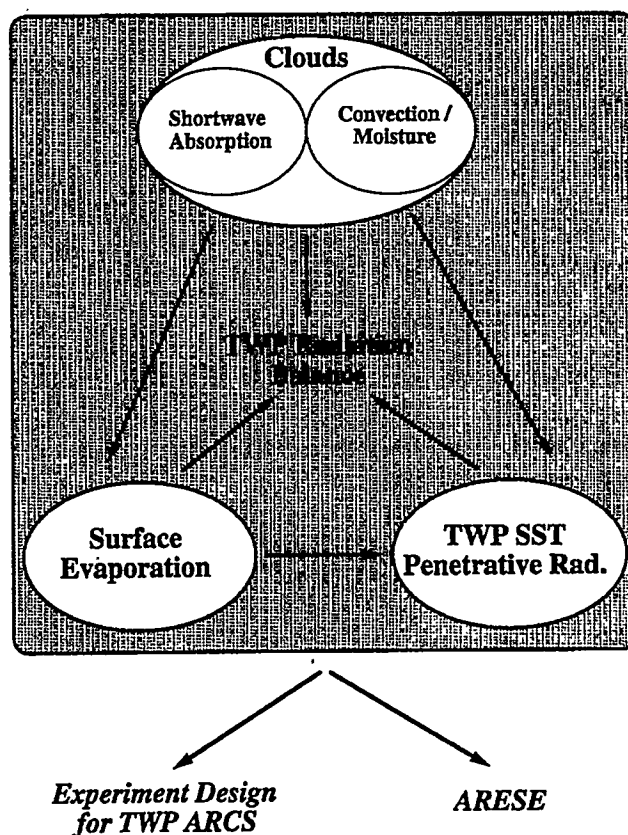


Figure 1. Conceptual relationships of studies presented in this report.

combination with heat losses at the surface, can cause vertical mixing. Thus, it influences TWP SST and thereby the state of the overlying atmosphere. Finally, the results of our studies will help the experimental design of the TWP Atmospheric Radiation and Cloud Station (ARCS) and the upcoming ARM Enhanced Shortwave Experiment (ARESE).

Treatment of Cloud-Radiation Interactions in GCMs

One of the fundamental objectives of our ARM related studies is to improve the treatment of cloud-radiation interactions in GCMs. Two recent ARM related studies (Cess et al. 1995 and Ramanathan et al. 1995) provide compelling evidence that the present generation of GCMs underestimate the absorption of solar radiation by clouds over the western Pacific warm pool by about 35 Wm^{-2} . We have begun a dedicated set of modeling and observational studies to track down the source of this serious deficiency and improve its treatment in GCMs.

Excess Solar Absorption in TWP Cloud Systems. To investigate the excess shortwave absorption by clouds, a numerical cloud generation model has been coupled to a discrete ordinates radiative transfer model (Lubin et al. 1995). The former was used to generate a cumulonimbus turret and three types of cirrus anvil (precipitating, extended, and detached) representing three stages of cloud evolution outward from the turret. The cloud particle size distributions, as a function of altitude, were used as input to the radiative transfer model using indices of refraction for pure water and pure ice, and equivalent sphere Mie theory. The radiative transfer model was used to calculate the ratio of cloud forcing at the surface to cloud forcing at the top of the atmosphere, both for the broadband shortwave and as a function of wavelength. Recent empirical studies have placed this cloud forcing ratio, f_s , at about 1.5, and our coupled model (Figure 2) easily yields this result for small solar zenith angles, when the cloud contains large ($>20 \mu\text{m}$) ice particles that absorb significantly in the near infrared (primarily the 1.6 micron window). The model result underscores the importance of large particles in cloud-radiative interactions and uncovered a flaw of GCMs that neglect drop and ice particles $>30 \mu\text{m}$. However, the empirical studies are based on diurnal averages, and our plane-parallel radiative transfer model yields an area and diurnally averaged cloud forcing ratio of only 1.2 for a tropical cumulonimbus and cirrus anvil system, primarily because of the rapid decrease of the ratio with zenith angle. The ratio decreases because of the increase in albedo with zenith angle, which is a characteristic feature of plane parallel clouds. This study identifies one potential cause for underestimation of solar absorption by models: the neglect of realistic topography of cloud tops and cloud shapes may overestimate the reflection for slant zenith angles, which in turn reduces the solar radiation available for absorption within the interior of clouds. Adding dust or aerosol to the cloud layers, to make them absorb at visible wavelengths, makes this ratio even smaller because the solar zenith angle

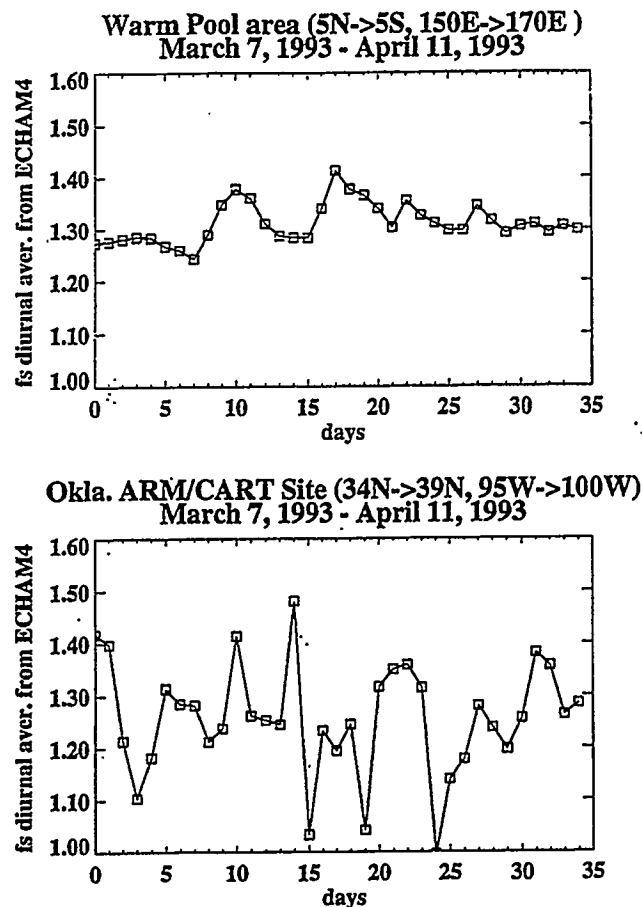


Figure 2. Time series of diurnally averaged cloud forcing ratio (f_s) for TWP and Oklahoma ARM/CART sites from a simulation with ECHAM4.

dependence in the cloud forcing ratio becomes more pronounced in the presence of small absorbing particles. This study suggests that the ARESE experiment should focus on three issues: 1) the zenith angle dependence of cloud albedos; 2) the understanding of gaseous absorption throughout the near infrared, and 3) three-dimensional cloud effects.

GCM Simulation Studies for the ARESE Experiment. The objective of the ARESE experiment is to confirm excess absorption by direct observations and unravel the physics behind the excess absorption. The aircraft experiment will be conducted over Oklahoma in the fall of 1995. To design the sampling strategy, we have analyzed the geographical, synoptic and diurnal variations in f_s . Figure 3 compares the computed ratios for the Warm Pool with that for Oklahoma. The model reveals significant day-to-day fluctuations, and variation with cloud types and cloud altitudes (not shown).

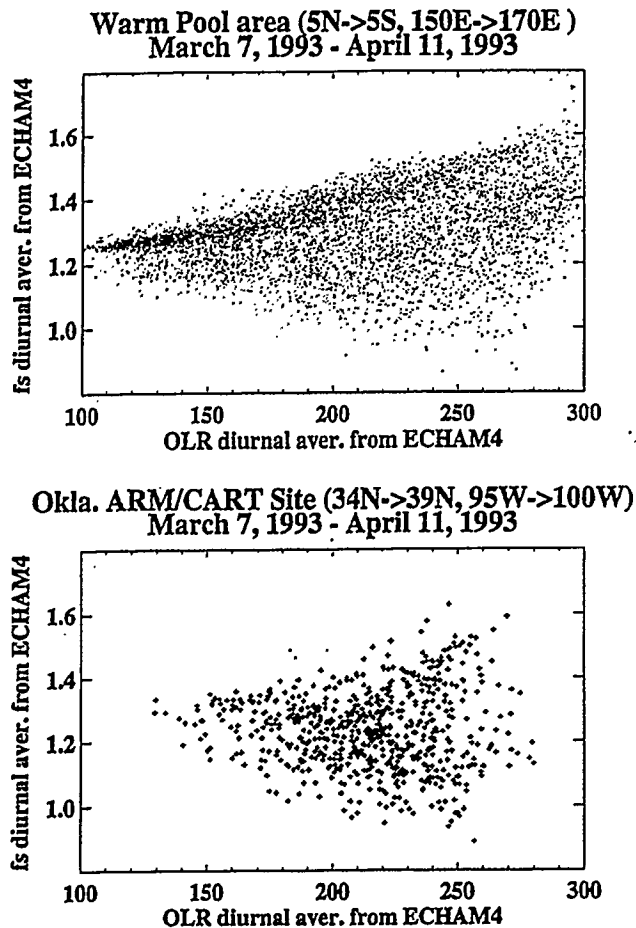


Figure 3. Scatter plot of diurnally averaged cloud forcing ratio (f_s) versus OLR for TWP and Oklahoma ARM/CART sites from a simulation with ECHAM4.

The magnitude and temporal scale of the fluctuations suggest that an observational period of at least six weeks and preferably two months would be required. Also, the ratio in the model is significantly different between the two ARM sites (TWP and SGP). The GCM seems to suggest that we cannot generalize the Oklahoma findings and that the ARESE experiment should be repeated over the TWP.

Comparison of Aspects of the Moisture Budget of ECHAM with CEPEX and Oklahoma ARM Site Measurement

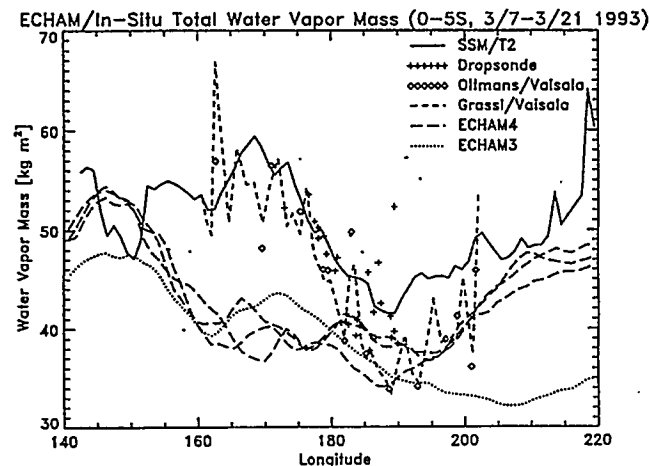
To demonstrate the use of observed data for GCM validation, aspects of the moisture budget are compared between

GCM simulations and field measurements taken during the CEPEX conducted from March 7 to April 5, 1993. One simulation of ECHAM3 and three simulations of ECHAM4 were carried out with slightly different initial conditions from the ECMWF analysis of March 1, 1993, and with the observed SST as lower boundary condition.

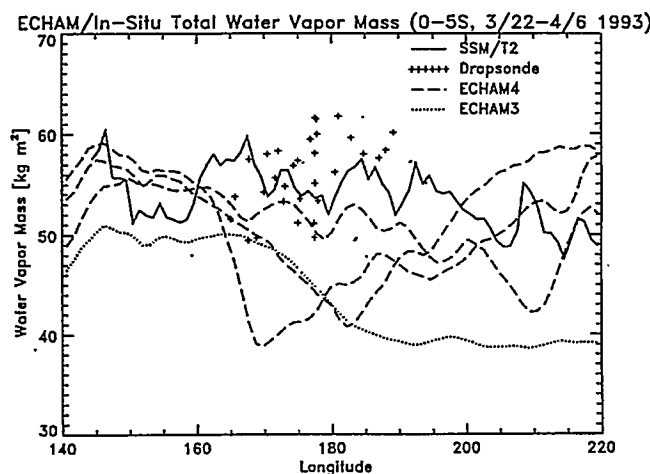
Radiosondes released between the equator and 5°S and the microwave satellite SSM/2 show a dry region east of the dateline in the first two weeks of CEPEX (Figure 4a), which vanishes in the second half of CEPEX (Figure 4b). ECHAM4 somewhat reproduces this change in convective activity, while ECHAM3 is too dry throughout and fails to pick up moisture in the second half in the eastern part. However, the three ECHAM4 simulations tend to develop different centers of convection later on, making comparison difficult.

Comparison of the mean relative humidity profiles indicates a dry bias in the mid troposphere and a moist one above 300 hPa, which shows up in ECHAM3, ECHAM4, and the ECMWF analyses (Figure 5). Microphysical cloud measurements conducted during CEPEX indicate an increase of ice water content (IWC) with temperature (McFarquahar and Heymsfield 1995). Its slope is similar but larger than data from a previous tropical dataset (Heymsfield 1993) indicate. While ECHAM3 systematically underestimates IWC, ECHAM4 results bear striking resemblance to the observations (Figure 6). Since many changes occurred from ECHAM3 to ECHAM4, such as new radiation code, detrained convective cloud water as a source for large scale water budget, and a change in the convection scheme, it is hard to determine what causes these differences.

To test the performance of ECHAM over land, a comparison was performed of the July/August total atmospheric water content above the Oklahoma ARM site from 8-day radiosonde measurements and the ECHAM3 (at T42 resolution) simulation (Figure 7). ECHAM3 was forced solely by global SSTs and was not initialized by observed atmospheric conditions. Thus, only the statistical properties of observations and simulation can be compared. It is impressive that the GCM simulated the correct mean water content over the month period. In addition, the simulated variability is reasonably close in magnitude and frequency to that observed, especially when it is realized that the comparison is between point observations and averaged properties of a GCM grid box (about 250 x 250 [km²]).



a)



b)

Figure 4. East-west section of total water vapor mass. Total water vapor mass is derived from the Grassl and Oltmans upsondes situated on the Vickers R/V, dropsondes from the Learjet aircraft, the microwave satellite SSM/2, ECHAM3 and ECHAM4 for the first a) and second half b) of CEPEX in the latitude belt 0-5°S.

Coupling Between Convection and Evaporation

The TWP SST is governed by surface energy fluxes due to the relatively small ocean heat transport. Solar radiation and latent heat flux are the most important components; both are significantly affected by convection. While the effect of convection and its associated cloud systems on reducing the surface solar radiation has been well recognized, the

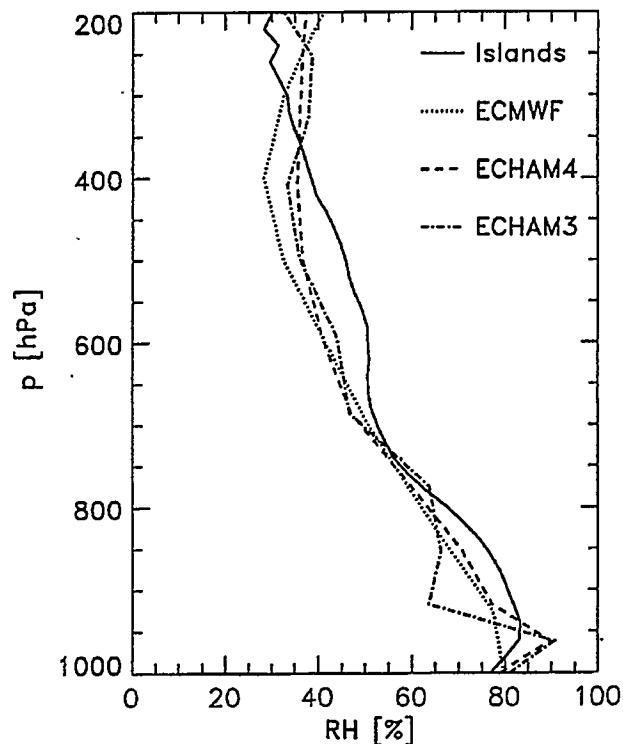


Figure 5. Mean relative humidity profiles. Mean relative humidity profiles averaged over the whole CEPEX time and domain from radiosondes released from all island stations, ECMWF analyses, ECHAM3 and ECHAM4.

coupling of surface evaporation with convection and its associated circulation is not clear. Since convection is abundant in the warm pool, estimates of the role of surface evaporation in limiting the warm pool SST without considering the coupling between convection and surface evaporation could be in serious error. Thermodynamic theories predict that evaporation increases with SST, due to the exponential increase of surface saturation water vapor pressure with SST. However, the predictions are not supported by available observations.

Using the data collected from the Tropical Atmospheric Ocean (TAO) buoy array (Figure 8) deployed in the equatorial Pacific for the period from 1991 to 1993, we computed the surface evaporation and analyzed its relationship with SST (Zhang and McPhaden 1995). Figure 9 shows the surface evaporation as a function of SST binned at 1°K SST intervals. Also shown as functions of SST are the major factors contributing to surface evaporation, surface wind speed and surface saturation specific humidity deficit. For SST less than 301°K, evaporation increases with SST as predicted by thermodynamic theories.

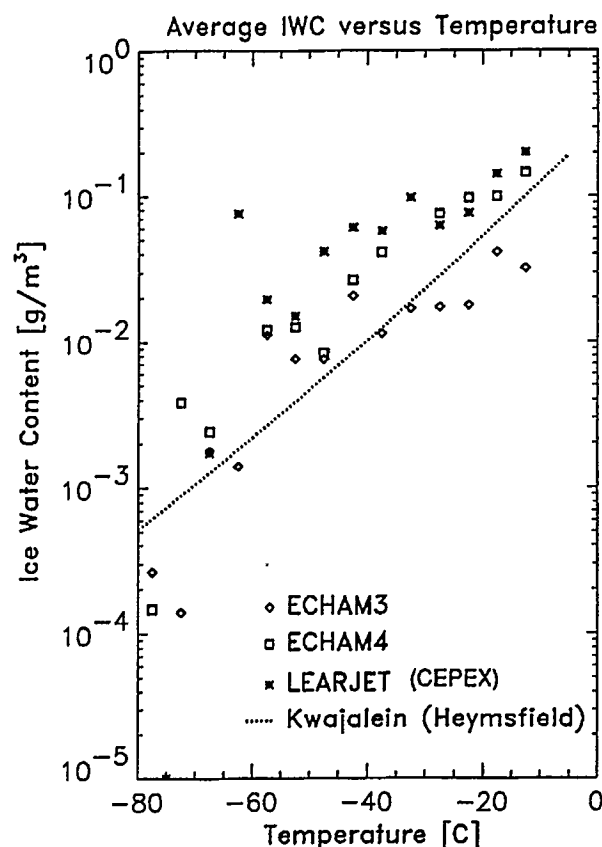


Figure 6. Ice water content (IWC) versus temperature. The average IWC for each 5 degree temperature bin versus temperature is derived from two-dimensional cloud probe data carried on the Aeromet Learjet (stars), ECHAM3 (rhombs) and ECHAM4 (squares). Additionally another derived relationship from a tropical dataset taken from the Kwajalein islands (Heymsfield 1993) is indicated as dashed line.

However, after reaching a maximum of about 110 Wm^{-2} at SST close to 301°K , evaporation begins to decrease with SST. Comparison between wind speed and humidity deficit indicates that this is a result of the wind speed coupling with SST. Despite a fast increase of humidity deficit with SST, which tends to enhance evaporation at high SST, the strong decrease of winds with SST leads to decrease of evaporation with SST.

Why are surface winds coupled with SST? One likely process responsible for it is the interaction between convection and the large-scale low level circulation (Zhang et al. 1995). The frequency of convection has been found by many observational studies to strongly depend on SST statistically: below a threshold of about 301°K , little convection is found; above this threshold, the frequency of

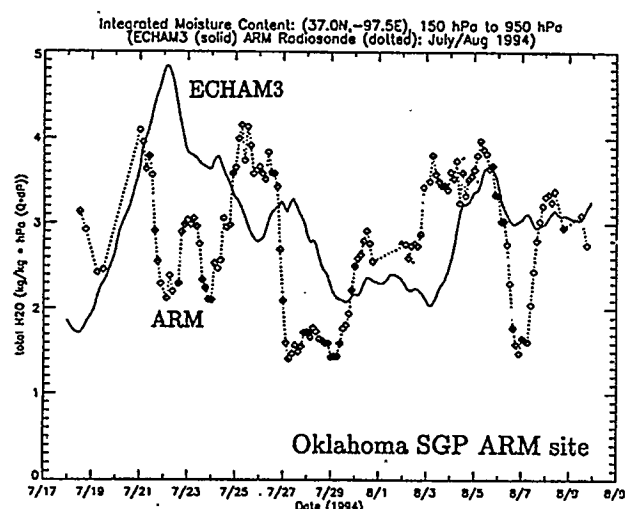


Figure 7. Comparison of the July/August total atmospheric water content above Oklahoma ARM site from 8-day radiosonde measurements (dotted) and the ECHAM3 AGCM at T42 resolution (solid).

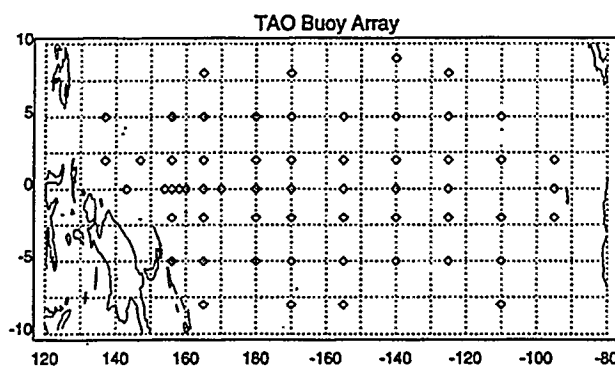


Figure 8. Positions of the TAO buoy array in the equatorial Pacific.

convection increases sharply with SST (Graham and Barnett 1987, Neelin and Held 1987, Zhang 1993). Regions of convection are also regions of strong large-scale low-level mass convergence and weak surface winds. Figure 10 shows the cumulative frequency distribution of surface winds as a function of outgoing longwave radiation (OLR), a surrogate for convection, from the AVHRR satellite data. Clearly, low winds are observed much more frequently in convective regime (OLR less than 240 Wm^{-2}) than in nonconvective regime. For instance, 50% of the observations have winds less than 4 m/s in convective regime, whereas only about 25% of the observations fall below this value in nonconvective regime.

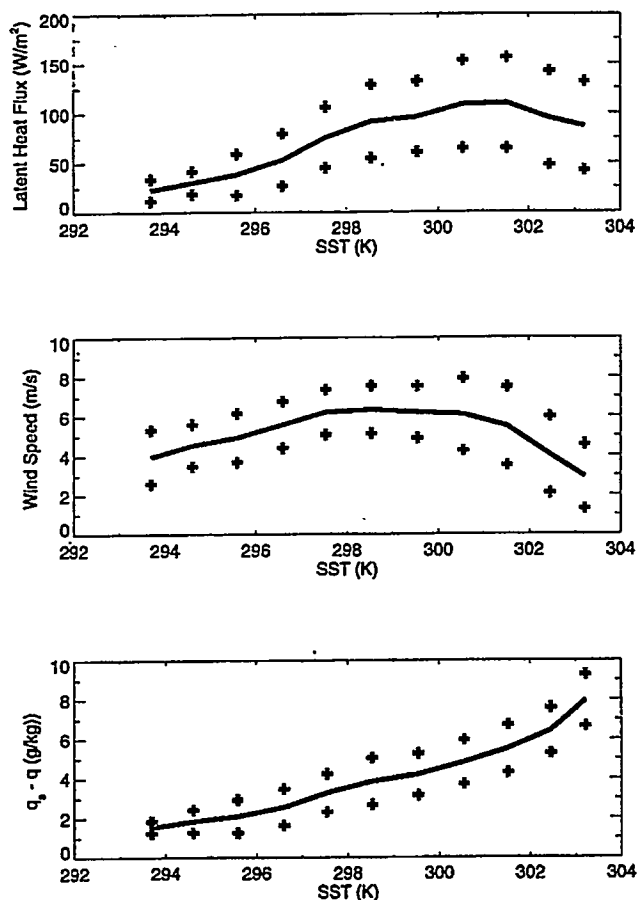


Figure 9. Surface evaporation, wind speed and air-sea humidity difference as functions of SST, as obtained from the buoy data. The means and the standard deviations for each SST bin are given.

In summary, contrary to common beliefs that surface evaporation should increase with SST, observations from the moored buoys show that for SSTs above 301°K, surface evaporation decreases with SST. This is a result of the sharp decrease of wind speed with SST, due to the coupling between convection and the low-level large-scale circulation. As SST increases above the threshold for the onset of convection, convection starts to develop. At the center of the large-scale low-level convergence that accompanies convection, wind speeds are low, resulting in lower surface evaporation.

Penetrative Radiation

The analysis of TWP in a coupled ocean atmosphere model (Schneider et al. 1996) suggested that penetrative radiation not only deposits heat below the surface, but, in combination

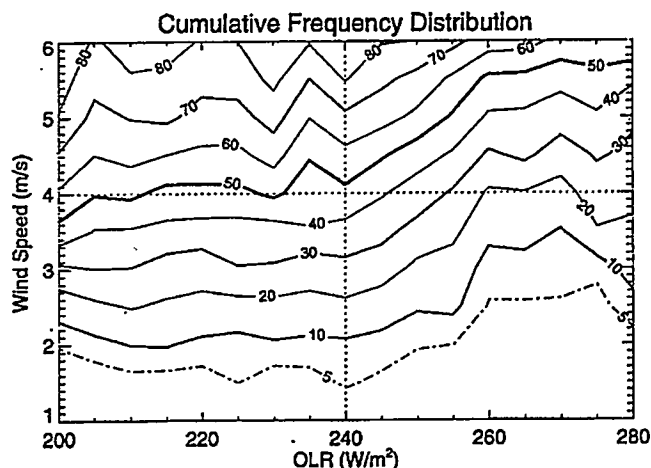


Figure 10. Cumulative frequency distribution of surface wind speed as a function of OLR (surrogate for convection) based on the two-year buoy data.

with heat losses at the surface, can destabilize the water column and cause vertical mixing. Thus the partitioning of incident shortwave radiation and cooling due to radiation and turbulent heat fluxes exert significant control on the mixed layer depth, the surface heat budget of the ocean and planetary boundary layer and, ultimately, the processes of atmospheric convection and cloud formation.

Incident solar radiation at the surface of the ocean is partially absorbed below the surface of the ocean, while turbulent and radiative heat losses apply at surface. Thus the water column can be destabilized and vertical mixing can result. This mixing occurs as part of the diurnal cycle, where the large heat gains during the day are mostly removed at night. In regions of net heat gain, such as the tropics, this diurnal mixing is only possible with the penetrating part of the shortwave radiation heating the interior of the ocean. Without penetrative radiation, daytime heat gains at the surface exceed nighttime heat losses, and gravitational instabilities cannot develop.

The following simple model is a prototype for these physics and quantifies when the surface fluxes are conducive for night time mixing. Its application to TOGA COARE demonstrates the importance of penetrative radiation in controlling diurnal mixing and the evolution of surface temperatures.

Consider a one-dimensional model where the time rate of change of surface temperature and salinity are determined by the surface fluxes of heat and fresh water. The turbulent heat fluxes, long-wave radiation, and the fresh water flux apply at the surface of the ocean, and incident shortwave

radiation is absorbed within the surface ocean following a double exponential profile (Paulson and Simpson 1977). Mixing in the upper ocean occurs in response to static instability only and homogenizes the water column until it is gravitationally stable. Thus, convection is nonpenetrative, since the potential energy released during convection is dissipated rather than used to entrain waters from greater depths.

For the average surface fluxes of heat and fresh water observed during TOGA COARE (156°E, 1°45'S), this model yields a mixed-layer depth of 36 m, surprisingly close to other estimates of the surface layer depth in the western Pacific. This is strong evidence that mixing induced by penetrative radiation and night time cooling can be as important as wind-induced mixing.

To further illustrate this point, Figure 11 shows the time evolution of SST from COARE mooring, and the solution to the model outlined above. The time period corresponds to October 22, 1992, to March 3, 1993. Despite the great

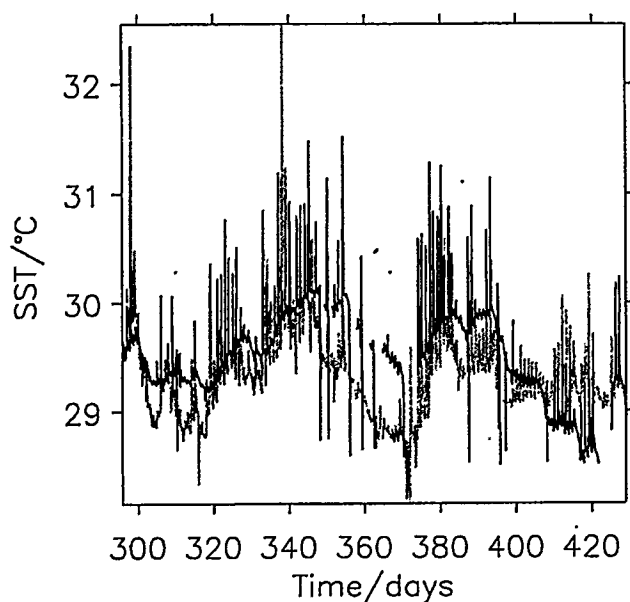


Figure 11. SSTs observed during TOGA COARE (156°E, 1°45' S) from October 22 to March 3, 1993 (gray line), and sea surface temperatures predicted from a model that includes only surface fluxes of heat and fresh water, penetrative radiation, and non-penetrative convection (solid line). The model was forced by observed fluxes of heat and fresh water. Gaps in the simulated surface temperature correspond to very low surface temperatures associated with rain storms.

simplicity of the model, and the complete lack of wind-induced mixing, the general SST tendencies are captured surprisingly well. The occasional short gaps in the simulation are due to very cool SSTs caused by heat losses and heavy precipitation that shut off nighttime mixing (Anderson et al. 1995). Largest differences between the simulation and observations are associated with strong winds, whose effect is not included in the model. The success of the simulation indicates that diurnal mixing due to penetrative radiation is a central mixing mechanism.

Neglect of penetrative radiation causes large errors in SST. Figure 12 shows observed, daily averaged SSTs, and those simulated by forcing a mixed layer model (Price et al. 1986) with surface fluxes observed during TOGA COARE. The model results with penetrative radiation reproduce the observed SSTs well, even though the one-dimensional mixed layer model underestimates the cooling during westerly wind burst. Disregard of the penetrative radiation yields a severe overestimation of surface temperatures by up to 3K during calm periods.

The importance of vertical mixing due to penetrative radiation that was hypothesized from a simulation with a coupled GCM was confirmed using observations. Thus, the relative magnitudes of incident radiation, and surface heat losses are of great importance for the heat budget of the ocean and planetary boundary layer and, ultimately, the processes of atmospheric convection and cloud formation.

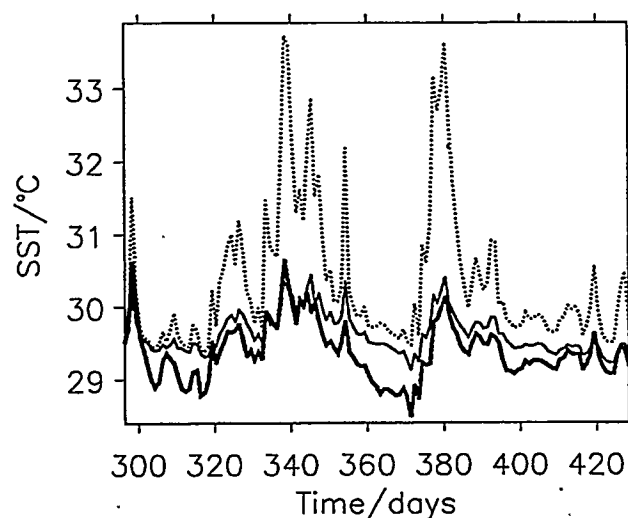


Figure 12. Daily averaged SSTs observed during TOGA COARE (solid thick line), simulated by a mixed-layer model (Price et al. 1986) with (solid thin line) and without (dotted line) penetrative radiation.

Conclusion

The results presented here suggest that, to zeroth order, the warm pool SST is governed by surface heat budget and mixed-layer dynamics. Furthermore, the surface and atmospheric radiation fluxes are strongly regulated by deep convection and associated cloudiness. Thus, the TWP is an ideal natural laboratory to address issues related to clouds-radiation-climate feedback processes. We list here the primary conclusions from our study:

- We have a long way to go before successfully modeling the excess cloud absorption over TWP or SGP. However, our study uncovered two flaws in the treatment of clouds in GCMs: 1) The neglect of cloud drops or ice particles with size greater than about $30\mu\text{m}$ (radius), a practice that tends to underestimate absorption; and 2) the plane parallel treatment of clouds tends to exaggerate the zenith angle dependence of cloud albedos, another reason for the underestimation of absorption in GCM clouds.
- Our GCM study suggests an observational period for ARESE experiment of at least six weeks and preferably two months.
- Comparison of observations of ECHAM3 and ECHAM4 and of observations from CEPEX and Oklahoma ARM site shows impressive simulations of aspects of the atmospheric moisture budget.
- Surface evaporation over TWP, where the annual mean SST is above 301°K , is strongly affected by deep convection through its interaction with the low level large-scale circulation. As a result, in the high SST and active convection regime, evaporation does not increase with SST, contrary to the predictions based on thermodynamic considerations alone.
- The importance of vertical mixing associated with penetrative radiation for TPW surface temperatures was confirmed using observations. Thus, the relative magnitudes of incident radiation and surface heat losses are of great importance for the heat budget of the TWP and planetary boundary layer and, ultimately, the processes of atmospheric convection and cloud formation.

We have used all available data to determine the long-term heat budget of the warm pool (Ramanathan et al. 1995). We need spatial and temporal variability of the radiation fluxes on a daily basis to relate the radiation fluxes to clouds, water vapor, and cloud microphysical properties. The three planned TWP ARCS stations along with 3-D analyzed fields will form the cornerstone of future study, which includes for the coming year:

- Use SGP radiation data in conjunction with ARESE data to unravel the physics behind the excess solar absorption.
- Use high spatial (Oklahoma) and high temporal (1 hour) resolution of GMS and GOES data to estimate the scale dependence of cloud radiation properties and develop a scale dependent cloud parameterization scheme for deep convective-cirrus clouds in TWP.
- Couple the convective parameterization with the cirrus cloud radiative parameterization in a GCM to unravel the role of convection and clouds and their associated physical, dynamical and radiative processes in the TWP heat budget.
- Investigate the coupling between convection, surface radiative heat flux, and evaporative cooling in interannual anomalies of the TWP using observations, atmospheric and coupled atmosphere-ocean circulation models.
- Investigate the dependence of the long-term average surface radiation and turbulent heat fluxes and SSTs on the specification of high-frequency (i.e., diurnal) forcing using a combination of an ocean mixed layer model and an atmospheric column model.

Acknowledgments

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Evaluation of Cloud Prediction and Determination of Critical Relative Humidity for a Mesoscale Numerical Weather Prediction Model

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Predictions of cloud occurrence and vertical location from the Pennsylvania State University/National Center for Atmospheric Research nonhydrostatic mesoscale model (MM5) were evaluated statistically using cloud observations obtained at Coffeyville, Kansas, as part of the Second International Satellite Cloud Climatology Project Regional Experiment campaign. Seventeen cases were selected for simulation during a November-December 1991 field study. MM5 was used to produce two sets of 36-km simulations, one with and one without four-dimensional data assimilation (FDDA), and a set of 12-km simulations without FDDA, but nested within the 36-km FDDA runs. The 36-km runs had 20 layers and the 12-km runs had 25 layers in the vertical direction. The 36-km simulation with (without) FDDA will be designated as CNTL (CFDA), while the 12-km simulation will be referred to as FGM. Validation data were obtained from the PSU 94-GHz cloud radar.

The comparison study of the bias score (Bs), threat score (Ts), and categorized forecast (α) described by Guo (1994) for all cases shows that cloudiness is predicted well in the low and middle layers and that FDDA improves the skill of the cloud prediction in the 36-km simulations (Table 1). The 12-km fine mesh simulations, with no FDDA, but using boundary conditions from the 36-km FDDA simulations, have skill very similar to the 36-km FDDA runs. Overall, however, the standard relative-humidity-dependent cloud diagnosis in MM5 (Benjamin 1983) is found to produce a non-negligible bias that over-forecasts cloudiness in the low and high layers, particularly for the 12-km simulations. For example, Figure 1 shows that all three experiments have a strong positive bias for low clouds, such that a bias score of 1.0 (perfect) is associated with relative humidities of 83% - 87%.

Table 1. MM5 diagnosed skill scores for clouds using the standard RH_c (Benjamin 1983).

Low Cloud	Skill Scores		
	Bs	Ts	α
CNTL	1.30	0.54	0.68
CFDA	1.40	0.58	0.70
FGM	1.50	0.58	0.69
Middle Cloud	Bs	Ts	α
CNTL	0.85	0.48	0.71
FGM	0.95	0.52	0.73
High Cloud	Bs	Ts	α
CNTL	1.19	0.42	0.53
CFDA	1.09	0.43	0.56
FGM	1.14	0.44	0.56
Standard RH_c (with respect to water)			
	Low Cloud	Middle Cloud	High Cloud
CNTL	75%	75%	60%
CFDA	75%	75%	60%
FGM	75%	75%	60%

These errors in forecasted cloud can be attributed to an improper specification of the critical relative humidity (RH_c) used to parameterize cloud occurrence in this model at these resolutions. In the original MM5, RH_c is not a function of grid scale. Results verify the hypothesis that, for mesoscale applications, RH_c should be defined as a function of model grid resolution, with higher values of RH_c used for

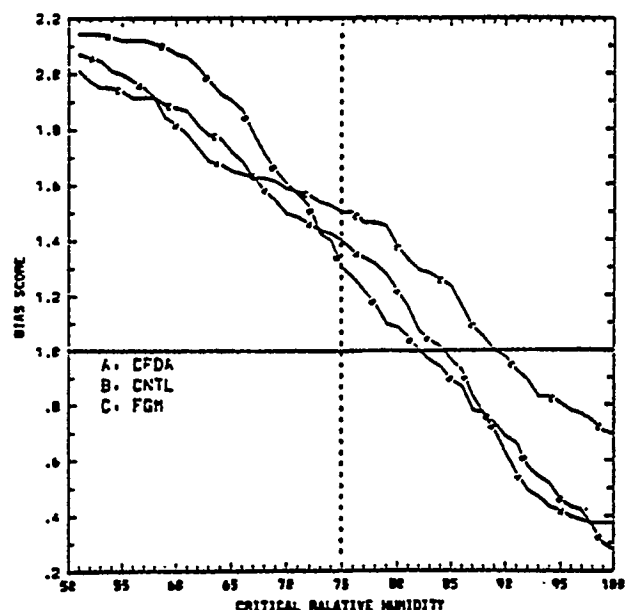


Figure 1. Evaluation of bias scores for low clouds at Coffeyville (all cases) based on standard critical relative humidity, Curve A = exp. CFDA; Curve B = exp. CNTL; Curve C = exp. FGM. Vertical dashed line is standard $RH_c = 75\%$, horizontal line is perfect score = 1.0.

finer resolution. New functional (optimal) relationships are determined for both 36-km and 12-km grids (compare Tables 1 and 2) and sensitivity analysis shows improved predictive skill for clouds in MM5. For example, Figure 2 indicates that the bias score remains consistently close to 1.0 for all levels in an evaluation that groups the output into eleven validation layers (data are not applicable above 9 km in the stratosphere). Finally, analysis shows that the model predictions of cloud fields have a greater correlation with

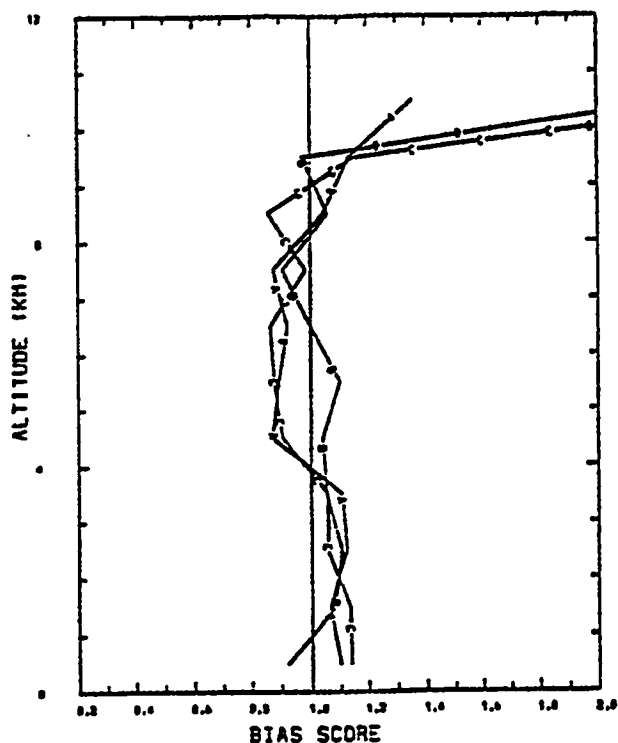


Figure 2. Bias score as a function of height calculated from 11-layer optimized values of critical relative humidity, RH_c , Curve A = exp. CFDA; Curve B = exp. CNTL; Curve C = exp. FGM. Vertical line is perfect score = 1.0.

the radar-observed clouds (assumed to have very high quality) than do the Real-Time Nephanalysis (RTNEPH) cloud fields from the U.S. Air Force for the same period (Table 2). That is, the optimized MM5 has higher skill than RTNEPH for low, middle, and high clouds.

Table 2. Optimized MM5 diagnostic skill scores for clouds.

Low Cloud	Skill Scores		
	Bs	Ts	α
CNTL	0.96	0.58	0.76
CFDA	1.04	0.59	0.76
FGM	1.00	0.58	0.75
RTNEPH	0.72	0.55	0.76
Middle Cloud	Bs	Ts	α
CNTL	0.99	0.51	0.72
CFDA	1.00	0.52	0.72
FGM	1.03	0.55	0.74
RTNEPH	1.14	0.45	0.62
High Cloud	Bs	Ts	α
CNRL	1.02	0.41	0.55
CFDA	0.96	0.44	0.60
FGM	1.05	0.43	0.57
RTNEPH	0.84	0.34	0.52
	Standard RH _c (with respect r=to water)		
	Low Cloud	Middle Cloud	High Cloud
CNTL	83%	71%	65%
CFDA	83%	71%	65%
FGM	89%	72%	65%

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A Shallow Convection Parameterization for the Non-Hydrostatic MM5 Mesoscale Model

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A shallow convection parameterization suitable for the non-hydrostatic Pennsylvania State University (PSU)/National Center for Atmospheric Research MM5 mesoscale model is being developed at PSU. The parameterization is based on parcel perturbation theory developed in conjunction with a 1-D Mellor-Yamada 1.5-order planetary boundary layer (PBL) scheme and the Kain-Fritsch (KF) deep convection model (Kain and Fritsch 1990).

Development is done within a 1-D framework to allow efficient experimentation in a prescribed (constant or time-dependent) large-scale environmental state. Later, the 1-D model will be introduced into the 3-D MM5 for fully interactive testing and evaluation. The first step was to incorporate the KF scheme into the Gayno-Seaman PBL. This composite PBL-convective model was run over Oklahoma for the convective case of June 10, 1993, beginning at 1200 Greenwich Mean Time (GMT). The initial profile (not shown) had an inversion in the lowest 60 mb. Figure 1 shows the simulated mixed layer at 1800 GMT (+6 h). By 2100 GMT, after the inversion is eliminated, parcels in the mixed layer with small positive temperature perturbations (1°C) and moisture perturbations (1 g kg^{-1}) become unstable (Figure 2). Convective available potential energy is then released by activating the KF scheme. Although the full impact of the convection on the environment cannot be represented in the 1-D model, the sounding is stabilized by evaporative cooling in moist downdrafts, thus halting the convection and drying the middle levels (not shown).

The model assumes shallow convection is mostly surface-based. Key variables include the cloud-parcel perturbation temperature, perturbation mixing ratio, and the vertical velocity at cloud base and the cloud-base height. For the first stage of the development, over land, we assume that the parcel perturbation temperature (T) and mixing ratio (q) are given by the simulated surface-layer values (level KL). The cloud-base height, Z_c , after Wang (1993), is given by

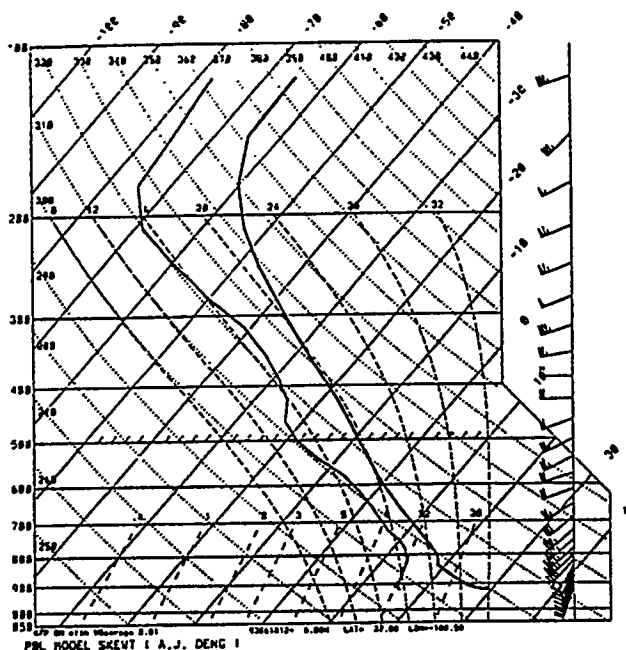


Figure 1. Simulation at 1800 GMT, +6 hrs.

$$Z_c = \frac{T_{KL} \ln[q_{KL}/q_{SO}]}{\left[1 - \frac{q_{SO}}{0.622} \right] \frac{Lg}{R_v T_{KL} C_p} - \frac{g}{R_d}}$$

Between a non-disturbed state and the initiation of deep convection, three states of shallow convection are hypothesized: 1) turbulent PBL, but no cloud (no condensation); 2) shallow stable cloud (does not reach level of free convection, LFC); and 3) deep non-precipitating cloud (reaches LFC, but too shallow to trigger KF parameterization - less than 4 km deep). As clouds grow through Types 1-3, they transition smoothly to KF deep convection.

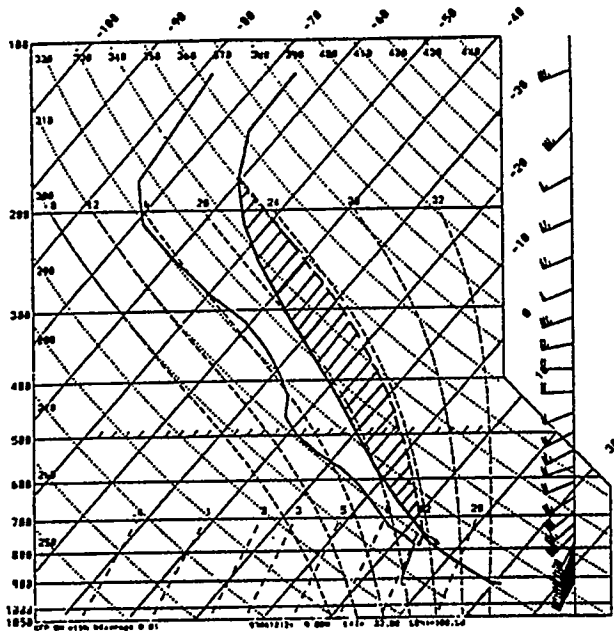


Figure 2. Simulation at 2100 GMT, +9 hrs., showing perturbation temperature and moisture. Hatched area indicated convective available potential energy (CAPE).

Current work involves testing and evaluation of the vertical-velocity calculations. The vertical velocity of a parcel is determined as the sum of five contributing factors: 1) sub-grid scale PBL turbulence due to surface heating, 2) resolved-scale vertical velocity at cloud base, 3) sub-grid scale effects due to land-use inhomogeneities, 4) sub-grid scale effects detectable from the resolved-scale terrain, and 5) sub-grid scale effects undetectable from the resolved-scale terrain. The dominant contribution to the parcel's cloud-base vertical velocity normally is from the turbulent component, $W_T' = 0.816 E_{\max}^{0.5}$, where $E_{\max}$ is the maximum turbulent kinetic energy in the PBL. An example of the evolution of W_T' is shown in Figure 3. Along with the temperature and moisture perturbations and the cloud-base calculation, this will enable determination of equilibrium cloud tops for undiluted clouds. Subsequent work will add entrainment with the environment.

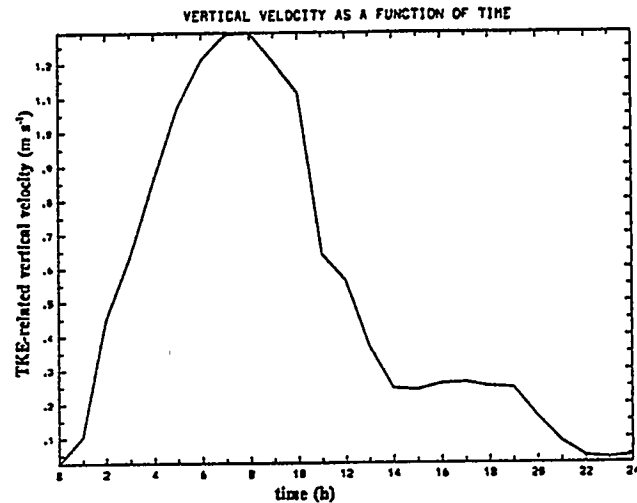


Figure 3. Turbulence-induced vertical velocity perturbations, W_T' , calculated for June 10, 1993.

When completed, the model will allow prediction of cloud coverage, cloud base and top, mass flux at cloud base, distribution of detrained cloud mass, and liquid water content. These parameters will better represent effects of non-precipitating cloud on radiation processes (and other meteorological or other chemical impacts) in numerical models than is currently possible. The scheme will also distinguish between cumulus and stratocumulus clouds and be applicable in continental and marine environments.

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Dual-Wavelength Millimeter-Wave Radar Measurements of Cirrus Clouds

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Introduction

In April 1994, the University of Massachusetts' 33-GHz/95-GHz Cloud Profiling Radar System (CPRS) participated in the multi-sensor Remote Cloud Sensing (RCS) Intensive Operation Period (IOP), which was conducted at the Southern Great Plains Cloud and Radiation Testbed (CART). During the 3-week experiment, CPRS measured a variety of cloud types and severe weather. In the context of global warming, the most significant measurements are dual-frequency observations of cirrus clouds, which may eventually be used to estimate ice crystal size and shape. Much of the cirrus data collected with CPRS show differences between 33-GHz and 95-GHz reflectivity measurements that are correlated with Doppler estimates of fall velocity. Because of the small range of reflectivity differences, a precise calibration of the radar is required and differential attenuation must also be removed from the data. Depolarization, which is an indicator of crystal shape, was also observed in several clouds (Matrosov 1991; Tang and Aydin 1995). In this abstract we present examples of Mie scattering from cirrus and estimates of differential

attenuation due to water vapor and oxygen that were derived from CART radiosonde measurements.

Cirrus Observations

Figures 1a and 1b plot models of the difference between 33-GHz and 95-GHz reflectivity and the mean velocity in cirrus clouds. The quantity used to describe differences in 33-GHz and 95-GHz reflectivity is referred to as the dual wavelength ratio (DWR) and is defined as

$$\text{DWR} = \text{dBZ}_{33} - \text{dBZ}_{95}.$$

DWR is generated in two ways. The first source of DWR is backscatter from larger particles that fall into the Mie region at either 95 GHz or at both frequencies used. DWR is a function of particle shape, size, and orientation (Matrosov 1991, 1995). The second source of DWR is frequency dependent attenuation by the atmosphere and clouds.

Although external calibrations of the radar were made with a trihedral corner reflector, differential attenuation, both from

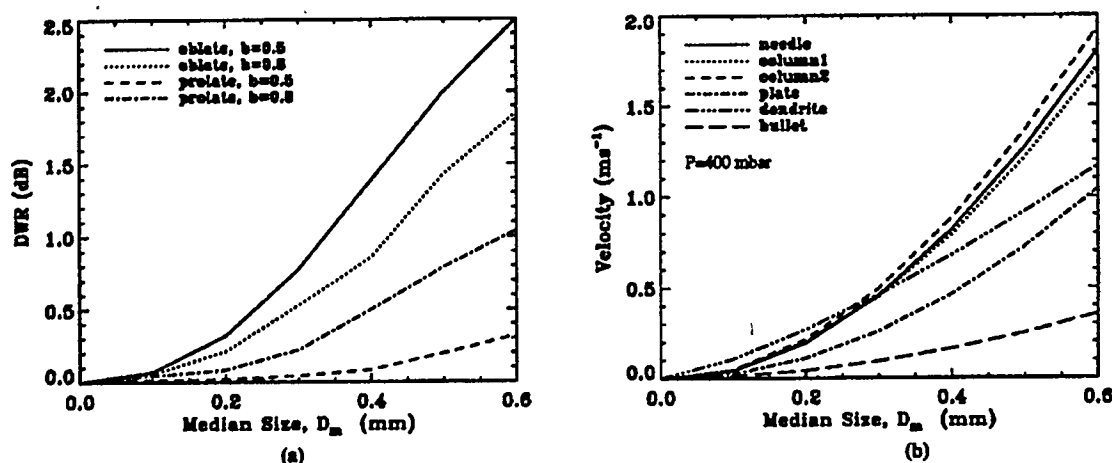


Figure 1. (a) Dual-wavelength ratio (DWR) models for gamma distributions of spheroid (adapted from Matrosov 1995) and (b) Mean velocity at 400 mbar for gamma distributions of spheroids.

intervening clouds and from the moist atmosphere itself, contributes a substantial bias to DWR measurements and must be taken into consideration. Because of the low values of DWR produced by Mie scattering, an important step in the analysis of DWR measurements is removal of differential extinction experienced by the radar signal in propagating between the ground and cloud base. The contribution to differential attenuation from water vapor and oxygen is plotted in Figure 2 at 1-km intervals.

Figures 3a and 3b show colocated 33-GHz and 95-GHz reflectivity measurements taken on April 15th, 1994. Gaseous extinction has been removed from these measurements, and there are no lower clouds present. Cross-sections of measurements are shown in Figure 4. Although these plots show values of DWR and velocity that lie within the range of values predicted by the models in Figures 1a and 1b, they are ambiguous because crystal shape and particle orientation are not known.

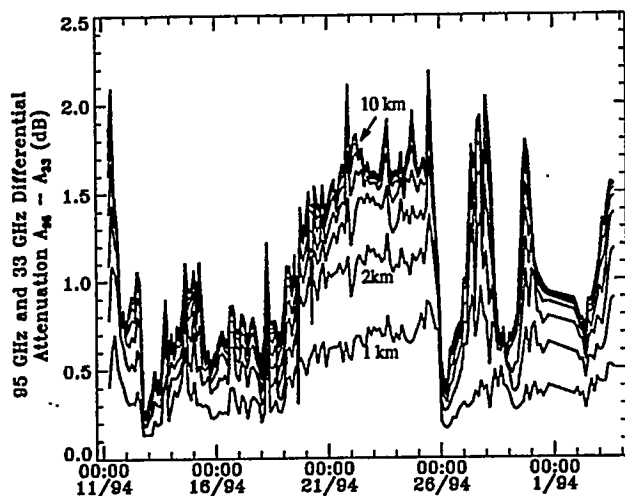


Figure 2. CART radiosonde-derived estimates of one-way differential attenuation between ground and indicated height for 1-km steps in altitude.

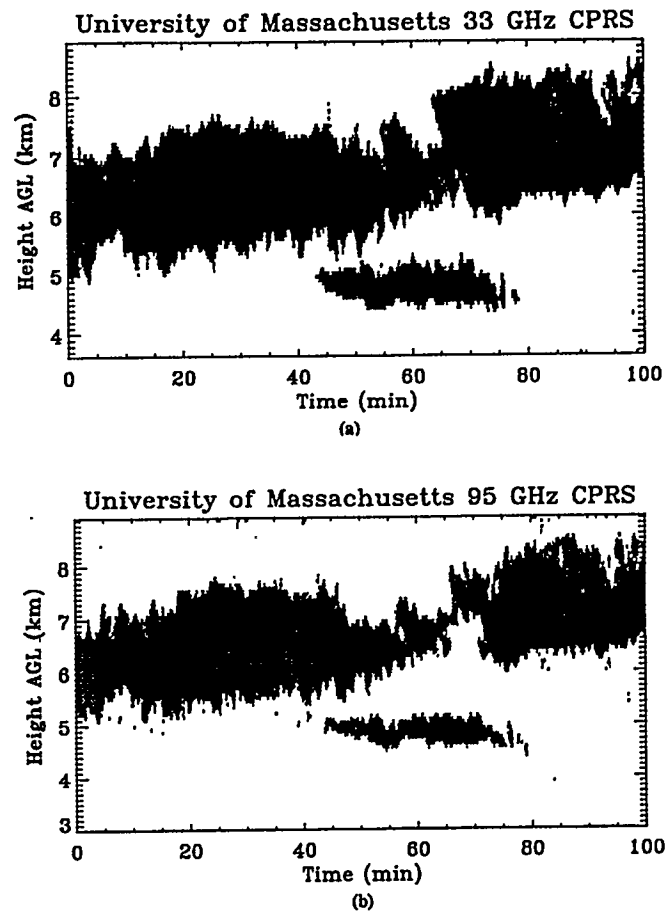


Figure 3. Colocated reflectivity measurements from April 15, 1994. dBZe values on relative scale at (a) 33 GHz and (b) 95 GHz.

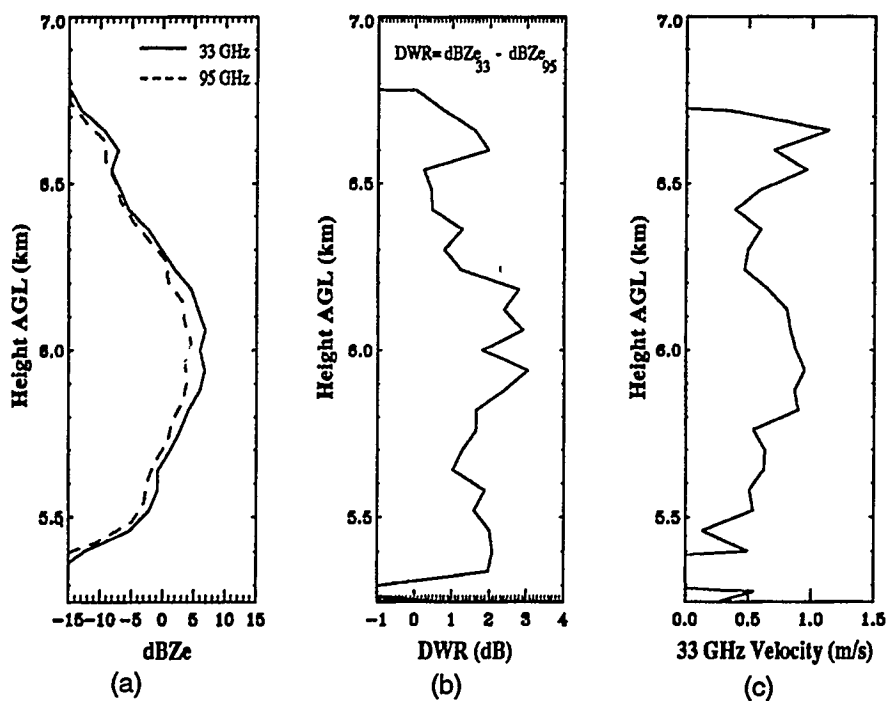


Figure 4. Profiles of (a) reflectivity, (b) DWR, and (c) fall velocity at 0016 GMT.

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Remote Cloud Sensing Intensive Observation Period (RCS-IOP) Millimeter-Wave Radar Calibration and Data Intercomparison

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Introduction

During April 1994, the University of Massachusetts (UMass) and the Pennsylvania State University (Penn State) fielded two millimeter-wave atmospheric radars in the Atmospheric Radiation Measurement Remote Cloud Sensing Intensive Operation Period (RCS-IOP) experiment. The UMass Cloud Profiling Radar System (CPRS) operates simultaneously at 33.12 GHz and 94.92 GHz through a single antenna (Sekelsky and McIntosh 1995). The Penn State radar operates at 93.95 GHz and has separate transmitting and receiving antennas (Clothiaux et al. 1995). The two systems were separated by approximately 75 meters and simultaneously observed a variety of cloud types at vertical incidence over the course of the experiment.

This abstract presents some initial results from our calibration efforts. An absolute calibration of the UMass radar was made from radar measurements of a trihedral corner reflector, which has a known radar cross-section. A relative calibration of between the Penn State and UMass radars is made from the statistical comparison of zenith pointing measurements of low altitude liquid clouds. Attenuation is removed with the aid of radiosonde data, and the difference in the calibration between the UMass and Penn State radars is determined by comparing the ratio of 94-GHz and 95-GHz reflectivity values to a model that accounts for parallax effects of the two antennas used in the Penn State system.

Analysis of Cloud Data

Figure 1 shows images of reflectivity values collected on April 11, 1994. Averaging time is 8 seconds, and range gates are spaced 75 meters apart. These data are unique in that the cloud was composed of small water droplets and because temperature and radar depolarization measurements suggest that no insects were present in the cloud volume sampled. Scattering from insects can generate significant biases in measurements at different frequencies and, possibly, in measurements at the same frequency in different locations if the effect cannot be statistically removed by averaging. Uncorrected reflectivity values averaged from 1900 GMT to 1946 GMT are plotted in Figure 2. Although the 33-GHz and 95-GHz data are very similar, difference in extinction rates though the cloud cause the 33-GHz and 95-GHz dBZe curves to diverge. CART radiosonde data and 33-GHz reflectivity measurements are used to correct for extinction by water vapor, oxygen, and liquid cloud droplets.

A substantial difference between UMass and Penn State estimates of reflectivity is seen at the lowest range gate and decreases with height. The large differences in reflectivity values can be accounted for by modeling a well-known parallax effect in bistatic radars (Ulaby et al. 1982) for various offsets in pointing angle between the two antennas. A numerical simulation was performed to determine sample volume overlap for the transmitting and receiving antennas

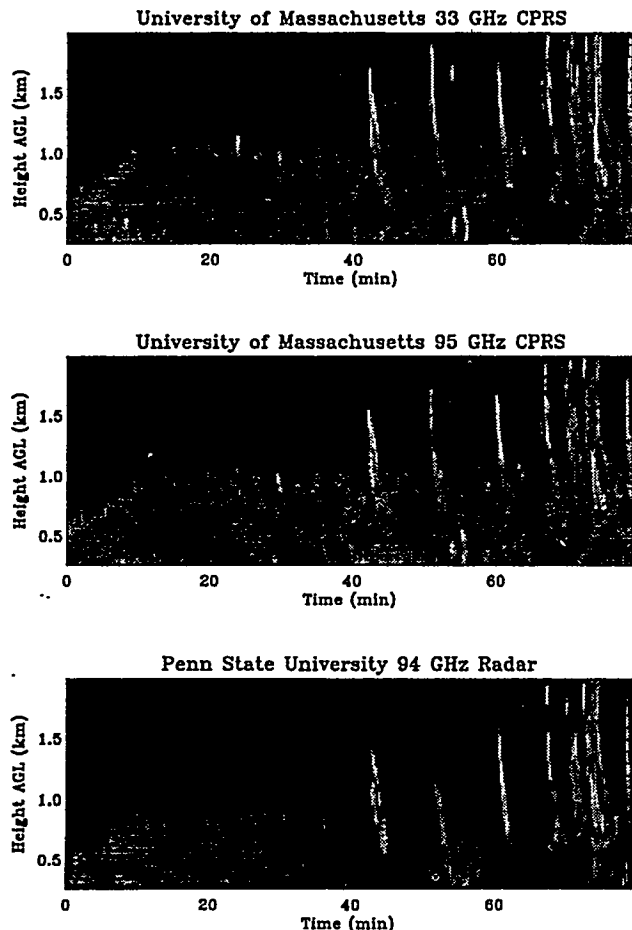


Figure 1. Liquid water cloud measurements collected with (a) CPRS 33-GHz radar, (b) CPRS 95-GHz radar, and (c) Penn State 94-GHz radar.

of the 94-GHz radar as a function of range and offset of one antenna from the boresight direction of the other. Figure 3 compares model data with measured ratios of $Z_{94.92}/Z_{93.95}$ (averaged over 46 minutes). A least mean square fit of the measured data to simulations was performed for 0.001° angular increments. Results indicate angular offsets of 0.052° and 0.068° in the pointing directions of the two antennas. A mean difference between the UMass and Penn State calibrations of 0.98 dB was determined from the mean offset of the $Z_{94.92}/Z_{93.95}$ curve from the simulated curve. Thus, the actual difference in calibration between the Penn State and UMass radars is -0.98 dB for this dataset after attenuation and parallax errors are removed.

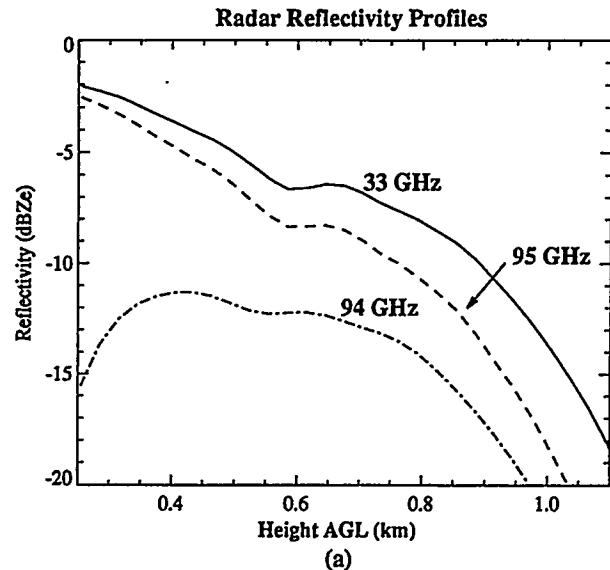


Figure 2. Vertical profiles of reflectivity averaged between 1900 and 1946 GMT without attenuation correction.

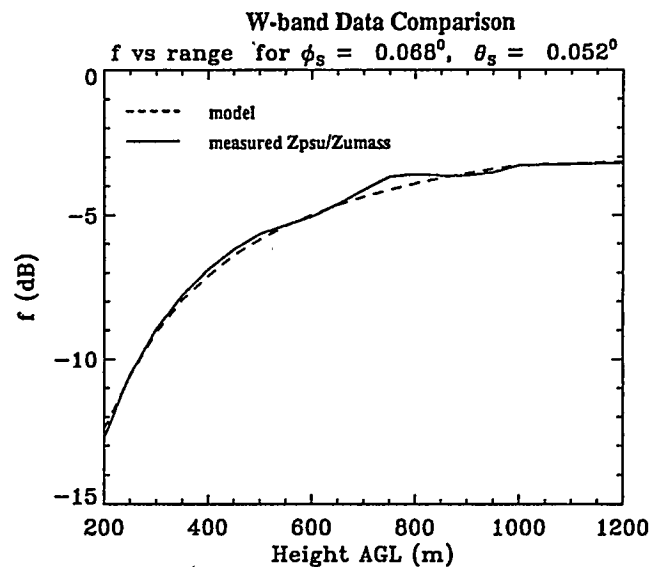


Figure 3. Ratio of non-parallel-pointing dual-antenna reflectivity values versus range to that for single antenna.

Summary of Liquid Cloud Measurements

Figures 4 and 5 summarize liquid cloud reflectivity measurements from the UMass and Penn State radars in the form of histograms. Figure 4 shows obvious differences in the positions of the peak and in the range of values spanned in the raw reflectivity measurements. Figure 5 plots the same data with much better agreement after corrections for attenuation are applied to each measurement set and after correcting the Penn State measurements for parallax. The analysis illustrates some important issues involved in comparing measurements collected with different sensors.

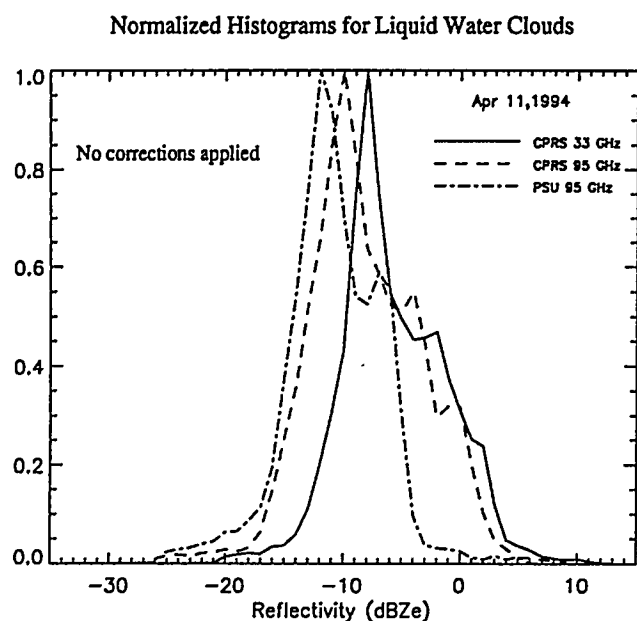


Figure 4. Uncorrected histograms of liquid water cloud measurements. Bin size is 1 dB.

Normalized Histograms for Liquid Water Clouds

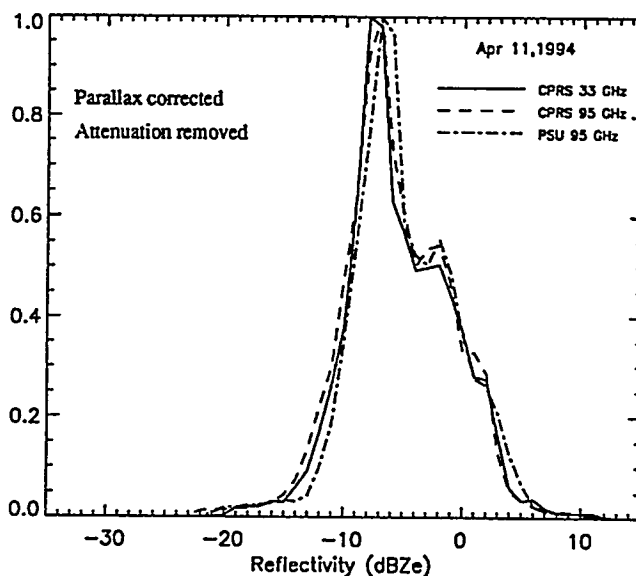


Figure 5. Histograms of liquid water cloud measurements after correction for attenuation and parallax. Bin size is 1 dB.

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Balloon-Borne Radiometer Profiler: Field Observations

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Project Objective

The objective of this project is to develop the capability of making routine soundings of broadband radiative fluxes and radiative flux divergences to heights of 1500 m above ground level (AGL). Using this capability, soundings will be made under different cloudiness conditions and will be compared to radiative transfer models.

The Radiometer Profiler

To meet these objectives, radiometers are carried on a stabilized platform (see Figure 1) in a harness inserted in the tetherline of a tethered balloon meteorological sounding system, approximately 30 m below the balloon. The platform is leveled by an automatic control loop stabilization system in which the level sensors are two single-axis, solid-state

rotation rate sensors. These sensors use vibrating micro-machined quartz tuning forks to sense angular rate. Rate sensor drift is corrected periodically by a long-term-average measurement of the earth's gravity vector as sensed using a pair of single-axis linear accelerometers.

The rate sensor/accelerometer leveling system is a new design completed in 1994. Other modifications were made to the prototype platform: the automatic control loop was modified to use the new sensors, the breadboarded electronic circuits were made into finished printed circuit boards, new radiometers (total hemispherical radiometers and pyranometers) were installed and tested, improved pulley housings were installed, the platform structure was stiffened, and platform components were miniaturized and relocated in a weatherproof housing. Initial field tests were also performed.

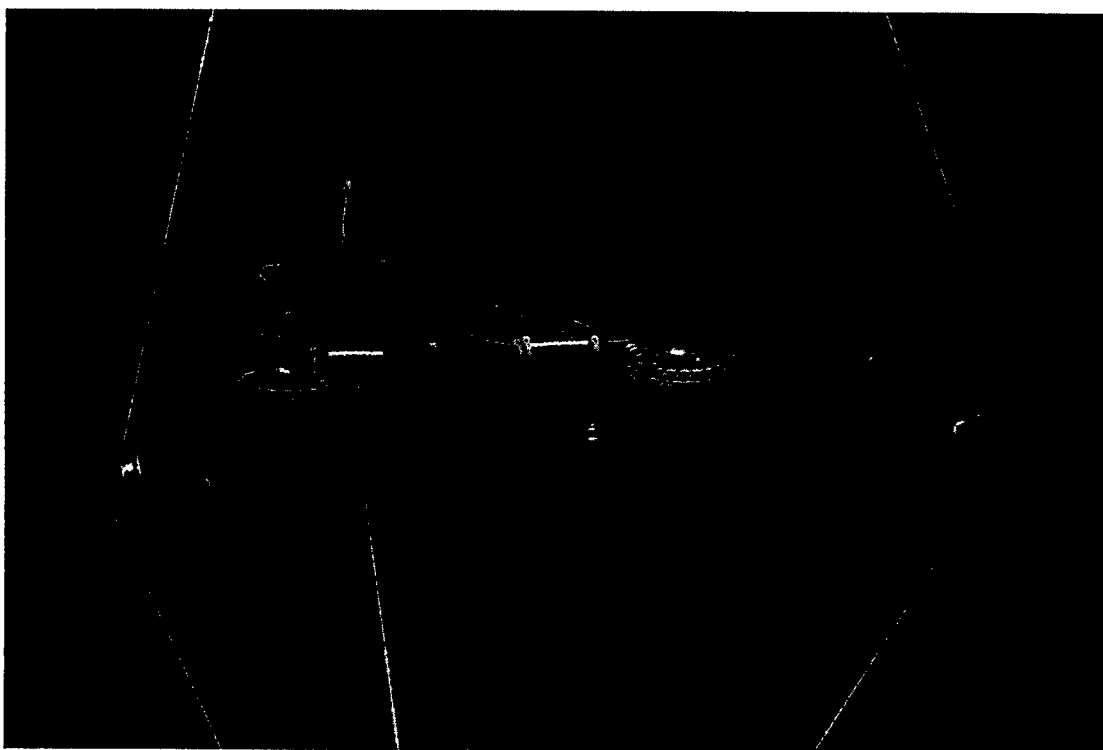


Figure 1. Photograph of prototype self-leveling balloon-borne radiometer platform.

Leveling Measurements

The radiometer platform stabilizes itself using feedback loops involving a pair of accelerometers and rate gyros active on axes separated by 60 degrees. The accelerometers are sensitive to lateral accelerations as well as the gravitation acceleration component when the platform is not level. They are, therefore, lowpass-filtered for use as a mean level indicator. The rate gyros measure angular velocity, but become insensitive to low-amplitude long-period rotations. They are high-pass filtered and are used to measure oscillations of the platform about its mean orientation.

It is important to know the inherent accuracy of the leveling system components, which can be estimated as follows. For the rate gyros, the full-scale ($\pm 5V$) is 100° s^{-1} . The analog outputs from these sensors are sampled using a 12-bit A/D converter, yielding a sampling resolution $\theta_{\text{res}} = 0.024^\circ \text{ s}^{-1}$. The angular displacement at a given frequency can be represented by

$$\theta = \theta_0 \sin(2\pi ft)$$

where θ_0 is the displacement amplitude. The rotation rate is then

$$\dot{\theta} = 2\pi f \theta_0 \cos(2\pi ft)$$

Therefore, the displacement resolution at a given frequency (or period T) is

$$\theta_{\text{res}} = \frac{\dot{\theta}_{\text{res}}}{2\pi f} = \frac{\dot{\theta}_{\text{res}} T}{2\pi}$$

For a nominal period of 60 s, this gives $\theta_{\text{res}} = 0.06^\circ$, which is well within the accuracy required to validate leveling requirements of 0.5° rms deviation from zenith. For the accelerometers, which measure mean departures from level, the acceleration due to an angular displacement is, to a first approximation,

$$a = g \sin \theta$$

The full-scale range of 4 g through a 12-bit A/D converter gives a resolution of 0.001 m s^{-2} , which corresponds through the above relation to 0.06° . As noted previously, the accelerometer measurements of angular displacement can be contaminated by lateral accelerations. As an estimate of this effect, consider that a reasonable maximum platform lateral velocity change over 60 s of 3 m s^{-1} yields an

average acceleration of 0.05 m s^{-2} . This corresponds to an error in the measurement of average displacement of 0.29° , which is comparable to mean leveling requirement for the platform. Thus, as long as the balloon is reasonably stable, the accelerometers are adequate to establish the required mean leveling.

Results of Flight Tests

The new rate sensors proved to be very sensitive, and the platform sides had to be stiffened to reduce the sensitivity of the control system to flexures of the structural members of the platform. These structural modifications increased the mass of the platform to about 2.5 kg. Flight experience showed that a low-speed 7.5 m^3 balloon should have adequate lift to carry the platform to its design altitude of 1500 m AGL. A 7 m^3 high-speed balloon with rigid tail fins should allow flights to this altitude in winds up to 35 kts.

An initial series of balloon-borne test flights was performed under waivers obtained from the Federal Aviation Administration. Figures 2 and 3 present the test results during a period of continuous ascent with a low-lift balloon on June 28, 1994. During this flight segment, the mean tilt of the radiometer profiler from horizontal on axis 1 was 0.5° with a standard deviation of 1.2° . The tilt on axis 2 was -1.7° with a standard deviation of 0.6° . We should be able to improve on these initial results with better pre-flight leveling adjustments and with control loop filters and sensitivity adjustments. Further test flights will concentrate on stabilization system performance, better sensitivity matching of the two axes, flight profiles (continuous versus stepped ascents), testing of a new high-speed balloon, calibration and testing of the radiometers, development of operating procedures, etc.

Field experiments were used to test the stable platform radiometers against transfer standards. Discrepancies between the airborne radiometers and the standard radiometers suggest that the airborne radiometers should, in the future, be calibrated while they are in position on the Radiometer Profiler; this will allow us to correct for the thermal conductivity of the platform radiometer mounts.

This project is cooperating with the National Science Foundation and the Office of Naval Research-funded investigators as part of the Surface Heat Balance of the Arctic (SHEBA) Program to develop new tethered balloon technologies for atmospheric research in the Arctic. Plans are being made to make data flights with the Radiometer Profiler at Point Barrow, Alaska, in 287 the summer of 1995.

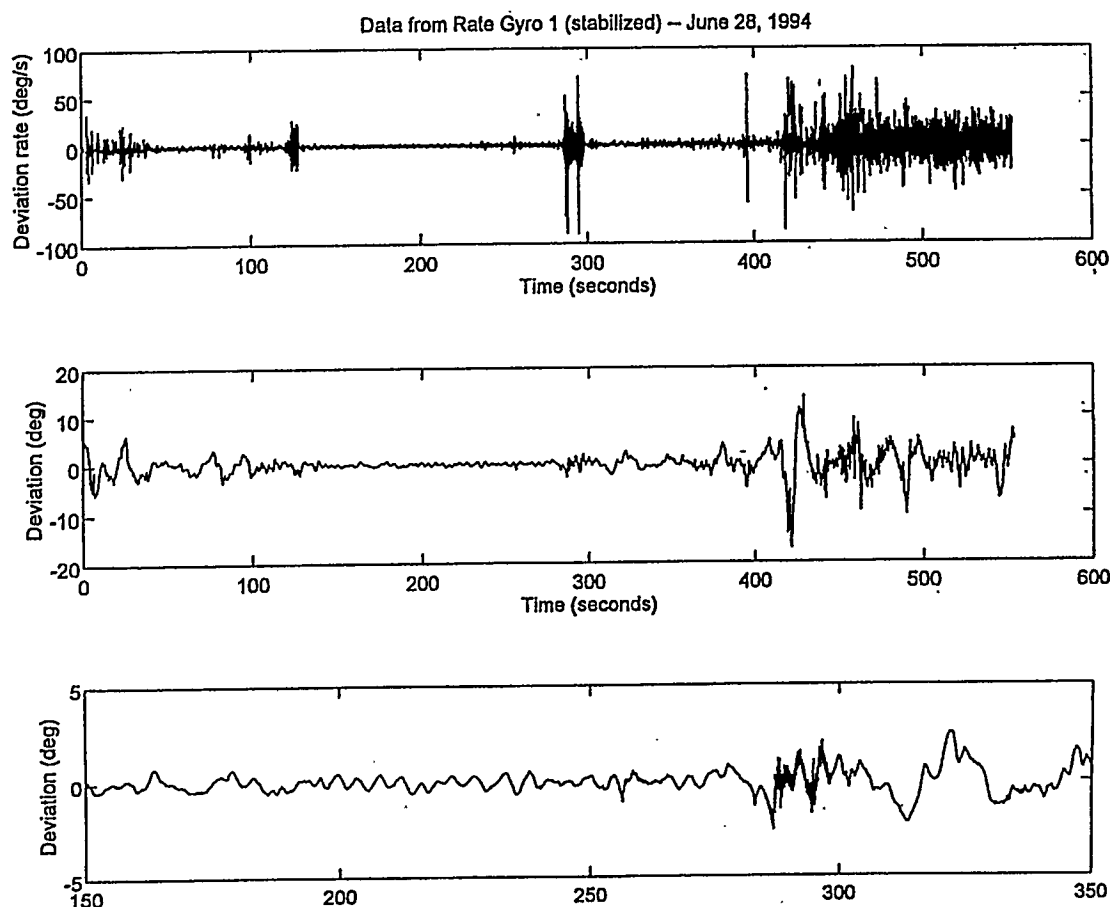


Figure 2. Time series from rate gyro 1. These data were collected during a steady ascent. The deviation is the instantaneous departure from level. It is obtained by removing periods longer than 30 s in the deviation rate and then integrating over time. The first 130 s of the flight were affected by wake turbulence from nearby buildings. The increased unsteadiness about 400 s into the flight corresponds, we believe, to early battery failure. The period between 150 s and 350 s is the period for which statistics have been calculated.

Acknowledgment

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Satellite Data Sets for the Atmospheric Radiation Measurement (ARM) Program

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Satellite measurement is an important component in the Atmospheric Radiation Measurement (ARM) Program data stream. Currently the real-time, polar-orbit and geostationary satellite data received at the ARM data processing facilities are provided by the SeaSpace Corporation. These data sets include the Advanced Very High Resolution Radiometer (AVHRR) and the TIROS Operational Vertical Sounder (TOVS) measurements from the National Oceanic and Atmospheric Administration (NOAA) satellites and measurements from Geostationary Operational Environmental Satellites (GOES)-7/GOES-8. The AVHRR acquires visible, near-infrared, and thermal infrared data through five channels at the horizontal resolution of approximately 1 km at nadir. The TOVS data produce vertical profiles of temperature, moisture, and wind. With the finishing of data-set transition from GOES-7 to GOES-8, all the five channel measurements of GOES-8 at the horizontal resolution of 4 km are provided to the ARM data processing facilities for the coverage of the continental United States. In addition, full resolution GOES-8 measurements are made available for the 1000 x 1000-km area centered at the ARM Southern Great Plains site. These satellite data sets have been widely used by ARM science team members to derive cloud-top altitude, cloud cover, snow and ice cover, surface temperature, and vertical profiles of temperature, water vapor, and wind, as well as to obtain continuous observation of weather features needed to predict, detect, and track severe weather. In addition to producing satellite data products directly for the ARM data processing facilities, we also examine the application of satellite data from a variety of satellite platforms. The emphasis is on the integration of different data sources, including data from satellite and non-satellite sources, for meteorological studies.

Besides data from the NOAA and GOES satellites, measurements from the Defense Meteorological Satellite Program (DMSP) are gaining recognition in the meteorological research and operational communities. The sensors carried aboard the DMSP satellites include the Operation Line Scanner (OLS), the Special Sensor Microwave Imager (SSM/I), the Special Sensor Microwave Temperature Sounder (SSM/T1), and the Special Sensor Microwave Water Vapor Sounder (SSM/T2). The OLS sensor is a visible/infrared radiometer, and the SSM/I, SSM/T1, and SSM/T2 sensors are microwave radiometers. The SSM/I is primarily used to estimate various geophysical and meteorological properties of the earth and atmosphere. These include total columnar atmospheric water vapor and liquid water, ocean surface wind speed, sea ice concentration, etc. The SSM/T1 has seven channels with the spot size approximately 175 km at nadir. The measurements of these channels can be converted to temperature, geostrophic wind, and thermal wind data at 15 standard pressure levels (1000 mb through 10 mb). The SSM/T2 sensor has five microwave channels. The spot size is approximately 45 km at nadir. The channel measurements can be converted to specific humidity, relative humidity, or dew-point temperature at 6 standard pressure levels (1000 mb through 300 mb). We are examining the temperature and water vapor retrievals with comparison to conventional radiosonde measurements. Considering the advantages of being capable of observing large areas cost-effectively and regularly through satellite remote sensing, it can be anticipated that satellite observation will play a more important role in the future in meteorological researches.

FASCODE for the Environment (FASE)

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Abstract

The Optical Physics Division of the Phillips Laboratory, with support from the U.S. Department of Energy (DOE) Atmospheric Radiation Measurement (ARM) Program, is developing a state-of-the-art line-by-line atmospheric radiative transfer model as the successor to FASCODE. The goal of this project is to create a computationally efficient model which contains the most up-to-date atmospheric physics. The new model, known as FASCODE for the Environment, or "FASE", will combine the best features of FASCODE and LBLRTM, the DOE's standard radiative transfer model. FASE will also contain new features such as new cross-sections for heavy molecules, an improved solar irradiance model, and improvements to the Schumann-Runge bands and continuum. The code will be optimized for vectorized and/or parallel processing, put under configuration control for easy maintenance, and structured into separate modules for each function: atmospheric profiles, layer optical properties, radiative transfer, multiple-scattering, etc. This modular structure will allow for increased flexibility and easy customization of the code for specialized applications, such as a forward model for iterative inversion algorithms. Ease-of-use will be enhanced with improved input control structures and documentation to accommodate the needs of novice and advanced users. This paper addresses changes which have been made to FASCODE and LBLRTM to create FASE, and gives an overview of the modular structure and its capabilities.

Introduction

The goal of the FASCODE for the Environment (FASE) program is to create an atmospheric radiance and transmittance model which is user-friendly and contains the

latest atmospheric physics. The prime focus of the program is to make available to the atmospheric spectroscopy community the results of on-going work sponsored by the U.S. Department of Energy (DOE), while incorporating the results of continuing research and development at the Air Force Phillips Laboratory. In addition, advances from the wider radiative transfer community will be incorporated where appropriate within program constraints and resources.

FASE has been developed by combining features from the line-by-line radiative transfer codes of the Air Force Phillips Laboratory (FASCODE) (Anderson et al. 1994) and the Department of Energy (LBLRTM) (Clough 1993). Both of these models were derived from FASCOD3 which was based on FASCOD1B, a four-function line-by-line code developed by Clough and Kneizys (Clough and Kneizys 1979). Our basic approach to developing FASE is severalfold: 1) to modify the overall program structure of FASCODE/LBLRTM so as to improve the flexibility and maintainability of the code without significant re-coding, 2) to incorporate a number of coding improvements (also to benefit the flexibility and maintainability), 3) to improve the user interface and access to individual portions of the code, and 4) to add new modules which incorporate updated physics and improved features. These tasks have been identified as those which would be the most beneficial to current users of the code. For those that use the code as a "blackbox" by supplying the appropriate input and examining the output, FASE will appear identical to FASCODE/LBLRTM. However, for users who wish to modify the code or incorporate it within other software, FASE will be a much easier code with which to work.

The key features of FASE are shown in Table 1. Items of scientific interest include a more accurate algorithm for the Voigt profile, updated non-local thermodynamic equilibrium

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Table 1. Key features of FASE.

Science Improvements	Coding Improvements
Improved Voigt Algorithm	Parameterization of Array Sizes
Updated NLTE Routines	Output in NCAR Graphics Format
Line Rejection Flag	Coded for Vectorized Computers
Schumann-Runge Bands and Continuum Programs	Output Formatted for Multiple-Scattering
Solar Spectrum Model	Embedded FFT Scanning Functions
Improved Hartley-Huggins and Chappuis Bands	Optional Fixed DV Output 2020 cm^{-1} Spectral Region (max) Improvements to Geometry Algorithm

(NLTE) routines, a line rejection flag to signal whether or not a particular spectral line was rejected for use in a layer,^(a) the inclusion of Schumann-Runge band and continuum features (Minschwaner et al. 1992), and the addition of a solar spectrum model (Kurucz 1994). Changes to the program source code include vectorization of routines to increase the computational speed, parameterization of dimension statements to allow for easy changes in the size of arrays (e.g., the maximum number of layers allowed), the ability to configure the model for a variety of computers (which is aided by the parameter statements), and improvements to the atmospheric path geometry formulation. Other features of FASE include the option for output that is formatted for input to multiple-scattering programs, such as CHARTS (Moncet and Clough 1992) or DISORT (Stamnes et al. 1988), and the ability to calculate spectra over a much wider region than currently allowed by FASCODE.

Because we are combining elements of FASCODE and LBLRTM and wish to make improvements to the overall structure and use of the code, there are two key issues which must be addressed: 1) what are the origins and magnitudes of numerical differences between FASCODE and LBLRTM, and 2) what coding improvements should be adopted which would maximize algorithm speed while allowing flexibility for the user community and maintainability for the code itself. This paper addresses the second of these issues.

(a) The line rejection flag is particularly useful when computing radiance derivatives with respect to temperature using a finite-difference scheme since a line could be rejected in the reference case and not after perturbing the temperature, or vice versa. In such cases large errors would be introduced unless the line was consistently rejected.

Program Structure

An essential point to be considered in the development of FASE is the determination of the structure of the program. In the current FASE structure (inherited from FASCODE) the optical depth and radiative transfer calculations are performed at a spectral resolution appropriate for the local pressure conditions. This results in a complex algorithm for merging layers with different spectral resolution. Further, the various core modules for the calculation of molecular optical depth, aerosol attenuation, atmospheric path characteristics, filtering, etc., are grouped into a single executable code with a complex driver controlling the sequence of calls to the different subroutines to perform any one of numerous predefined functions (e.g., radiance in a clear atmosphere, radiance in a cloudy atmosphere, or weighting functions). This structure was adopted in response to a desire to reduce the number of computations and also to accommodate the needs of users not familiar with radiative transfer. Aside from the fact that an extensive set of instructional inputs must be provided by the user in order to run the code, the current structure has a serious drawback for research applications in the sense that the code cannot be easily tailored to meet specific user needs (Gordley et al. 1994) or to optimize the execution time for specific applications.

One option under study to solve this issue is to split the various core modules into independent subroutines or executable modules with clearly documented inputs and outputs. A set of simple, easily customized, standard radiative transfer routines that perform single functions would be provided, thus eliminating the need for an extensive input control file and giving the user the capability of organizing a sequence of functions to meet the requirements of a particular problem. The modules would be structured such that a novice user need only create an input file and examine the output (as with the current FASCODE), with all intermediate steps transparent to the user. However, what would be different from the current program format is that the structure of the modules would make it easy for the experienced user to rapidly implement new features or modify parameters in existing features.

As mentioned earlier, the complexity of merging layer optical depths in FASCODE/LBLRTM results from the desire to calculate layer transmittances at the local spectral resolution in order to save on the required number of exponentiations when computing the radiative transfer. However, for a non-homogeneous atmospheric path this approach requires that calculated transmittances be interpolated to the spectral resolution of the next layer.

Computational savings can only be derived with this approach by merging the successive layers from bottom to top (in the direction of decreasing pressure). We believe that this approach is obsolete since performing the exponentiations is only a small fraction of the total computer time on modern computers. Furthermore, for many applications such as multiple-scattering, inversions, etc., much greater gains can be obtained by merging the layers from top to bottom and bottom to top at the same time (Moncet and Clough 1992). Thus it is best not to attempt to satisfy the above merging constraint. Finally the spectral overlap needed for performing the interpolation defeats the requirements for parallel processing.

One way to solve this problem is to build the code around a core module which computes the molecular optical depth at the same spectral resolution for each layer. The actual computation occurs at the spectral resolution appropriate for the layer, but the result is then interpolated to the final spectral resolution (which corresponds to the highest resolution in the problem, usually prescribed by the geometric line-of-sight). The advantage of this technique is that the total spectral interval can be broken into sub-intervals with uniform end-points for all layers, making customization of the code easier. The size of these sub-intervals would depend on the amount of core memory available. Because of the uniform output for all layers, a complete calculation can be performed in memory for each sub-interval. Furthermore, this approach simplifies the extension to parallel processing of the sub-intervals.

The drawback of interpolating the molecular optical depths to the same resolution is that the radiative transfer must be computed at high resolution even for layers where a lower resolution would suffice for computational accuracy. The impact of this on the timing has been found to be minimal (Snell et al. 1995). In addition, the structure of the program is simplified, and it would be very easy for the user to adapt the modules to meet the needs of specific problems. For calculations consisting of multiple runs (different atmospheres, instrument functions, viewing geometry, etc.), this structure would be more efficient and easier to use because it would eliminate the need for multiple input files and would allow the processing of different operations in parallel (saving read/write time).

Conclusions

Creating the FASE from the existing routines of FASCODE and LBLRTM and restructuring the source code will increase the flexibility of the code and allow for easy customization of the algorithm to fit specific user problems.

Acknowledgments

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Cirrus and Aerosol Lidar Profiler - Analysis and Results

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Introduction

A cloud and aerosol lidar data set from over a year of near continuous operation of a micro pulse lidar (MPL) instrument at the Cloud and Radiation Testbed (CART) site has been established. MPL instruments are to be included in the Ames Research Center (ARC) instrument compliments for the SW Pacific and Arctic ARM sites. Operational processing algorithms are in development for the data sets. The derived products are to be cloud presence and classification, base height, cirrus thickness, cirrus optical thickness, cirrus extinction profile, aerosol optical thickness and profile, and planetary boundary layer (PBL) height. A cloud presence and base height algorithm is in use, and a data set from the CART site is available. The scientific basis for the algorithm development of the higher level data products and plans for implementation are discussed.

MPL Instrument

The unique features of micro pulse lidar are the ability to profile all significant cloud and aerosol structure of the atmosphere with a compact, fully eye safe instrument that is capable of full-time, long-term unattended operation. The initial instrument was developed at Goddard Space Flight Center as a spinoff of research on efficient lidar for space borne applications. An MPL instrument was first used in the SW Pacific PROBE study in early 1993 and in October 1993 an instrument began full-time measurements at the Atmospheric Radiation Measurement (ARM) Southern Great Plains (SGP) field site.

MPL is a conventional time-gated, incoherent detection lidar approach. Beyond that there are three basic differences between the micro pulse lidar and most previous lidar systems. First the laser pulse repetition frequency (PRF) is

much higher, kilohertz rather than Hertz, and the pulse energies are much lower, micro-Joules rather than milli-Joules or greater. The low-pulse energy is the factor that permits the systems to be eye safe. The second difference is that the laser is diode-pumped rather than flashlamp-pumped. The solid state lasers are much more efficient and smaller. The third difference is that signal detection is by photon counting. Photon counting is generally a more accurate and problem-free means of signal acquisition for low level signals than analog detection. The details and advantages of the MPL design are discussed by Spinhirne (1993), and Spinhirne et al. (1995) give the specifications of the MPL system in use at the SGP site. Upgraded instruments are to be installed at the SW Pacific and Arctic ARC sites. A high-performance instrument with much higher pulse energies and polarization measurements is in development.

MPL Data

The lidar data are stored as 1-minute average profiles. The signal return in photo electron counts per second $n(r)$ at range r may be given by

$$n(r) = CE(r)J[\beta_m(r) + \beta_p(r) + \beta_c(r)]T^2(r)/r^2 + n_b \quad (1)$$

where β is the backscatter cross section due to molecular, particulate, and cloud scattering as indicated by the subscript and T is atmospheric transmittance. C is a dimensional system calibration constant, J is the laser pulse energy, and E is a near range overlap factor that becomes unity at several kilometers range. n_b is a background signal from daytime solar photons entering the receiver. Signals are acquired for 400 μ s or 60-km range. The last 100 μ s of the signal is negligibly different from n_b and thus gives a

value for the background. A basic signal display form is the range corrected signal, or relative attenuated backscatter cross section, and is given by

$$p(r) = [n(r) - n_b]r^2 \quad (2)$$

Image displays of $p(r)$ in a time versus range format are valuable for initial data assessment and may be used for initial image algorithm processing of data. An example of one week of data from the SGP site in this format is given in Figure 1. Such images in graphical interface format (GIF) format are to be a standard data product from the MPL instruments.

Lidar Data Processing Strategy

A two-step processing strategy data processing algorithm is being implemented. In the first step, the goal is to classify the observational case and determine the location of the signal discontinuities, such as cloud base height, that are present. The analysis needs to be done before data has been calibrated and processed. The method that is used is threshold detection analysis but image-based analysis of the data displayed in time-height cross section form is in development. The second step is lidar equation signal analysis of the lidar data. A classification of the data by factors such as cloud presence and cloud type is needed before this step may be completed. The goal of the lidar equation signal analysis is to derive quantitative measurement parameters of cloud and aerosol optical thickness and cross section profiles.

Image Analysis and Classification

The most basic classification approach is threshold crossing analysis to detect the presence of cloud boundaries. A limitation of the approach is that for low signal-to-noise conditions, the threshold must be kept above noise fluctuations, which degrades sensitivity. In addition, the noise is a rapidly changing variable of background and range. However, when only a single profile is tested, improvements due to horizontal correlations are not realized. The sensitivity of thresholding may be maximized by adjusting the threshold as a function of the noise given by square root of $n(r)$. The current cloud base data set is determined by this technique. Cloud height statistics for a ten-month period are shown in Figure 2.

Image analysis takes advantage of spatial correlations. By using image processing approaches to determine the spatial boundaries of clouds and haze layers, advantage can be taken of horizontal correlations of the data as well as the sharp vertical gradients that indicate a cloud return. Edge detection algorithms extract this information and may be used to find cloud base and top. Defining cloud/background regions may be automated using computer vision techniques such as image segmentation, which partitions an image into a set of non-over-lapping regions whose union is the entire image. Local statistics such as mean and variance help determine the partitions. Other techniques to be explored in the future include line extraction and noise cleaning methods.

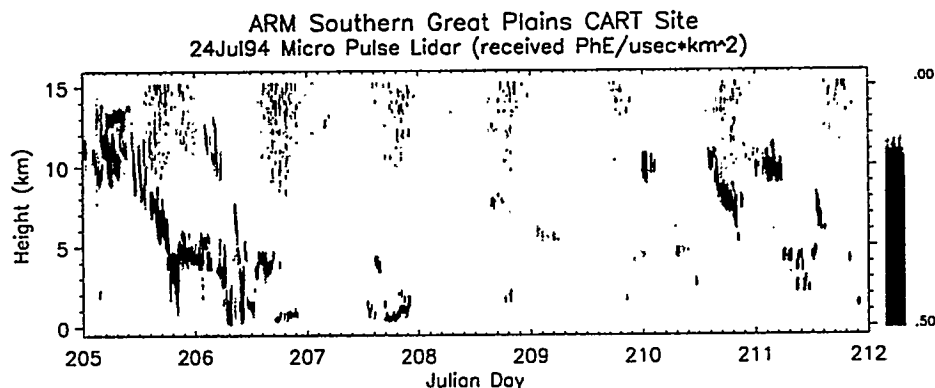


Figure 1. Image of MPL data from the Oklahoma CART site. The data is displayed in range corrected format but with no near range overlap correction of the near ground signal. Similar browse images for all MPL data are to be made available in GIF format.

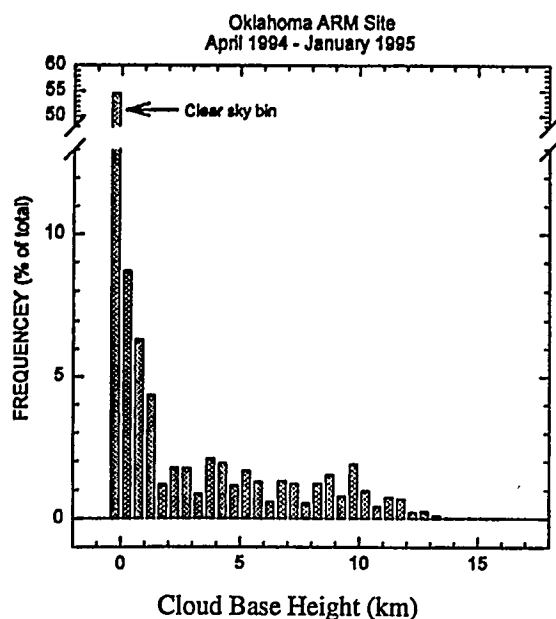


Figure 2. Cloud base height statistics from the MPL system for ten months of operation at the Oklahoma CART site.

Lidar Signal Equation Analysis of Data

The research objective of quantitative analysis of lidar data is to derive independent height resolved radiation variable data sets as appropriate for classified atmospheric conditions. A main goal is to obtain the cirrus optical thickness and extinction profile. Due to small field view of the MPL system ($100\mu\text{rad}$), a multiple scattering correction is not required. For optical thickness less than approximately one, the attenuation of the molecular and aerosol return signal from beyond the cloud top directly gives the cloud optical thickness of thin clouds. Signal averaging is generally required. When the optical thickness is approximately greater than one and less than two, an alternate approach is to correlate the integrated cloud return signal with the thermal or visible background radiance. The correlation provides a boundary condition to relate the measured backscatter cross section to the extinction cross section which allows the optical depth and extinction to be determined. Cloud optical depth of greater than approximately two cannot be measured directly by lidar.

The aerosol optical thickness and extinction profile may be similarly analyzed. The basic measurement parameter of lidar is the product of backscatter cross section times atmospheric attenuation. The backscatter may be calibrated by assuming molecular scattering for the cleanest conditions of

the upper troposphere and stratosphere. With the assumption of constant extinction to backscatter ratio, the lidar equation may be solved for extinction and optical thickness.

A generic description of the procedure is given here. The lidar equation may be given in term of the range and calibration normalized signal as

$$P(r) = (n(r) \cdot n_b) r^2 / CJE(r) \cdot \beta(r) T^2(r) \quad (3)$$

The well-known equation for the solution is

$$T^2(r_1, r_2) = 1 - \left(\frac{2}{k'}\right) I_{1,2} \quad (4)$$

where

$$I = \int_{r_1}^{r_2} \beta(r') T^2(r') dr' \quad \text{and} \quad k' = \beta(r) / \sigma(r) \quad (5)$$

if σ is the extinction cross section. The integral I is obtained directly from the lidar signal if the relative calibration constant C is known. C is obtained by normalization in the upper troposphere to the known molecular scattering cross section once the transmission is found. If $I_{1,2}$ is the integral of the signal for the clear air region in the upper troposphere and I_0 is the integral with no clouds or aerosol present below r_1 then

$$T^2(r_1) = I_{1,2} / I_0 \quad (6)$$

it follows that

$$I_{0,1} = \left(\frac{k'}{2}\right) \left(\frac{k'}{2I_0}\right) I_{1,2} \quad (7)$$

If a range of optical thickness is encountered in time, as is typical for PBL aerosol or cirrus layers, then (7) represents a system of linear equations by which k' may be determined from least squared analysis. The overall solution of the lidar data for the system calibration constant, the extinction to backscatter ratio, and the desired cirrus or aerosol optical depth when there is a long time history of measurements would be an iterative procedure. A more complete description of the theory involves the separation of terms for the molecular and cirrus or aerosol scattering components

(Spinhirne et al. 1990, 1994). The utility of the analysis procedure for the MPL data is to be tested.

Summary

Cloud boundary classification of MPL data is currently being processed by a threshold analysis technique. A data set for over a year's operation at the ARM CART site has been created. An improved classification procedure based on image analysis techniques will provide cloud top and thickness results up to the limit of signal attenuation and also the presence and extent of aerosol layers including PBL height boundaries. A quantitative data analysis procedure based on lidar equation analysis is in development to provide vertically resolved measurements of cirrus and aerosol extinction cross sections. The cross section measurements will be applied with ground visible and infrared radiance measurements to study basic cloud and aerosol radiation effects. An upgraded MPL instrument with greater sensitivity, dual wavelengths, and polarization is in development and should provide for greater accuracy of measurement results.

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Aspects of the Quality of Data from the Southern Great Plains (SGP) Cloud and Radiation Testbed (CART) Site Broadband Radiation Sensors

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A systematic evaluation of the performance of broadband radiometers at the Southern Great Plains (SGP) Cloud and Radiation Testbed (CART) site is needed to estimate the uncertainties of the irradiance observations. Here, net radiation observed with the net radiometer in the energy balance Bowen ratio (EBBR) station at the central facility is compared with the net radiation computed as the sum of component irradiances recorded by nearby pyranometers and pyrgeometers. In addition, data obtained from the central facility pyranometers, pyrgeometers, and pyrhemometers are examined for April 1994, when intensive operations periods were being carried out.

The data used in this study are from central facility radiometers in a solar and infrared observation station (SIROS), an EBBR station, the so-called "BSRN" set of upward pointing radiometers (where BSRN previously implied Baseline Surface Radiation Network), and a set of radiometers pointed down at the 25-m level of a 60-m tower. The 25-m radiometers view a wheat field. The SIROS, BSRN, and EBBR radiometers are all located in a pasture and are positioned within 25 m of each other; the upward pointing radiometers are at a height of about 1.5 m above the local surface, the net radiometer is at a height of 2.3 m, and the downward pointing SIROS pyranometer and pyrgeometer are at a height near 9.5 m. All of the upward pointing pyranometers and pyrgeometers are ventilated, but the net radiometer and the downward pointing radiometers are unventilated.

The uncertainties of readings from net radiometers are known to be substantial at times and differ considerably for varying radiometer designs (e.g., Field et al. 1992; Halldin and Lindroth 1992). The absolute uncertainties of net radiometer readings might optimistically be viewed as being no better than about $\pm 10 \text{ W m}^{-2}$ or $\pm 5\%$, whichever is larger, for any particular half-hour irradiance average. In

this paper, suggested uncertainty values correspond to the range for confidence levels of 90-95% for averaged readings. For pyranometers subjected to the broadband outdoor radiometer calibration (BORCAL) procedure by the National Renewable Energy Laboratory (NREL), the uncertainty of individual readings is reported to be typically about 3% (Wells et al. 1993). Pyranometer data reported here and by the Atmospheric Radiation Measurement (ARM) experiment center are all computed with calibration coefficients generated from BORCALs. Recent comparisons, however, of BORCALs sensitivity coefficients to those produced originally with integrating spheres by the manufacturer, Eppley Laboratory, Inc., suggest that the NREL BORCAL procedures result in irradiance values that are approximately 3% larger on average than those that would be obtained with the Eppley coefficients.^(a) For pyrgeometers, past experiences have shown that irradiance estimates using factory sensitivity coefficients have typical uncertainties of about $\pm 20 \text{ W m}^{-2}$. By careful comparison of pyrgeometers and subsequent adjustment of calibration coefficients, however, agreement within 5 W m^{-2} for any group of pyrgeometers appears to be possible;^(b) NOAA/SRRB has carried out such comparisons for the pyrgeometers used in this study. All pyrgeometer data reported by the ARM experiment center are currently derived with the Eppley factory calibrations.

Net Radiation

Differences between 30-min averages of EBBR net radiometer (R_n) values and values computed as the

(a) J. DeLuisi, National Oceanic and Atmospheric Administration (NOAA), Surface Radiation Research Branch (SRRB), Private communication.

component sum from the SIROS data are shown in Figure 1 for the months of December 1994, January 1995, and February 1995. With the pyrgeometer factory (Eppley) calibrations, the nighttime differences increase from near zero to about 15 W m^{-2} as the net radiation decreases from zero to about 80 W m^{-2} . The differences are displaced by about 15 W m^{-2} for all net radiation values when the NOAA/SRRB calibration coefficients are used for the pyrgeometers. We hypothesize that the nighttime trend evident for both sets of data indicates that the longwave sensitivity of the EBBR net radiometer is slightly different than the sensitivity of the pyrgeometers. For the positive values of net radiation in Figure 1, most of the differences would probably be made larger on average by approximately 3% of the values on the horizontal scale if the Eppley sensitivity coefficients had been used rather than the BORCAL coefficients for the pyranometers.

Condensation and frost are known to occur on the radiometer domes, particularly on the upper dome of the EBBR net radiometer when exposed to cloudless nighttime skies. Condensation on the domes of the upward pointing pyrgeometers and pyranometers is less likely because they are ventilated. When frost or condensation at night was visually seen only on the EBBR radiometer domes, the

EBBR net radiation was sometimes as much as 50 W m^{-2} larger (less negative) than the SIROS pyrgeometer estimates made with the Eppley calibration factors. Such large discrepancies disappeared after the condensation or frost evaporated from the domes. We suggest that condensation or frost on the EBBR net radiometer dome is considerably warmer than the sky and thus causes overestimates of downward longwave radiation. Further analyses have indicated that largest overestimates of nighttime values from the EBBR net radiometer during December 1994 through February 1995 tended to occur when the relative humidity was high, when condensation was particularly likely. Some of these cases can be seen in the upper left-hand corner of Figure 1.

Recent analyses indicate that readings from the EBBR net radiometer during the afternoon at the central facility during December 1994 through February 1995 were sometimes suppressed by as much as 100 W m^{-2} . By visual inspection, condensation was seen on the interior of the EBBR domes; this condensation could have shaded the sensing element slightly from direct-beam solar radiation. Cloudless skies and southerly winds seemed to be correlated with this situation. The reduced values for the data points seen in the lower right-hand corner of Figure 1 might have

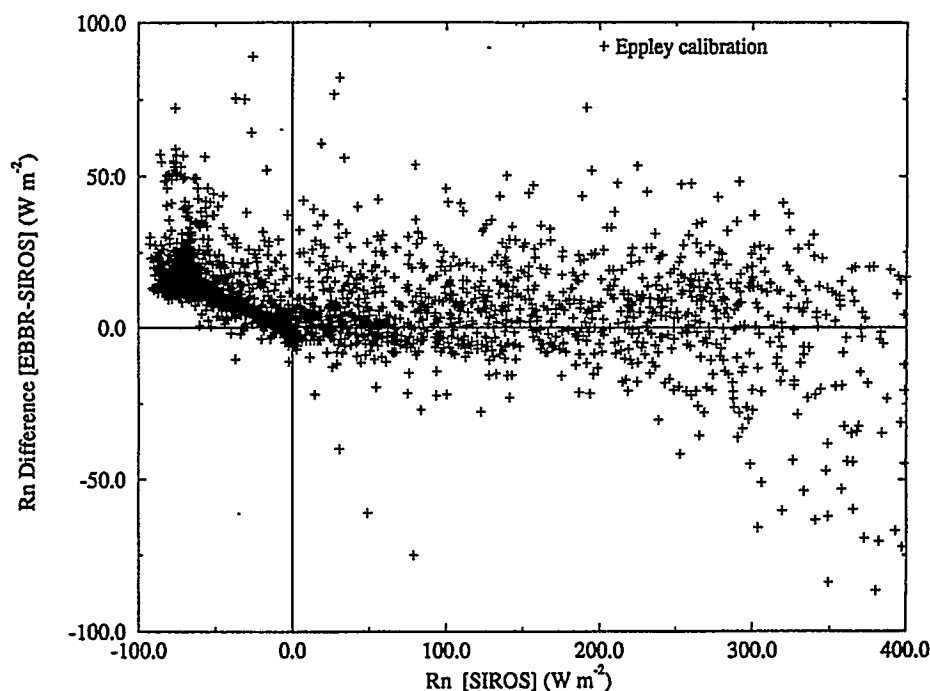


Figure 1. Differences between 30-min-average net radiation estimates from the central facility EBBR and SIROS during December 1994, January 1995, and February 1995 versus the SIROS net radiation values. The SIROS net radiation estimates were computed with the Eppley factory calibration coefficients.

been caused by this condensation problem. We suspect that the source of the condensation might be eliminated by an improved seal around the net radiometer domes and more frequent change of desiccant in a tube leading to the dome air, but further investigation is needed.

Replacement of the domes on the EBBR net radiometers could cause a change of as much as 3% in its sensitivity, according to the manufacturer. A rather large change was seen when the domes were replaced on December 2, 1994, after which the nighttime R_n difference dropped on average from about 30 W m^{-2} to 15 W m^{-2} when the net irradiance was near 80 W m^{-2} . Other instances of dome changes will be studied to characterize more thoroughly the effects of dome replacement on the EBBR net radiometer readings.

The effects of using SIROS net radiation estimates for the calculation of sensible and latent heat fluxes can be evaluated by replacing the EBBR net radiation values with SIROS-derived values in the energy balance equation. Figure 2 shows a case study, in which the Eppley coefficients were used with the pyrgeometer data. Our analyses suggest that the latent heat fluxes, on average,

appear to be mostly unaffected by use of the SIROS net radiation values, but the sensible heat fluxes are substantially altered at times, somewhat in proportion to the changes in the net radiation values used.

April 1994 Data

A set of graphs displaying daily broadband radiation data obtained at the central facility in April 1994 can be obtained from M. Wesely at mlwesely@anl.gov or J. Tichler of Brookhaven National Laboratory at tichler@met.das.bnl.gov. Figure 3 shows the types of data addressed. Observations of diffuse and direct-beam solar irradiances with the central facility SIROS were not made during this month. Because the BSRN global solar irradiance ($K_{\downarrow}$) computed as the sum of direct and diffuse components agreed well on April 18 with the BSRN and SIROS $K_{\downarrow}$ observed directly with unshaded pyranometers, the BSRN pyranometer used to observe the diffuse irradiance was apparently shaded quite well on this day. Such comparisons on other days indicate that the shading disk was improperly positioned for periods of several hours on April 6, 23, and 24.

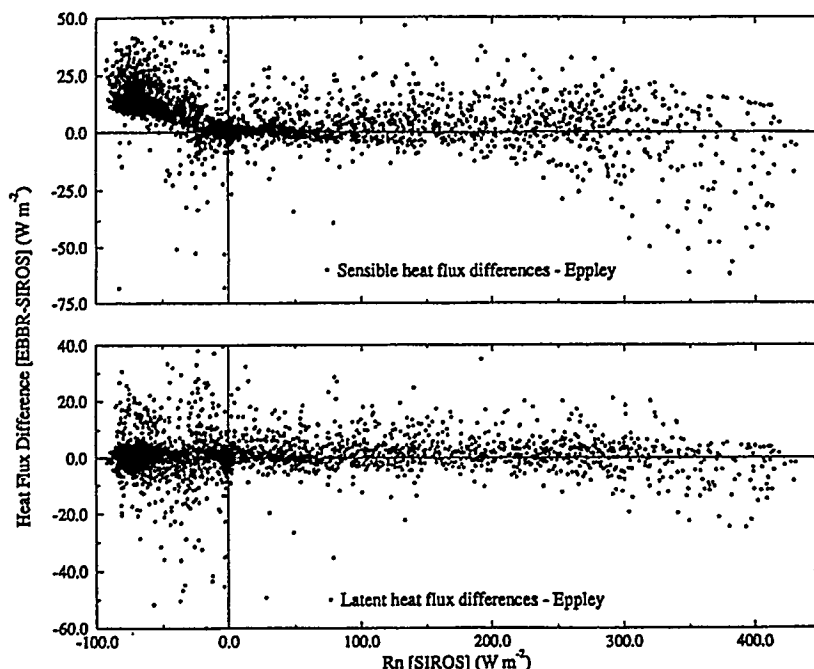


Figure 2. Differences between 30-min-average surface flux estimates from the EBBR and recalculated fluxes using the SIROS net radiation values (with the Eppley factory calibration coefficients) for (a) sensible heat flux and (b) latent heat flux for December 1994, January 1995, and February 1995.

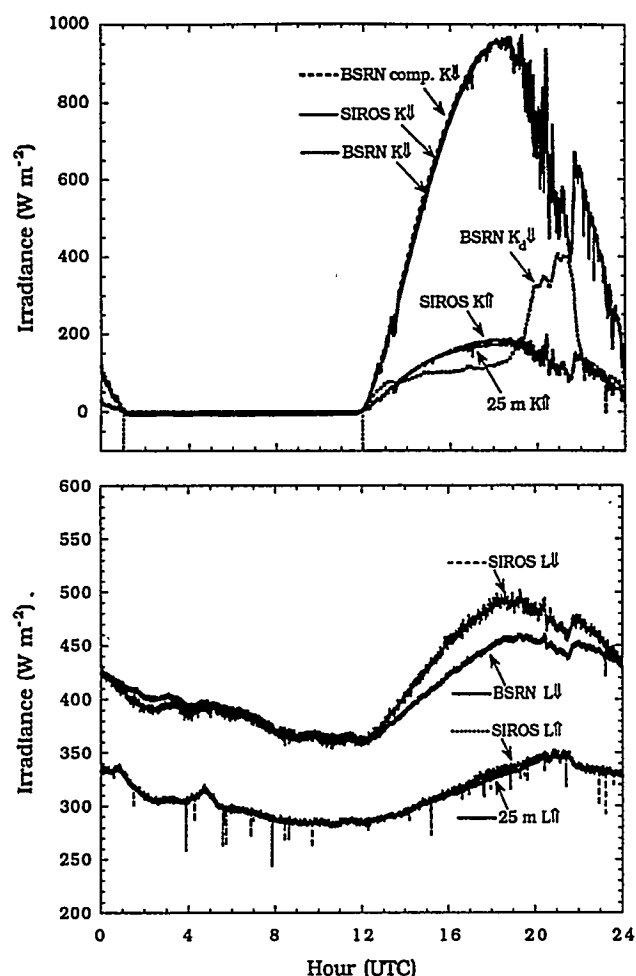


Figure 3. Irradiance observations made with pyranometers, pyrgeometers, and a pyrheliometer during April 18, 1994, where "K" indicates solar radiation, "L" longwave radiation, the downward pointing arrow to downwelling radiation, the upward pointing arrow to upwelling radiation, "comp." to the component sum of diffuse plus direct-beam irradiance on a horizontal surface, and the subscript "d" to diffuse solar radiation seen with a shaded pyranometer.

"Spikes" can be seen in Figure 3 on several of the SIROS data streams. All SIROS data streams from the central facility except for the upwelling longwave irradiance exhibited this problem. It was eliminated on September 20, 1994, by moving an electric cattle fence surrounding the SIROS instruments to a greater distance from the instrumentation. These spikes also occurred at the two Ashton and Ringwood extended facility SIROSs in operation in April 1994 and continued until the electric

fences there were relocated in September 1994. From the dates of the SIROS installations to the dates when the electric fences were moved, spikes on the SIROS data were common whenever the fence was activated. In general, the spikes tended to bias the SIROS readings low by a 0.2–1.5 W m^{-2} for 30-min averages for all affected sensors except the SIROS pyrheliometer, for which the reduction could be as large as 10 W m^{-2} . For 5-min averages, the reductions could be 3–4 times larger.

Calculations show that the observations of diffuse solar irradiance with the BSRN shaded pyranometer are 4–8% larger than comparable, midday values from the BSRN and SIROS unshaded pyranometers on overcast days in April 1994. The cause of this discrepancy is unknown, but we recommend that users of the data multiply the diffuse readings by 0.94. The upward pointing SIROS pyrgeometer read about 2% larger on average than the BSRN pyrgeometer, and the two upward pointing hemispherical solar sensors appear to give mean values within 1% of each other. Table 1 lists surface albedos computed for days selected for consistently overcast skies and for cloudless skies between 1600 and 2000 hr (UTC). The albedos for overcast conditions do not appear to have changed significantly from day to day and were about the same for the two fields. For cloudless conditions, the albedos might have decreased slightly during the month, with the wheat field having a perceptibly smaller albedo than the pasture.

Table 1. Computed values of surface albedo at two locations at the central facility during April 1994.

Day of Month	Pasture	Wheat
Overcast		
5	0.186	0.201
9	0.205	0.202
11	0.147	0.185
22	0.196	0.199
Cloudless		
3	0.222	0.214
6	0.203	0.182
15	0.191	0.175
16	0.194	0.174
17	0.194	0.184
26	0.187	0.165

Acknowledgments

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The Multi-Filter Rotating Shadowband Radiometer (MFRSR) - Precision Infrared Radiometer (PIR) Platform in Fairbanks: Scientific Objectives

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Arctic Qualification of the Multi-Filter Rotating Shadowband Radiometer (MFRSR) and Precision Infrared Radiometer (PIR) Instruments

The MFRSR and the PIR have been deployed at the Geophysical Institute in Fairbanks to check their performance under arctic conditions. Drawing on the experience of the previous measurements in the Arctic, the PIR was equipped with a ventilator to prevent the build-up of frost and moisture on the its dome.

We adopted the Solar Infrared (IR) Observing System (SIROS) concept from the Southern Great Plains Atmospheric Radiation Measurement Site to allow implementation of the same data processing software for a set of radiation and meteorological instruments.

The ongoing work is focused on practical problems and is in the stage of testing the sensors and data acquisition system.

To validate the level of performance of the whole SIROS prior to its incorporation into the North Slope of Alaska (NSA) Cloud and Radiation Testbed Site instrumental suite for flux radiation measurements, the comparison between measurements and model predictions will be undertaken to assess the MFRSR - PIR Arctic data quality.

Potential for Determining Cloud Properties from MFRSR Data

A preliminary feasibility study of the MFRSR capabilities in conjunction with comprehensive radiative transfer computations to infer cloud optical properties was performed

(Leontieva and Stamnes 1994). We assume that the surface albedo is known from the measurements of a MFRSR unit that operates without a shadowband looking down.

We propose to consider transmittances rather than irradiances because this allows us to deal only with the uncalibrated outputs of the instrument and makes the precision to which the filter functions are needed to be known less critical. The use of transmittances is highly beneficial from the computational point of view since an efficient algorithm is available to compute the transmittance for the earth-atmosphere system including vertical atmospheric inhomogeneity (Stamnes 1982). Figure 1 shows the look-up tables of computed transmittances for a specific day at Barrow, Alaska, having a step of 0.005 over $\cos(\text{SZA})$. Figure 1 demonstrates the possible effect of aerosol extinction on transmittance implying the usefulness of commonly observed atmospheric optical depths at a particular location that will be continuously retrieved from the MFRSR measurements of direct solar irradiance under true cloud-free conditions. The use of observational values of aerosol optical depths becomes especially desirable at high surface albedo.

Model simulations show that both cloud optical depth and cloud droplet effective radius could be inferred from the measurements of bispectral transmittance, using a combination of any wavelength without liquid water absorption (for example, 862 nm) and a 2250 nm wavelength. Two sets of contour curves presented in Figure 2 illustrate the possibility of simultaneously retrieving cloud optical depth and effective radius at least in the range of τ from 3 to 30. Better results are also anticipated at smaller, $r_e < 6 \mu\text{m}$, drop sizes.

A method for simultaneously retrieving both cloud optical depth and surface albedo from measurements of downwelling spectral irradiances in the visible part of the spectrum has been introduced by Ricchiuzzi et al. (1995). We

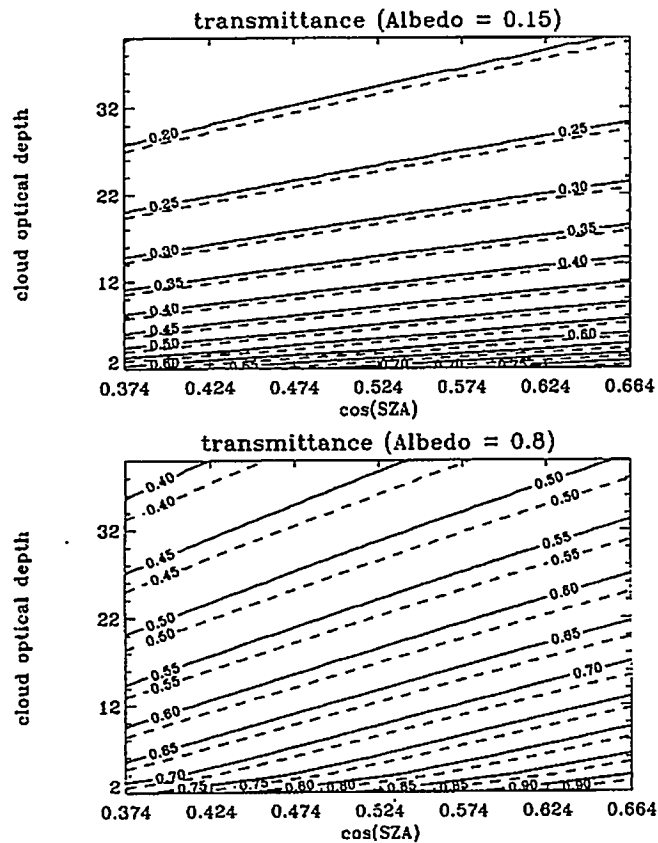


Figure 1. Contours of the computed transmittance at 862-nm MFRSR passband, for subarctic summer atmosphere and two aerosol models: MODTRAN h6, atmosphere optical depth $\tau = 0.105$ (solid lines) and the scaled model $\tau = 0.2$ (dashed lines). The range of $\cos(\text{SZA})$ corresponds to DAY=180, Barrow, Alaska ($\text{SZA} < 70^\circ$).

propose to test and apply this method to the measurements of transmittances in any pair of MFRSR channels (excluding 940 nm). Figure 3 provides a graphical illustration of the retrieval method and displays the contour plot of computed transmittances for two MFRSR channels, namely at 415-nm and 862-nm passbands.

Thermal IR Irradiance at the Surface: Model Predictions and Observations at Barrow, Alaska

The downward longwave irradiances measured by the PIR at the National Oceanic and Atmospheric Administration Climate Monitoring and Diagnostics Laboratory (CMDL) station at Barrow, Alaska, during 1988 and 1992 have been

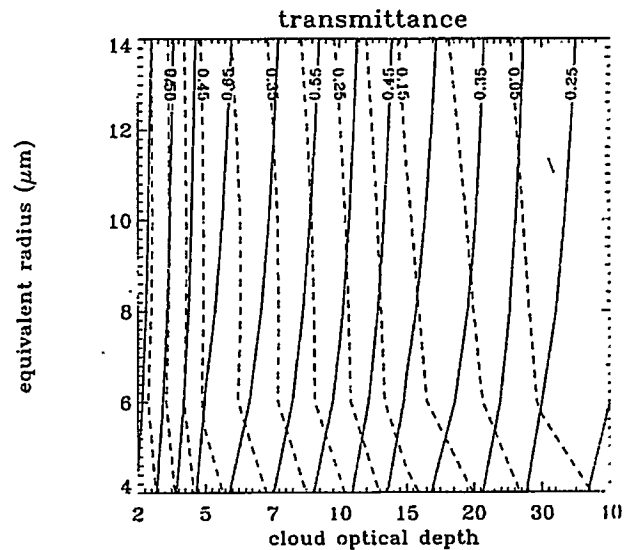


Figure 2. Contours of the computed transmittance at 862-nm MFRSR passband (solid lines), and at 2250 nm (dashed lines) for subarctic summer atmosphere and MODTRAN haze=6 tropospheric aerosol model at surface albedo $A = 0.15$. $\text{SZA} = 45^\circ$.

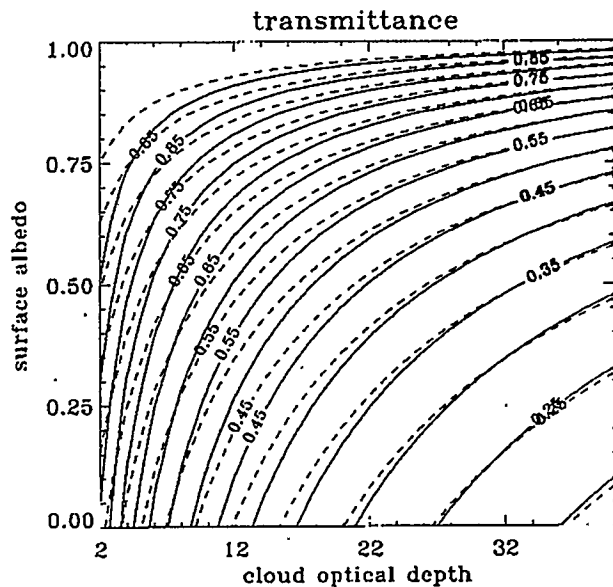


Figure 3. Contours of the computed transmittance at 862 nm (solid lines) and at 415 nm (dashed lines) of the MFRSR for subarctic summer atmosphere and MODTRAN haze=6 tropospheric aerosol model. $\text{SZA} = 45^\circ$.

compared with the model estimations based on radiosonde sounding data. The analysis was restricted to the times reported to be cloud free. Scatterplots of measured hourly-averaged IR fluxes at the surface versus calculated values are shown in Figure 4. The measurements made by the ventilated version of PIR in 1992 produce better

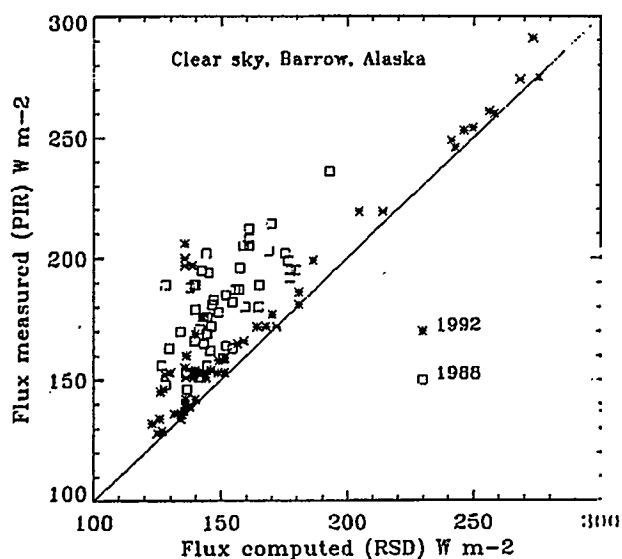


Figure 4. A scatterplot of measured versus computed downward IR irradiance at the surface.

agreement with the modeled values as compared with corresponding magnitudes in 1988 (PIR without ventilator). Table 1 presents the linear regression results for both data sets.

Although a better agreement between measurements and computations in 1992 might also be caused by different atmospheric conditions, we believe that these results demonstrate an improved quality of the IR flux measurements by the modified PIR in the Arctic. Generally, measured "clear

sky" downward IR flux in winter appeared to be greater than expected from the model predictions that by implication were associated with lower tropospheric ice crystals.

The impact of ice crystals as compared to water droplets (fog) on the downward IR flux at the surface is shown in Figure 5 based on model simulations. The radiative properties of clouds were modeled in terms of particle effective size and ice (liquid) water content using parameterizations of Fu and Liou (1993) and Hu and Stamnes (1993) for ice and water cloud, respectively. An increase in effective particle size (both drops and ice crystals) causes a decrease in downwelling IR flux at the surface and an increase in outgoing IR flux to space since the clouds become less opaque. For the same effective size and water path, the ice cloud is more "black" as compared with water cloud.

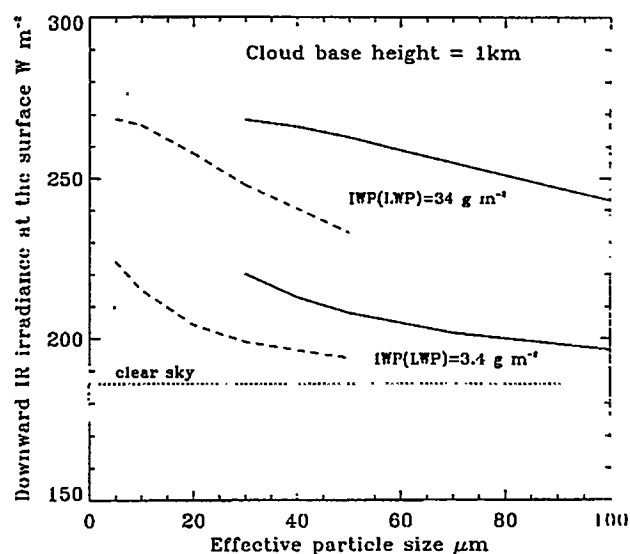


Figure 5. Downward IR irradiance at the surface versus effective cloud particle size (ice crystals shown as solid line; water droplets shown as dashed line).

Table 1. Comparison between measured and calculated downward longwave fluxes under clear skies.

Year	Number	Rms difference W m^{-2}	Bias (Meas-calc)	Slope	Intercept	Corr. Coeff.
1988	52	33	30	0.9	45	0.70
1992	55	20	13	0.9	29	0.94

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Observations of a Cold Front With Strong Vertical Undulations During the ARM RCS-IOP

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Passage of a cold front was observed on the night of April 14-15, 1994, during the Atmospheric Radiation Measurement (ARM) Remote Cloud Sensing (RCS) Intensive Observations Period (IOP) at the Southern Great Plains Cloud and Radiation Testbed (CART) site near Lamont, Oklahoma. This front produced severe thunderstorms on a line from Wisconsin to southeastern Oklahoma (OK). The series of water vapor profiles obtained by the Goddard Space Flight Center (GSFC) Scanning Raman Lidar (SRL) captures well the dramatic sequence of events at the CART

site (Figure 1). The SRL was operated in alternating vertical and near-horizontal pointing modes, each with 1-minute dwell, where the former yielded high vertical resolution (75 m) profiles from 165 m to above 8 km and the latter provided ultra-high vertical resolution profiling from near the surface. Details of the SRL and modes of operation may be found in Melfi et al. (this volume).

The initial observations show a fairly sudden deepening of the boundary layer with progressive moistening after about

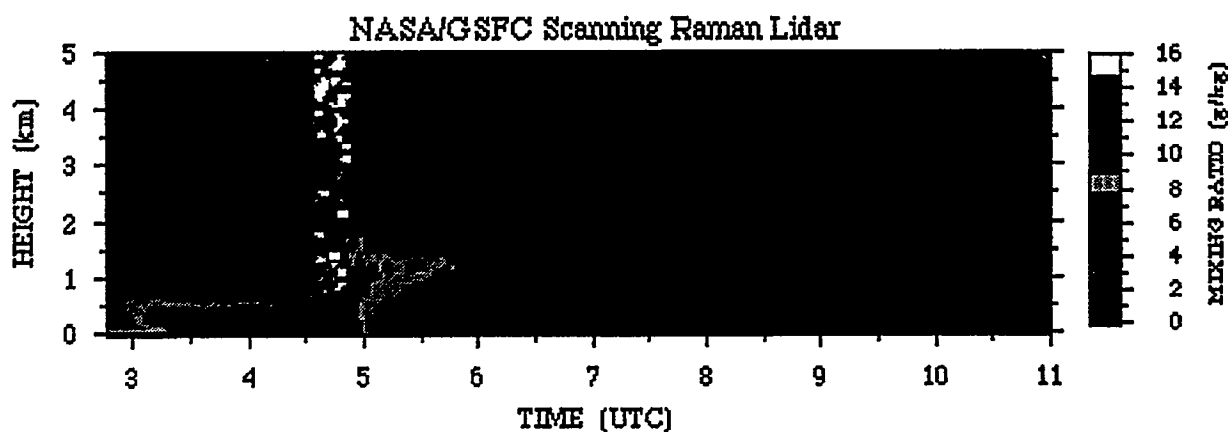


Figure 1. Observations of water vapor mixing ratio profiles by the GSFC Scanning Raman Lidar on 15 April during the 1994 ARM RCS IOP.

0300 UTC. At about 0430 UTC, very rapid growth in the planetary boundary layer (PBL) was observed with strong apparent wave activity. The vertical banding in the data indicate the presence of an attenuating cloud capping the first and second wave peaks (also indications of haze or very tenuous cloud on the third wave peak). Our real time display (and a strong wind gust) induced us to quickly make a visual observation and two roll clouds were sighted. Very high resolution observations by the University of Utah Polarization-Diversity Cloud Lidar revealed a sculpted cloud base (sloping upward) while the Penn State 95-GHz Doppler Radar showed turbulent dust plumes extending upward from the surface and alternating regions of strongly upward and downward motion in association with these waves.

This was followed by a change to a drier boundary layer capped by an upward sloping frontal surface. It should be noted that undulations, though more moderate, continued to be observed in the elevated frontal zone (note that the wave at 0340 UTC is purely an artifact of a brief 10-minute data lapse and processing for graphical display). Further analyses (not shown) show a fairly well-mixed layer extending from above the PBL or frontal zone to about the 5-km level where relative humidity reached a maximum. Our interpretation is that this is a dry "high-plains" PBL from the preceding day. Clouds developed just above the 5-km level after 1000 UTC with light showers reaching the surface after lidar operations were terminated at 1120 UTC. This mid-level cloudiness was observed by satellite (Geostationary Operational Environmental Satellite) as a cloud band oriented parallel to (and well behind) the surface front and was tracked by satellite and National Weather Service (NWS) radar from western Kansas.

An interesting structure is also seen in the post-frontal cold air mass where there is an indication of a developing PBL (moistening) near the surface. The moisture pattern around

the 1-km level is suggestive of roll circulations (0900-1000 UTC). Rawinsonde data at 0900 UTC indicate this minimum level in water vapor mixing ratio was associated with an internal stably stratified layer that separated the developing PBL (becoming well-mixed) from the upper region of the post-frontal air mass that is also fairly well-mixed and capped by the stable frontal zone.

The vertical profile data obtained by the Sandia Raman Lidar (Goldsmith et al., this volume) at 1-minute temporal resolution yields an even better picture of the wave structure associated with the frontal passage (Figure 2). These observations indicate a displacement amplitude of about 1 km or more for the first wave diminishing to about 400 m by the fourth wave. The period of the first two waves was about 9 minutes, while the trailing waves appeared shorter (about 7 minutes). Assuming propagation at about the same speed as the cold front (15 m s^{-1}) yields wavelengths of about 8 and 6 km, respectively. Our interpretation of this wave phenomena is that it is an undular bore induced by the piling up (deepening) of the moist pre-frontal boundary layer immediately ahead of and resulting from the propagation of the cold front. This interpretation is supported by observations of a sudden strong pressure jump, surface wind gust and direction shift, and the rolling cloud bands (Stull 1988). Bores have been previously observed in association with thunderstorm outflow boundaries (Koch et al. 1991) and propagating sea breeze fronts (Smith et al. 1982).

Ancillary Observations - A Key to Understanding

Interpretation of even continuous high-resolution vertical-profile data may be ambiguous in complex situations as on this night. The Oklahoma Mesonet, a unique dense network of more than 110 automated surface stations making routine

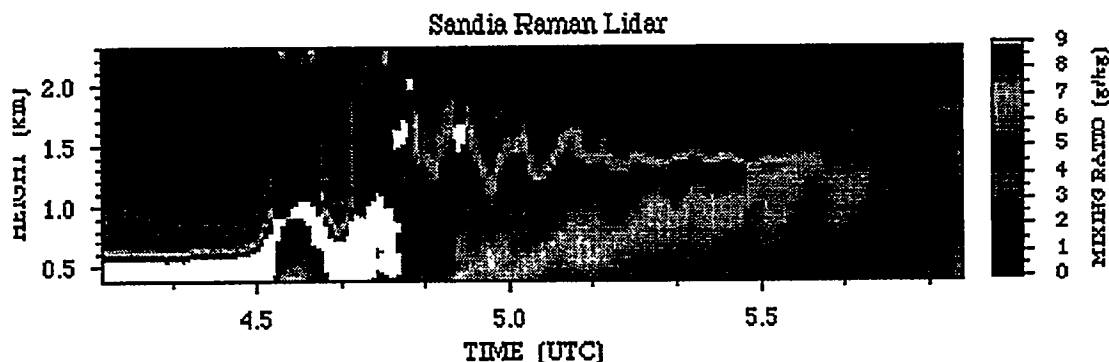


Figure 2. Observations of water vapor mixing ratio profiles (preliminary calibration) by the Sandia National Laboratory Raman Lidar on 15 April during the 1994 ARM RCS IOP.

5-minute observations, provides a wealth of complimentary meteorological data. In comparison, there are 16 NWS surface stations (half-hourly) and 3 ARM surface sites (1-minute data) in the state of Oklahoma (plus 2 ARM sites in southern Kansas). These surface data, together with data from the ARM rawinsonde sounding sites and the NWS profiler network, may be used to greatly enhance understanding of phenomena observed at the CART site. This case illustrates the high utility of these data (all available through the ARM data system).

Lidar observations indicated strong moistening and a rapid increase in the depth of the boundary layer about 2 hours prior to the frontal passage (left of Figure 1) when lidar operations were just commencing. Comparison to the 0300 UTC rawinsonde sounding (launched at 0230 UTC) more clearly shows the strong change. The next sounding (0600 UTC) was after the frontal passage (soundings not shown). Thus, even with the enhanced 3-hourly soundings at the CART site, this significant change in the pre-frontal boundary layer was not resolved. A strong thunderstorm with copious lightning was also visible during this time on the northeastern horizon from the CART site. Initially, the dramatic passage of the leading edge of the frontal zone with strong wind gusts and roll clouds could easily have been mistaken for a thunderstorm outflow boundary.

Analysis of the mesonet data provides a clear picture of what occurred (Figure 3). Shown are streamline analyses of the surface wind observations superposed on an analysis of the observed dew point temperatures ($^{\circ}\text{F}$). We have added our analysis of frontal position and also indicated the locations of the ARM sites. At 0100 UTC, the cold front was just penetrating into extreme northwestern OK. A convergence line is evident with a roughly N-S orientation through west-central OK and with southerly flow to the east. Also very evident is a strong moisture gradient through central OK that was oriented SSW-NNE with dew points near 70°F to the east transitioning to values near 50°F to the west and lower still (30°F) in the far west. The change to a more moist and deeper pre-frontal boundary layer seen in the Raman lidar observations (Figure 1) is a reflection of the northward movement of the moisture gradient on the southerly flow in advance of the front (Figure 3 at 0430 UTC). The 0600 UTC sounding from Morris in east-central OK shows the deep moist and warm boundary layer momentarily seen at the CART site. The intrusion of the moist boundary layer did not extend much to the north or west of the CART site. It is apparent that the subsequent coincidence of the convergence zone, the moisture gradient, and the front (wind

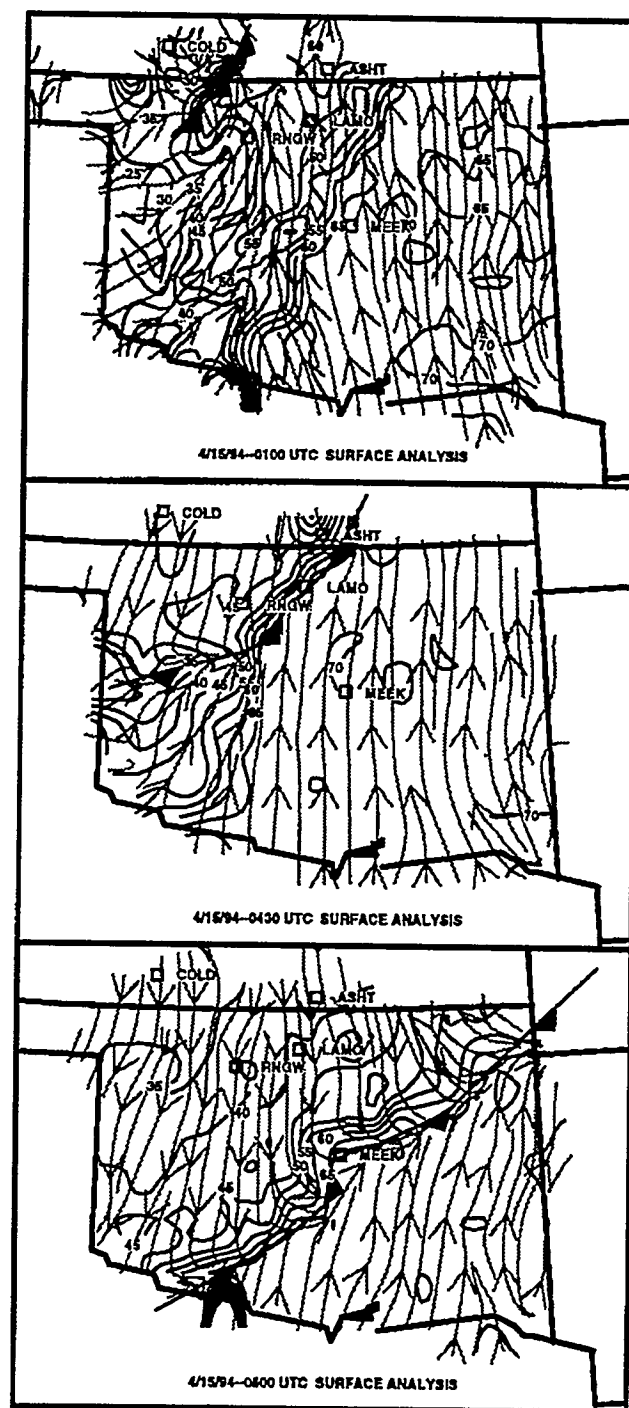


Figure 3. Streamline and dew point temperature ($^{\circ}\text{F}$) analysis of surface observations from the Oklahoma Mesonet, NWS, and ARM stations on 15 April 1994.

shift and temperature gradient) in central OK evolved from three distinct and separate earlier features. Thus, our influenced the observations at the CART site, although the frontal propagation speed was enhanced in northeastern OK (from about 15 m s^{-1} to 20 m s^{-1}).

Characterization of Structure

One goal of this project is to develop quantitative descriptions of subgrid-scale water vapor structure and variability in relationship to large-scale parameters (e.g., static stability and wind shear) that would be useful for validating or improving parametric treatments of subgrid-scale cloudiness or latent heat flux employed in mesoscale models or general circulation models. The rawinsonde and wind profiler data from the ARM and NWS networks enable characterization of the large-scale environment, while the Raman lidar provides a high-vertical and high-temporal resolution description of the water vapor field.

Because of the well-developed wave structure evident in the water vapor field (Figure 2), the 15 April case is very useful for evaluating techniques to quantify variability. Spectral analysis of the water vapor mixing ratio time series data at various levels does not reveal strongly dominant wavelengths. Minor peaks were found at periods of just greater than 7 and just less than 9 minutes. Wavelet analysis (Weng and Lau 1994) yields a very useful alternative characterization of the wave structure (Figure 4). Here, the results

depict the relative amplitude of local water vapor undulations (averaged over a 375-m layer about the 0.69-km level) as a function of time and wavelength delineated by frequency octave where a period of 12.5 minutes corresponds to the zero octave and doubles (halves) for unit octave increment. Very apparent is the wave structure at a period of 8.5 minutes seen in the alternating pattern of highs and lows between the 0 and -1 octaves that is especially strong around 0430 UTC (peak/high near 4.5 hours). This phenomena persists, though with moderated amplitude, until 0600 UTC. The period of the higher frequency but weaker wave structure (between the -1 and -2 octaves) corresponds to the sampling interval (2 minutes). The power seen at lower frequencies (between octaves 1 and 3) corresponds to the gross changes associated with the transition from the relatively dry pre-frontal conditions at this level to the very moist bore zone to the frontal zone itself, and then to the post-frontal air mass. Analyses at other altitudes yield very similar results, but with some low-frequency differences depending on the sequence of gross events at the level, and with indications of a second important oscillation period of about 18 minutes at higher levels.

Other techniques, including a combined spectral analysis, structure function analysis, and singular measure analysis to derive a multifractal characterization (Davis et al. 1994) of water vapor variability and textural measures of variability have been applied to data from this case and others observed during the ARM RCS IOP and the 1991 First International

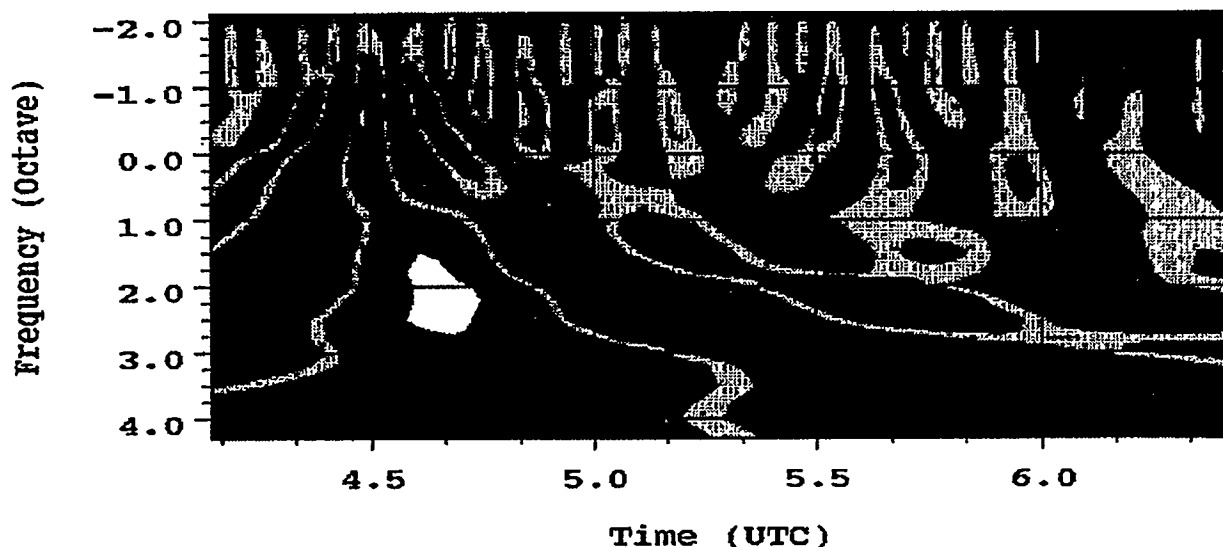
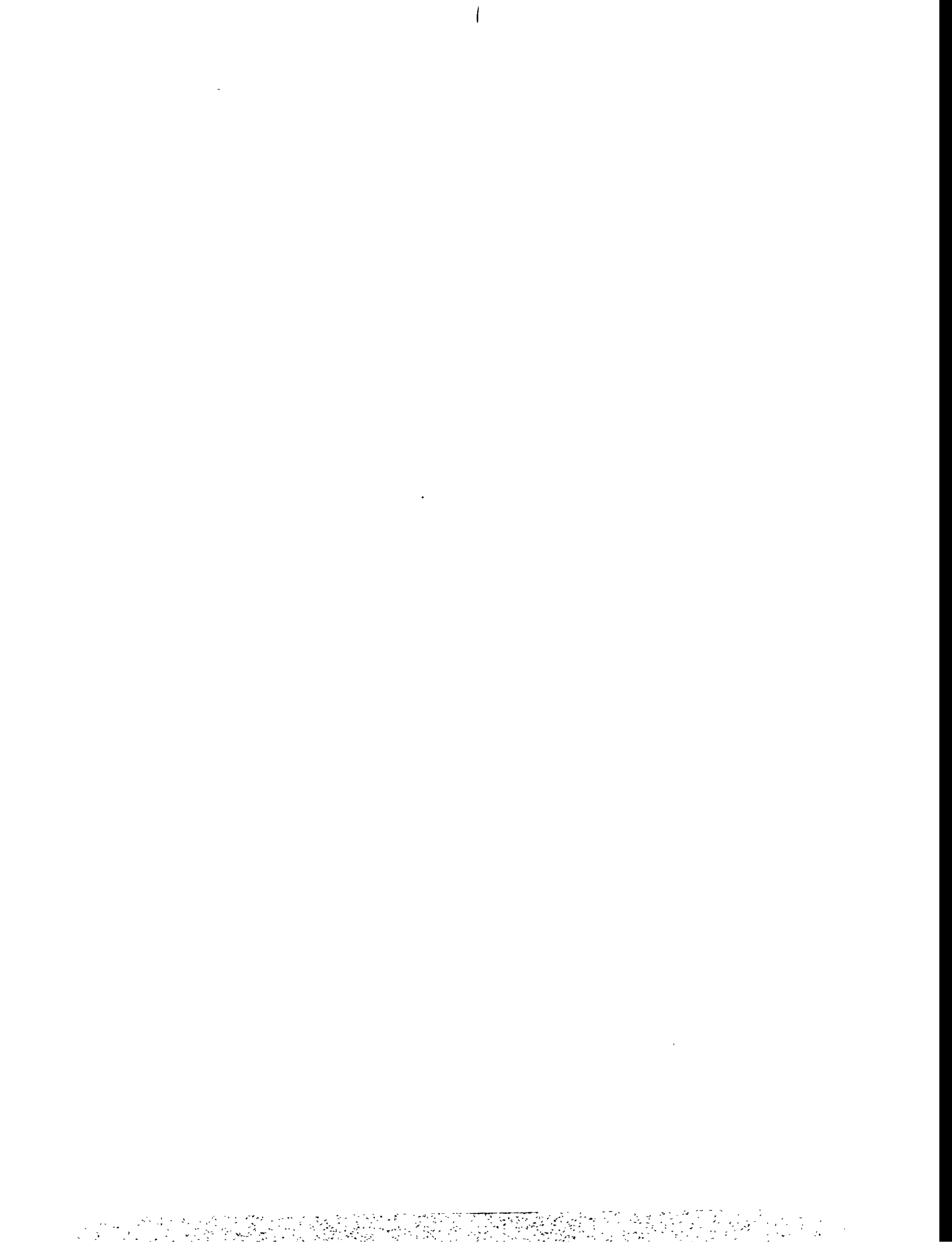


Figure 4. Wavelet analysis of GSFC Raman lidar water vapor mixing ratio observations, vertically averaged over 375-m layer centered on a height of 0.69 km (Figure 1), around the time of a cold front passage on 15 April 1994 at the ARM CART site in Oklahoma during the ARM RSC IOP (see text).

Satellite Cloud Climatology Project Regional Experiment/Spectral Radiance Experiment field campaign in Coffeyville, Kansas, in conjunction with analysis of the sonde and profiler data. The preliminary results (shown on poster) are encouraging. Specifically, the derived measures (stationarity or smoothness and intermittency or singularity) are found to encompass a limited but potentially useful range of the parameter space (i.e., the measures of variability do vary between cases or with altitude or with time which is requisite for producing a useful model of water vapor variability as a function of large-scale environmental conditions).

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Summertime Low-Level Jets Over the Great Plains

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Introduction

The sky over the southern Great Plains Cloud and Atmospheric Radiation Testbed (CART) site of the Atmospheric Radiation Measurement (ARM) Program during the predawn and early morning hours often is partially obstructed by stratocumulus, stratus fractus, or cumulus fractus that are moving rapidly to the north, even though the surface winds are weak. This cloud movement is evidence of the low-level jet (LLJ), a wind speed maximum that occurs in the lowest few kilometers of the atmosphere. Owing to the wide spacing between upper-air sounding sites and the relatively infrequent sounding launches, LLJ evolution has been difficult to observe adequately, even though the effects of LLJs on moisture flux into North America are large (Rasmussen 1967).

Beginning in the spring of 1994, a 915-MHZ radar wind profiler became operational at the ARM CART site near Lamont, Oklahoma (Stokes and Schwartz 1994). The hourly profiler data from this site show that the LLJ is a common feature of the nighttime environment over the southern Great Plains during the summer. A total of 56 LLJs are indicated in the available data between 1 June and 2 October 1994. Even though the level of maximum wind speed within the LLJ occurs throughout a large range of heights (Figure 1), the level of maximum wind speed is below 500-m AGL for 46% of the observed LLJs. This indicates that summertime LLJs frequently develop very close to the ground surface and therefore are impossible to observe accurately with the National Oceanic and Atmospheric Administration (NOAA) 404-MHZ radar wind profilers as presently configured with the lowest data level at 500 m above ground level (AGL).

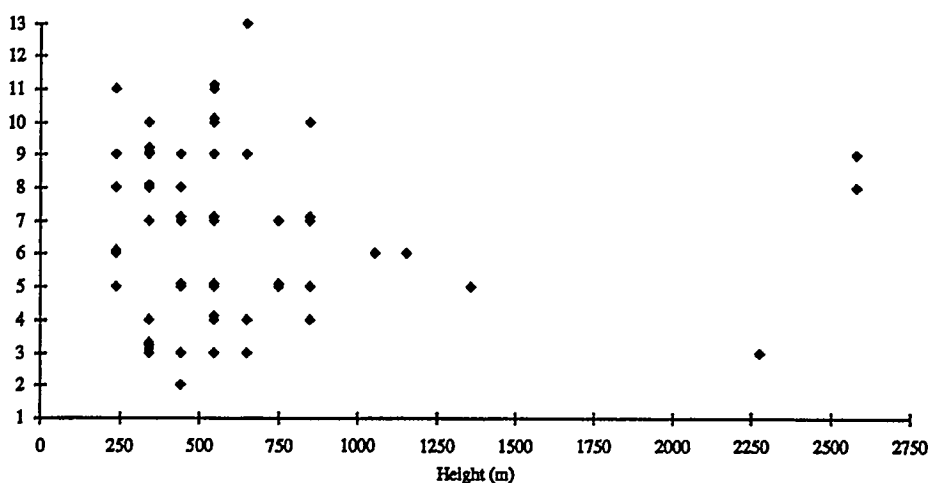


Figure 1. Plot of the time of maximum wind speed (UTC) from the low-level jets observed using the ARM program 915-MHZ radar wind profiler vs. the height of this maximum wind speed (m).

This lack of routine wind observations within the lowest 500 m AGL presents a significant challenge to the accurate observation of the LLJ. Fortunately, a number of WSR-88Ds^(a) are operational across the Great Plains (Figure 2). Radial velocity data from these radars are used routinely to calculate the horizontal wind components from the Velocity-Azimuth Display (VAD) technique (Browning and Wexler 1968) every 304-m mean sea level (MSL) in the vertical and at 30-min time intervals. Therefore, radar coded messages (RCMs), which contain the VAD wind information, are examined in order to better assess the diurnal and horizontal structure of the LLJ over the southern Great Plains.

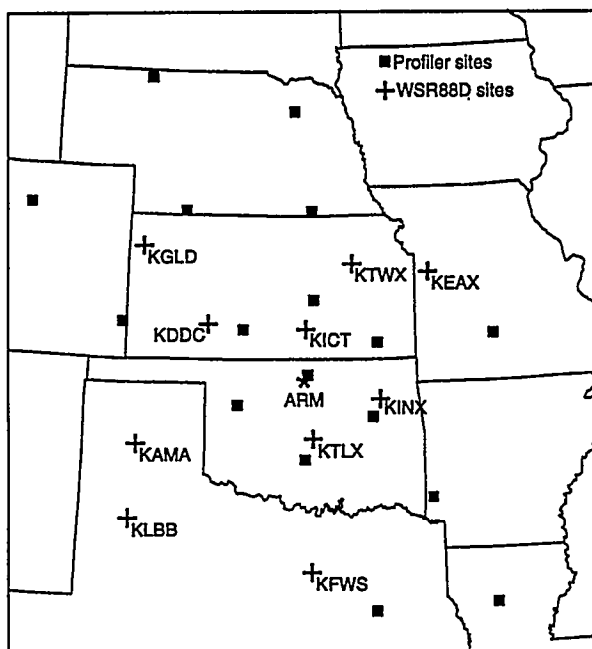


Figure 2. Locations of the WSR-88D radar sites, the ARM 915-MHz radar wind profiler, and the NOAA 404-MHz wind profilers.

Observations

During June 1994, the ARM profiler data documented the presence of 28 LLJs out of 29 days with data. Similar numbers of LLJ cases are found in an examination of the VAD winds from the WSR-88Ds, with similar structures in the LLJ observed by the ARM profiler and the nearest several WSR-88Ds. In general, these data suggest that the LLJ forms earlier in the western plains than in the eastern plains, although this is not true for every case. As an

example, data from 4-5 June show a LLJ stretching from Texas to Kansas with the strongest wind speeds located in western Kansas (Figure 3). While the LLJ is well-developed by 0600 UTC, it does not reach maximum amplitude across much of the plains until 0900 UTC. The time of the wind speed maximum within the LLJ increases from 0900 UTC at Dodge City, Kansas, to near 1200 UTC at Twin Lakes, Oklahoma, as winds veer throughout the night.

Model Simulation

The development and evolution of these summertime LLJs potentially are a challenge to numerical weather prediction models, owing to the finite number of vertical model levels, difficulties in parameterizing the nocturnal planetary boundary layer, and the lack of data with sufficient horizontal and vertical resolution to create a realistic model initial condition. To assess the ability of present numerical weather prediction models to simulate the complex structure of the LLJ in the Great Plains region, the 4-5 June event is simulated using the Penn State - National Center for Atmospheric Research (NCAR) mesoscale model. This

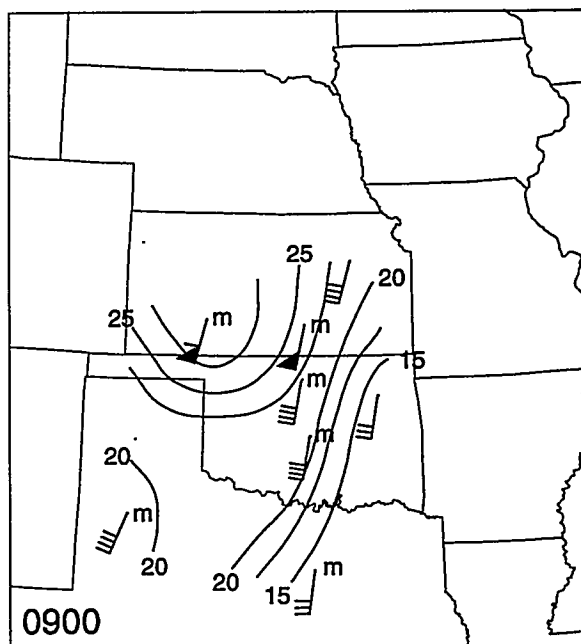


Figure 3. Plot of maximum wind speeds within the lowest 1.5 km of the atmosphere at 0900 UTC 5 June 1994. Each barb represents 5 m/s. Letter m indicates that the wind speed at this time was the maximum observed for this event.

(a) Weather Surveillance Radar - 1988 Doppler

model includes the Kain-Fritsch convective parameterization scheme, a Gayno-Seaman 1.5 order turbulence closure scheme for the planetary boundary layer, a surface energy budget that includes the effects of clouds, and explicit microphysics (see Stensrud and Fritsch 1994 for details). The model simulation from 4-5 June reproduces many of the structures deduced from the remotely-sensed wind observations (Figure 4). The model places the maximum wind speeds within the LLJ near DDC as observed, has the correct orientation of the shear zone, has the time of maximum LLJ wind speed earlier to the west than to the east, and has the level of maximum wind speed higher in Kansas than in either Oklahoma or Texas. However, the model winds are everywhere slightly less than observed and the model fails to reproduce the strong low-level winds at Lubbock, Texas. Thus, while the model captures much of the general LLJ structure, it misses several details.

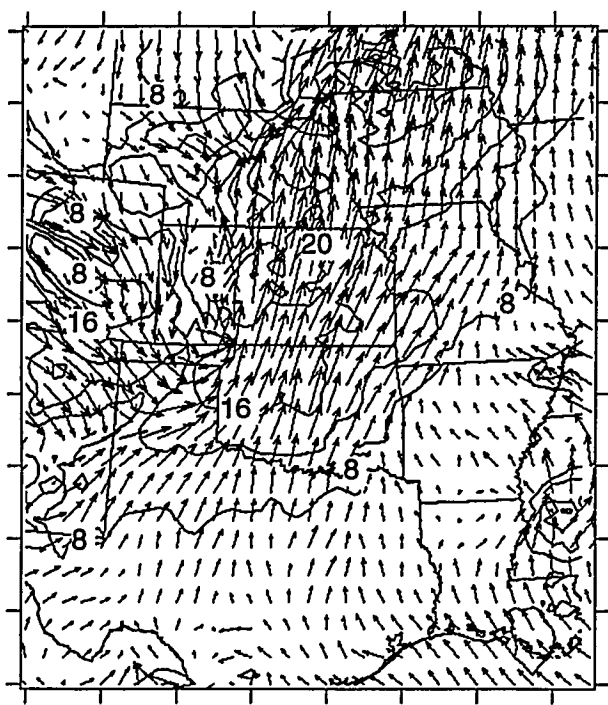


Figure 4. Model wind vectors with wind speed contoured every 4 m/s at 0900 UTC 5 June 1994. Model sigma level of 0.95.

In addition to examining the ability of the model to simulate the LLJ, the model also is used to identify the physical mechanisms of LLJ development and evolution. Low-level jets over the Great Plains have been hypothesized to be produced by inertial oscillations (Blackadar 1957),

baroclinicity over sloping terrain (Holton 1967), boundary currents (Wexler 1961), and isallobaric forcing (Uccellini 1980). Often none of these mechanisms alone can explain the observations. However, during the summertime, when the influences of large-scale systems are reduced, it may be that the important mechanisms for LLJ development can be identified. In addition, it is possible that these mechanisms will be different in different regions of the Great Plains, depending upon the underlying terrain slope and typical boundary layer structures.

To investigate the importance of the various mechanisms of LLJ development, all of the tendency terms from the model momentum equations are saved every 1.5 min for selected grid points (Figure 5). The v-component tendencies calculated from grid points near DDC indicate that the pressure-gradient force is the most important mechanism for LLJ development on this day, with both horizontal and vertical advective components acting to offset one another after the simulation has been going for 12 hours. In contrast, the v-component tendencies at TWX, located farther to the east, indicate that the Coriolis and pressure-gradient forces drive the acceleration of the wind (not shown). This result, while preliminary, does lend some credence to the idea that the mechanisms of LLJ development can be ordered according to importance based upon where the jet develops during the summertime.

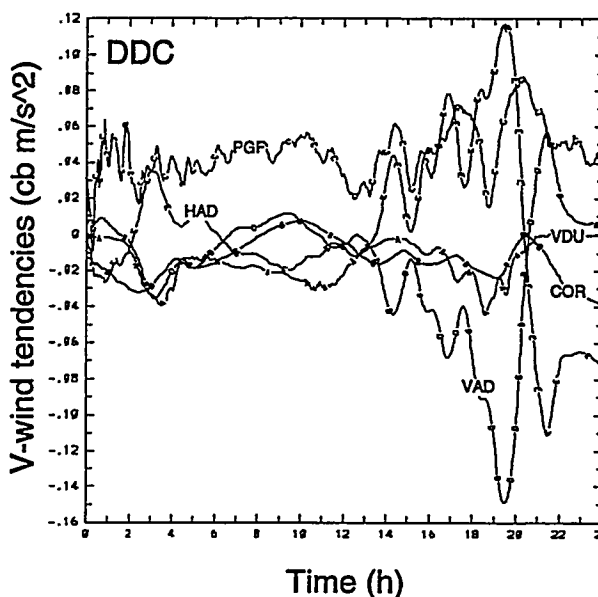


Figure 5. Average model grid point tendencies at sigma 0.95 near DDC from the pressure gradient force (PGF), horizontal advection (HAD), vertical advection (VAD), vertical diffusion (VDU), and Coriolis (COR). Units in cb m s^{-2} .

Discussion

The simulation of one LLJ case during June 1994 has been examined and shown to reproduce many of the observed features documented using the remotely sensed winds from the ARM 915-MHZ wind profiler and the WSR-88Ds. In order to better understand LLJ development and the ability of present numerical weather prediction models to simulate the LLJ, the entire month of June 1994 will be simulated and the model results compared with observations. This study will present the most detailed look to date at summertime LLJ structures over the southern Great Plains that are so important to the water vapor flux into and within North America. The identification of the most important mechanisms for LLJ development and evolution should lead to improved representations of LLJs in global circulation models, since the important physical processes necessary for jet development during the summertime can then be incorporated into the general circulation models.

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Microphysical Structure of Simulated Marine Stratocumulus: Effects of Physical and Numerical Approximations

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Introduction

Over the past decade or so the evolution and equilibria of persistent decks of stratocumulus climatologically clinging to the edge of summertime subtropical highs have been an issue of increased scientific inquiry. The particular interest in the microphysical structure of these clouds stems from a variety of hypotheses which suggest that anthropogenic influences or biogenic feedbacks may alter the structure of these clouds in a climatically significant manner (e.g., Twomey 1977; Baker and Charlson 1990; Albrecht 1991; Ackerman et al. 1993; Pincus and Baker 1994).

Most of these hypotheses are quite tentative, based as they are on simple formulations of boundary layer structures and interactions between drops and aerosols. However, given the potential importance (from the perspective of the global energy balance) of the nature of the equilibrium of the cloud topped boundary layer, there is considerable interest in flushing out their validity; whether it be through ingenious observations or detailed numerical simulations of the range of interactions involved in a particular hypothesis.

Because the hypotheses discussed above are fundamentally coupled to the dynamics of the droplet spectra, their detailed simulation requires that both the droplet distribution function and its interaction with the aerosol distribution function be explicitly represented. Over the course of the past few years two groups (e.g., The University of Oklahoma group [Kogan et al. 1994] and the Colorado State University group [Feingold et al. 1994]) have begun to address this requirement by fully representing the turbulent evolution of the boundary layer via a large eddy simulation (LES) representation of the dynamics coupled to a detailed or explicit representation of the microphysics. Hereafter, the

model which results from coupling an explicit microphysics (EM) component to the LES model will be referred to as an LES-EM model. The advantage of such an approach is that it provides a consistent representation of the evolution of the cloud dynamical and microphysical structure. Its major drawback is that in order to capture even the most rudimentary interactions between drops and ambient aerosol (e.g., cloud drop activation), several million degrees of freedom must be represented.

Given this background, the present work concerns itself with an assessment of the microphysical structure of marine stratocumulus as simulated by an LES-EM model. The purpose of this assessment is to address the following questions (which we believe to be relevant toward any attempt to evaluate, on the basis of numerical experiments, the hypotheses outlined above): 1) To what extent is the microphysical structure of the simulated stratocumulus similar to that observed? 2) To what extent is the simulated microphysical structure physically consistent? and 3) What is responsible for the structure of the simulated microphysical fields?

Method

In addressing the questions outlined in the introduction we make use of a family of models. Our approach is to use the LES-EM model to generate two sets of information. The first is the structure of an ensemble of trajectories which characterize the boundary layer circulations. The second is the microphysical structure of the cloud as predicted by the LES-EM model. The ensemble of trajectories is then used to drive a trajectory ensemble model (TEM) with different degrees of complexity in the representation of the microphysics. To the extent that the ensemble of

trajectories dynamically characterize the simulated cloud, we can compare the microphysical structures predicted by the differing model environments and isolate the effects of a variety of model assumptions, or physical processes.

A detailed description of the models used is given by Stevens et al. (1996a), as well as an overview of the initial conditions, grid configurations, and boundary conditions. The present investigations are limited to the condensation nucleation process only, as this formulation uses the TEM to its best advantage.

Fields

In Figure 1 the layer-mean as well as the up/downdraft-mean simulated droplet concentrations are plotted. In contrast to the gradual and approximately linear increase in liquid water content (not shown), the number concentrations are quite blunt. Such a picture is consistent with activation of cloud drops at cloud base and no subsequent activation through the depth of the cloud. Increases in liquid water content are absorbed by a shifting of the mode diameter on a constant droplet population, as evidenced by the increase in average diameter (not shown) through the depth of the cloud. All of this is in qualitative agreement with the classical picture of

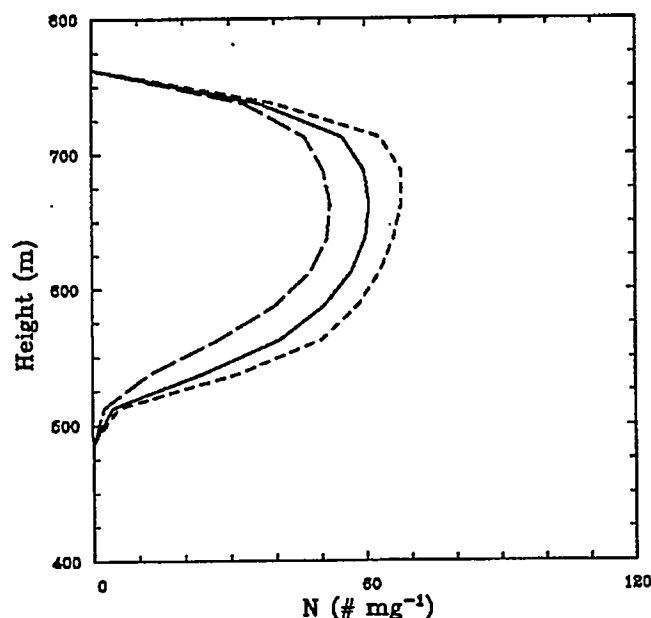


Figure 1. Ensemble average profile of number concentration field predicted by LES-EM model. Layer mean (solid line), updrafts (short-dash), downdrafts (long dash).

stratocumulus microphysical structure developed on the basis of numerous observations (e.g., Slingo et al. 1982; Noonkester 1984; Nicholls 1984).

Although supersaturations in clouds are not directly measurable, several investigators have made indirect estimates of cloud base supersaturations. For instance, Hudson and Frisbie (1991) found that, for stratocumulus measured during the First International Satellite Cloud Climatology Project Regional Experiment, median effective supersaturations were between 0.24 and 0.42%, while Martin et al. (1994) show results which limit cloud base supersaturations in stratocumulus to under 0.8% with median values of order 0.4%. Such results are consistent with the predictions of the model (see updraft average in Figure 2). Moreover, if we estimate the supersaturations produced by the model using our knowledge of the specified cloud condensation nuclei (CCN) spectra and the mean value of the activated fraction of CCN, we obtain an effective cloud base supersaturation of 0.25-0.30%.

The mean supersaturation plotted in Figure 2 increases with height, but is negative through the depth of the cloud, indicating that the important structure of the supersaturation field is not revealed in the mean. More information may be extracted by conditionally sampling over up and downdrafts, which through the depth of the cloud are positively and negatively supersaturated, respectively. Furthermore, the

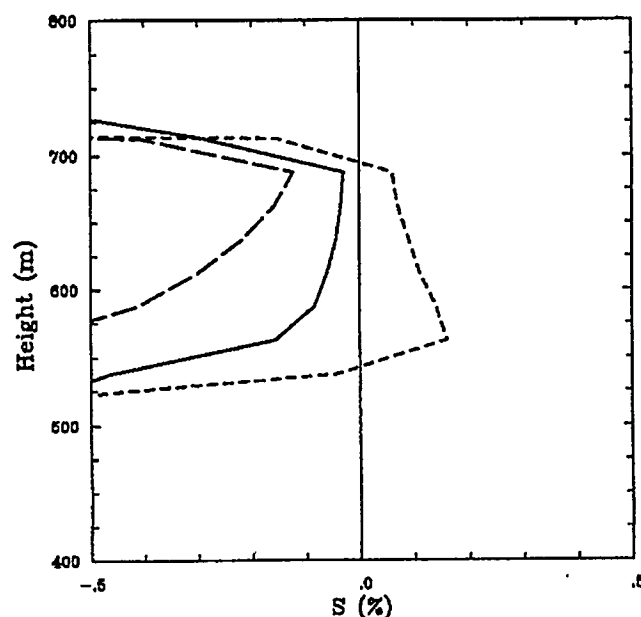


Figure 2. Same as Figure 1, but for supersaturation field.

supersaturation in the updrafts is peaked near cloud base and falls off through the depth of the cloud as one expects based on classical theories.

Updrafts are considerably moister than downdrafts so that departures from adiabaticity in the mean are largely a function of the quantity of entrained air in the downdrafts. Furthermore, because we are considering layer averages, the presence of dry air in downdrafts at a given level also leads to observed decreases in the drop concentrations (see Figure 1) and the liquid water content (not shown).

Noonkester (1984) plotted profiles of dispersion (standard deviation normalized by the mean) from his observations of stratocumulus off the Californian coast. Using these as a basis for comparison we have constructed similar profiles for the number concentrations, liquid water, and mean diameter. Qualitatively we found that the model generated a similar structure in the dispersion profiles, with cloud top and cloud base maxima. Quantitatively the LES-EM simulations also agreed quite well, although the predicted incloud minimum in diameter dispersion was slightly less while the number dispersion was slightly higher than observed by Noonkester.

Sensitivity to Modelling Framework

Using the ensemble of trajectories produced by the LES to drive the TEM with different microphysical representations, we were able to examine the effects different model assumptions had on the simulated fields. We found that the effect of spatial averaging implicit in the Eulerian framework led to a slight underprediction of cloud base supersaturations. As a result, fewer drops were activated.

Two other factors led to a significant underprediction in both the number of activated drops and the cloud base supersaturation. The first was the neglect of gas kinetic effects in our condensational routines. This neglect allowed cloud base drops to grow more rapidly than they otherwise would have, thereby decreasing the phase relaxation time for the parcel and mitigating the production of cloud base supersaturations. Rooth (1957) explained this basic process; nevertheless, we were surprised at the magnitude of the differences (of order 20%). The second was the assumption that drops once activated immediately grew into the first bin. Similar to the neglect of gas kinetic effects, this assumption decreases the phase relaxation time too rapidly and underestimates the number of activated drops by 5-10%.

In addition, we found that the manner in which mixing is done in Eulerian models leads to two related, yet spurious results. First is the generation of cloud top peaks in the supersaturation and second is an underestimation in the number of drops in the downdrafts in the LES-EM data relative to the TEM data. These are not artifacts of the numerical operators (see Stevens et al. 1996b for more details), as they would occur even in the context of perfect advection. Instead they appear to be fundamentally related to the fact that the continuous evolution of a cloud boundary is being represented within a discretized spatial domain.

With respect to numerical dispersion, we found that the Eulerian representation of physical space in the LES-EM model tended to generate reduced values of droplet dispersion in the interior of the cloud. In contrast, the Eulerian representation of mass space generated values of diameter dispersion which were too large, especially in the downdrafts.

Nature of Contributing Trajectories

In order to understand how different trajectories contribute to the simulated cloud microstructure, we conducted 4 sensitivity tests. All simulations were conducted with the TEM coupled to the Lagrangian microphysical component with gas kinetic effects included. In the first sensitivity run (experiment S1) we turned off the mixing so the conserved variables were constant at their initial condition along the trajectories. The second sensitivity run (experiment S2) differed from the first in that the initial conditions of the ensemble of trajectories were set to the mean initial conditions. The third sensitivity run (experiment S3) differed from the second in that the vertical velocity was kept fixed along the trajectory. The fourth and last sensitivity run (experiment S4) differed from the third in that we only collected statistics along an updraft segment of one of the trajectories.

Table 1 shows how the dispersion in different fields responded to the experiments. Data were taken from a level in the interior of the cloud, near the minimum value of the dispersion. The baseline dispersion associated with the single parcel stems from the fact that data were collected over a finite height interval (7 m) so some of the dispersion is associated with variances amongst parcels over slightly different levels. Nevertheless, this table shows that the majority of the variance in the cloud liquid water and drop spectrum is due to mixing, where most of the mixing occurs

Table 1. Values of dispersion for liquid water, number concentration, and diameter from different simulations at 650m.

Experiment	σ_{q_L}/q_L	σ_N/N	σ_D/D
Control	0.252	0.138	0.129
S1	0.110	0.116	0.071
S2	0.017	0.116	0.059
S3	0.019	0.014	0.042
S4	0.014	0.000	0.040

at cloud top. The number concentration dispersion receives its largest contribution from the inhomogeneity in cloud base vertical velocities leading to the activation of different numbers of drops along different trajectories. This also appears to be important in the droplet dispersion.

Summary and Conclusions

We have demonstrated that an LES model with an explicit microphysics component is able to generate realistic representations of the microphysical structure of stratocumulus. Mean and variance quantities agree qualitatively and, in most cases, quantitatively with available observations.

Using a family of models we have explored the sensitivity of the simulated microphysical structure of stratocumulus to a variety of modelling assumptions. While the simulated fields qualitatively agree with observations, the neglect of gas kinetic effects and the impact of Eulerian representations in mass and physical space lead to significant under-predictions of cloud base supersaturations and the activation of too few drops.

Using a trajectory model driven by an ensemble of trajectories generated from the LES data, we have explored the causes for the dispersion in several simulated fields. In all cases, we found mixing along the trajectories (which is greatest at cloud top) to be the most significant source of dispersion.

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Size Distributions of Boundary-Layer Clouds

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Cumulus Triggering by Surface-Layer Thermals

Scattered fair-weather cumulus clouds are triggered by thermals rising from the surface layer. Not all surface-layer air is buoyant enough to rise. Also, each thermal has different humidities and temperatures, resulting in inter-thermal variability of their lifting condensation levels (LCL). Thus, even some of the rising thermals might not rise high enough to trigger clouds (Figure 1).

For each air parcel in the surface layer, its virtual potential temperature θ_v and its LCL height z_{LCL} can be computed.

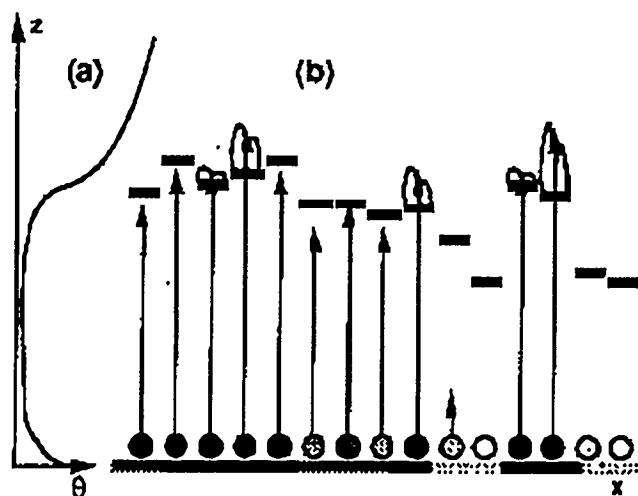


Figure 1. Heterogeneous land use, illustrated with the black and gray line segments at the bottom, cause turbulent temperature fluctuations in the surface-layer air, illustrated with air parcels of different shading. These parcels rise to their level of neutral buoyancy, shown by the vertical arrows. Moisture also varies from parcel to parcel, causing corresponding fluctuations in the height of the lifting condensation level (LCL), shown by the black horizontal dashes. Those parcels that rise above their own LCL can create cumulus clouds.

These variables measure the potential for a parcel to rise, and the amount of rise needed to form a cloud. By sorting all parcels in the surface layer, a joint frequency distribution (JFD) can be computed, giving the number of air parcels having various θ_v and z_{LCL} values. Such a JFD is idealized with the elliptical frequency contours in Figure 2.

If the mean environmental sounding is plotted on the same JFD diagram (Figure 2), then the portion of the JFD that is to the right of the sounding indicates the fraction of thermals that can rise to form cumulus clouds. The shaded region in Figure 2 shows those air parcels that are warm enough to rise and for which their LCL is below their level of neutral buoyancy. This region gives the cloud coverage.

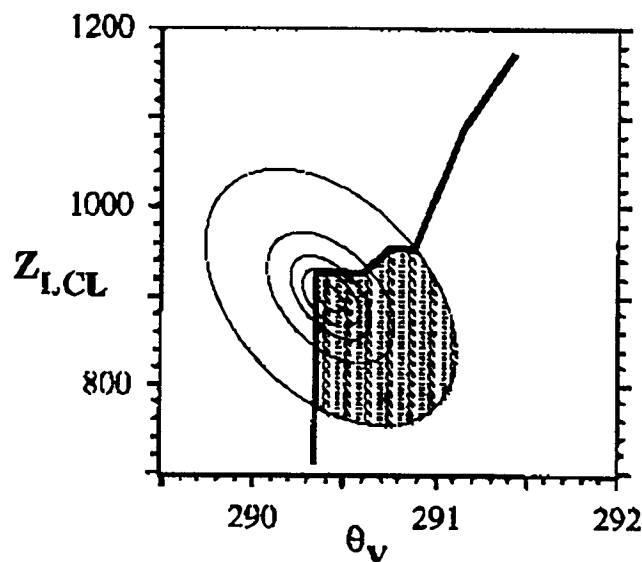


Figure 2. Idealized smooth contours represent values of constant frequency distribution of virtual potential temperature θ_v (K) vs. z_{LCL} (m) for surface-layer air. Superimposed as the heavy line is a plot of the mean environmental sounding of θ_v vs. height z . Shaded region indicates those surface-layer thermals that can form cumulus.

Distributions of Cumulus Bases, Tops, and Depths

Cloud ensemble size distributions can also be determined from the joint frequency distribution. To do this, the joint distribution must be conceptually (Figure 3) divided into three sectors that represent different physics:

1. points cooler than that of the mean mixed layer (ML) temperature (extrapolated vertically across the whole figure)
2. points warmer than the mean ML temperature but which are above the sounding
3. points warmer than the ML but which are below the sounding.

As sketched in Figure 4, Sector 1 indicates the fraction of air that will not likely rise. Sector 2 contains those points that will rise from the surface layer, but which will not likely form clouds. Sector 3 contains those points that rise and form clouds. The range of cumulus base heights and top heights is marked by the starting and ending points of the diagonal arrows in Figure 4.

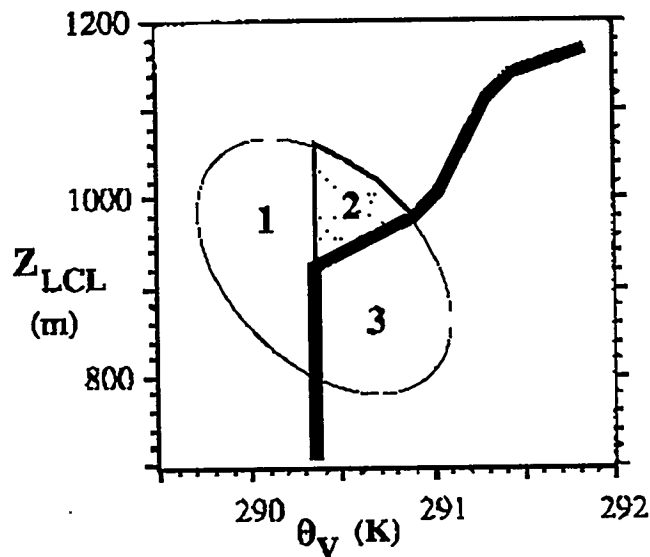


Figure 3. Three sectors of the joint frequency distribution: Sector (1) contains parcels that probably will not rise; (2) contains parcels that will rise without forming a cloud; and (3) contains parcels that will rise and form clouds.

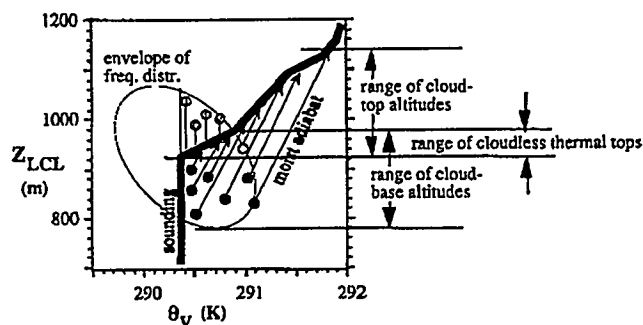


Figure 4. Parcels in Sector 2, indicated with gray shading, rise from the surface layer until they hit the sounding, without forming clouds. The other parcels, colored black, reach their LCL, and then continue rising along a moist adiabat. For many of these latter parcels, such as the one colored white, they rise without condensation past other cloud bases before reaching their own cloud base.

The set of cloud tops and bases for each parcel determines the cloud size distribution. Preliminary results from the Atmospheric Radiation Measurement Cloud and Radiation Testbed (CART) Southern Great Plains (SGP) site give a lognormal depth distribution. Case studies have been made for May 1, 1994, during the spring intensive observation period (IOP), and for July 27, 28, and 31, 1994, during the summer IOP.

Three Layers Associated with Boundary-Layer Cumulus

When the extent of buoyant rise for points from Sectors 2 and 3 of Figure 4 are plotted, taking into account rise along moist adiabats from Sector 3, the result indicates that there are three layers in the cloudy boundary layer (Figure 5). The bottom layer is the well-known sub-cloud layer, where all the updrafts are cloud-free. The top layer is the cloud layer, where all the updrafts are cloudy.

In between is an "overlap" layer, with both cloudy and cloud-free updrafts. This layer is not quite like the transition layer described by Malkus (1958) for tropical cumulus, nor is it quite like the entrainment zone described by Deardorff et al. (1980) or the LCL zone described by Wilde et al (1985). The nature of this overlap zone determines the coupling between the subcloud and cloud layers.

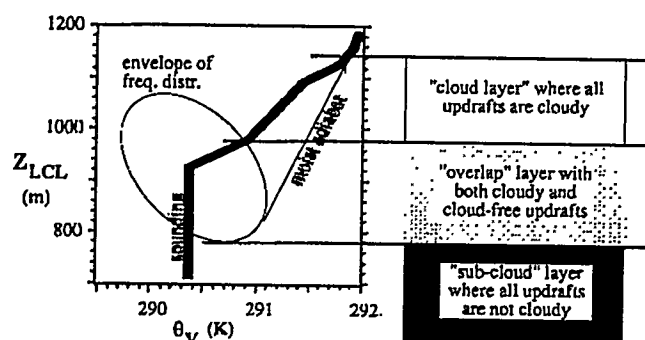


Figure 5. Three layers can be identified in the boundary layer. The cloud and subcloud layers have the conventional definitions of cloudy and non-cloudy air. The middle "overlap" layer has both cloudy and cloud-free updrafts.

For more details about these methods, see the following papers: Schrieber et al. (1996) and Berg and Stull (1996).

Planned Research

The heuristic model will continue to be refined and validated against new IOP data from the SGP CART site. As other field sites come on line, we will test our parameterization for other parts of the globe.

Daily forecasts of cloud attributes will be made by inserting this cloud parameterization scheme into an operational numerical forecast model.

During the upcoming July 1996 IOP, we propose to bring the University of Wyoming King Air aircraft to the SGP CART site to make turbulence measurements in the surface layer. These will be used to define the JFD for that area.

Acknowledgment

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Inferring Spatial Cloud Statistics from Limited Field-of-View, Zenith Observations

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Introduction

Many of the Cloud and Radiation Testbed (CART) measurements produce a time series of zenith observations, but spatial averages are often the desired data product. One possible approach to deriving spatial averages from temporal averages is to invoke Taylor's hypothesis where and when it is valid.

Taylor's hypothesis states that when the turbulence is small compared with the mean flow, the covariance in time is related to the covariance in space by the speed of the mean flow. For cloud fields, Taylor's hypothesis would apply when the "local" turbulence is small compared with advective flow (mean wind).

Thus, if Taylor's hypothesis holds

$$R(dt) = R(dx) \text{ and } dx = U dt \quad (1)$$

along the direction of the mean wind, where $R(dt)$ is covariance in time, $R(dx)$ is covariance in space, and U is the speed of the mean wind. That is, the temporal covariance of the cloud field at a fixed location and the spatial covariance along the direction of the mean wind are the same when Taylor's hypothesis holds.

The objective of this study is to determine under what conditions Taylor's hypothesis holds or does not hold for broken cloud fields. We use one-dimensional and two-dimensional correlations to evaluate the applicability of Taylor's hypothesis. By way of illustration, we use a time series of small field-of-view, zenith measurements of cloud cover to estimate the cloud cover fraction over a 112° view of the sky dome. This is analogous to using a time series of ceilometer measurements indicating cloud/no cloud to determine the cloud cover fraction. We identify the conditions under which the time series can yield reliable estimates of the cloud cover fraction. Whole-sky images (WSIs) provide a

useful data set for this evaluation, because the time series cloud cover fraction results can be validated with the 2-D images.

The WSI Data Set

The WSI data set used in this analysis was collected at White Sands, New Mexico, during May 4-5, 1992. Images were acquired using two Scripps, E/O 5 WSIs placed about 5 km apart. This data set was originally used to develop algorithms to determine the cloud base height from paired WSIs (Allmen and Kegelmeyer 1994). The WSI has a fish-eye lens that provides wide field-of-view images (~ 165 degrees) of the sky dome with an angular resolution of $\sim 1/3$ degrees. The images are equal-angle projections with the distance from the zenith in the image being proportional to the zenith angle in the scene. Depending on the cloud height, horizontal cloud distributions from 10 km to 50 km in horizontal extent can be recorded.

A few comments on the character of the WSI images as related to processing the images are appropriate. Because the WSI views clouds from beneath, primarily bottoms of clouds are seen near the center of the WSI image, and progressively more of the sides of the clouds are seen as the view approaches the horizon (i.e., edge of the image). In addition, because of the projection properties of the fish-eye lens, the clouds appear to be more strongly distorted from their apparent shape in the scene, the closer they are to the horizon. Because a common focus is difficult to achieve for a wide field of view, the clouds scenes are slightly blurred at the edge of the image. Also, because of the changing scattering angle, the image is darker away from the sun. These effects will affect the accuracy of the cloud statistical calculation.

Several corrections are done to reduce these effects that would otherwise bias the statistical calculation. First, the dependence of the gray level on the scattering angle is calculated from an overcast stratus cloud image. Based on

the relationship of gray level to scattering angle, a range of thresholds is determined to detect and to filter out bright cloud sides from the image. Then the images are normalized to make the gray level more uniform. Finally, the image is "flattened" using a Pseudo-Cartesian coordinate transformation to reduce the projection distortion of the fish-eye lens.

Testing Taylor's Hypothesis

Taylor's hypothesis is often used in field experiments to save the expense and effort of data sampling over a wide area (Powell and Elderkin 1974). Taylor's hypothesis states that if the turbulence is small compared with the mean wind, then the temporal covariance of the cloud field at a fixed location and the spatial covariance along the direction of the mean wind are the same (Taylor 1938). In Atmospheric Radiation Measurement CART sites, many of the measurements are restricted to zenith views of the atmosphere. To see how the fixed point temporal observation can be used to infer the spatial cloud length scale, Taylor's hypothesis is tested for a variety of cases.

First, the horizontal mean wind of the cloud scene is obtained by using cross-correlation techniques (described in the next section). Then the time series of a fixed point in the image in the down-wind direction is extracted from the image. The correlation function of the time series is calculated and compared with the correlation function of the spatial line sampled along the mean wind direction. Figure 1 shows where Taylor's hypothesis holds [i.e., where $R(dx) \sim R(Udt)$]. The images used for this analysis were acquired during a period of one hour in which individual cumulus clouds with a mean cloud base height of about 2.5 km were distributed in a region of about 10 km in the WSI field of view and the mean wind was near constant and about 3.1 m/s (7.5 pixel/min). The results shown in Figure 1 indicate that in about 125 pixels (3.1 km or 17 min), the temporal and spatial correlation functions are similar. Taylor's hypothesis holds during this period of time and over this distance.

Deriving Cloud Cover Fraction from Time Series Measurements

If Taylor's hypothesis holds, spatial averages can be determined from temporal averages. To illustrate this for broken cloud fields, we determine the cloud cover fraction from the

Spatial and temporal correlation function (P1)

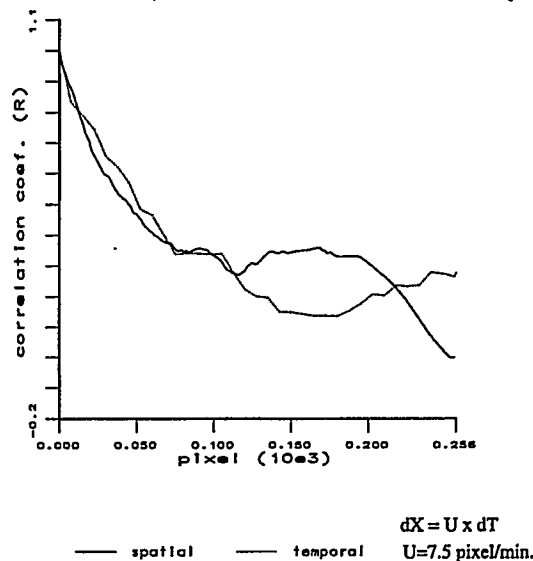


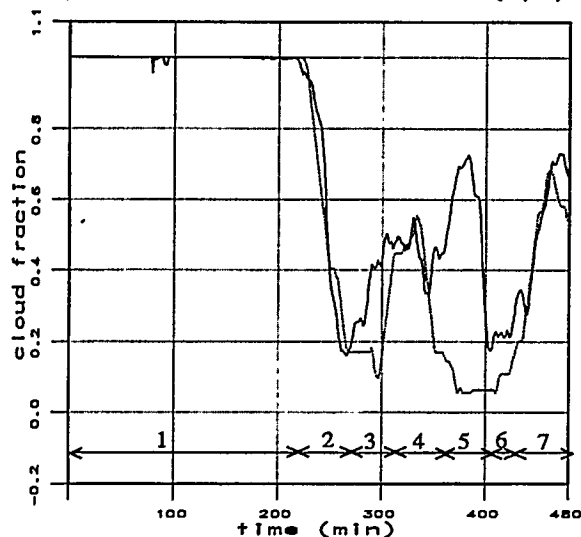
Figure 1. Taylor's hypothesis holds over the range where $R(dx) \sim R(Udt)$, about 125 pixels. Where Taylor's hypothesis holds, time-series averages can be used to derive spatial averages.

WSI data from a small 15 x 15 pixel ($5^\circ \times 5^\circ$) field of view of observation with a moving-window average of 40 minutes. We compare this with the actual cloud cover fraction obtained from wide-field-of-view images (~ 112 degree). Figure 2 shows the comparison for May 4, 1992. The marks, 1-7, represent different clouds and processes types in the time periods indicated. They are

1. St cloud overcast
2. Broken and decaying St
3. Decaying St and some Cu
4. Individual Cu
5. Clustering and vertically developing Cu
6. 1/2 image clear and horizontal wind shear
7. New, big clustering Cu moving in to view.

When Taylor's hypothesis holds, the time series cloud cover approximation is close to the real situation (i.e., the 1/2-spatial average).

Comparison of cloud fraction (5/4/92)



— ws1(56fov) — 15x15pixel(40min)

Figure 2. To test Taylor's hypothesis, we use a time-series in the zenith cloudiness to infer the spatial cloudiness (i.e., the cloud cover fraction). Taylor's hypothesis holds where the two curves are nearly equal.

After analyzing different cases for various cloud types and conditions for the May 4-5, 1992, data, we find the following for the applicability of Taylor's hypothesis.

Mean Wind Determination

We calculate the mean wind from the cross-correlation function of two sequential cloud images. First the cloud base height is calculated from pairs of WSI images which are taken from two WSIs separated by 5 km (Allmen and Kegelmeyer 1994). Next, the cloud horizontal displacement can be found from the spatial lag of the maximum of the cross correlation coefficient from the zero lag position. The mean wind is obtained by dividing the displacement by the time lag. Our results show that while the mean wind velocity can be determined in this way, sometimes the calculated wind speed is slower than the real wind. Additional image correction will have to be done to make this approach more reliable. The possible errors come from some stationary but small population noises. These come from either very bright or dark spots in the original images or are introduced when doing the flattening and pixel interpolation following flattening.

2-D Cloud Correlation Function

In this study, 2-D auto- and cross-correlation functions are calculated by first calculating the fast Fourier transform (FFT) of the two images to be correlated (same image for auto correlation; original and time-lagged image for cross correlation), then multiplying the two images together and applying inverse FFT with appropriate normalization to obtain the correlation function.

The 2-D auto-correlation function of the image can be used to show the spatial morphology of the cloud field. By comparing the auto-correlation at various times, the changes in the cloud field can be shown. Figure 3 shows an example of the 2-D auto-correlation function of a cumulus cloud field. The positive contour lines show the coherent structure of the individual cloud and the negative contour lines show the clear sky. The coherent cloud length scale is about 120 pixels (i.e., about 3 km in this case). This is similar to the results shown in Figure 1 for the 1-D correlations.

Combining the correlation function and the mean wind, we can demonstrate when the Taylor's hypothesis holds.

ACF of Cu (frame=320 5/4/92)

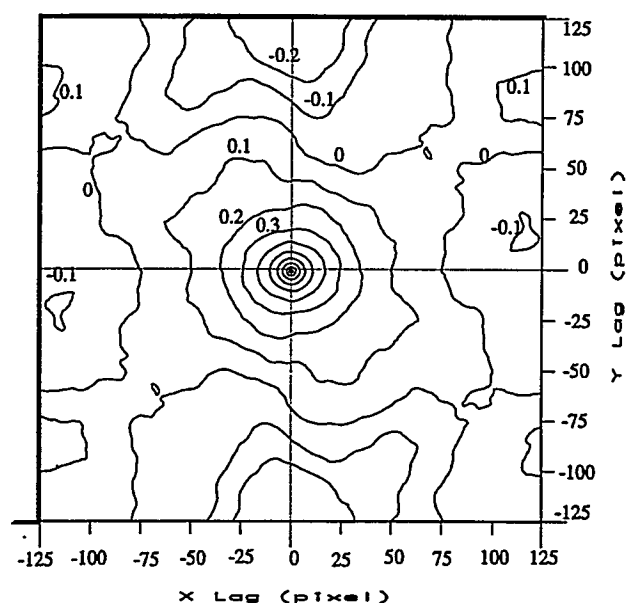


Figure 3. 2-D auto-correlation of a cumulus cloud field. Positive contours indicate the coherent structure of the field. The coherence length is about 120 pixels, consistent with the 1-D correlations shown in Figure 1.

Conclusion

It is often desirable to infer spatial properties of cloud fields from CART zenith, time series observations. This can be done using Taylor's hypothesis when it applies. In this study we have identified some conditions in which Taylor's hypothesis holds and when it does not hold for broken cloud fields. These are summarized in Table 1. In brief we have found that

1. Spatial averages can be inferred from time averages when Taylor's hypothesis holds.

Table 1. Conditions and cloud types for which Taylor's hypothesis holds.

Holds for these conditions:

- Wind has uniform speed and direction
- Statistically homogeneous over horizontal region of greater than 10 km

Generally holds for these clouds:

- Stratus (mostly overcast)
- Individual Cumulus
- Advection of large cumulus

Does not hold for these conditions:

- Wind changing in speed and direction
- When cloud are developing rapidly
- When clouds are dissipating rapidly
- When there is horizontal wind shear
- When more than one cloud deck exists

Generally does not hold for these clouds:

- Cirrus
- Clustering cumulus with vertical development
- Mostly broken stratus

2. Taylor's hypothesis holds for a variety of broken-cloud conditions, generally for cumulus and for mostly overcast stratus.
3. Correlation techniques (giving spatial correlation and mean wind) provide an indication for when Taylor's hypothesis is valid.
4. Correlation techniques can also be used effectively to determine the mean wind speed and direction from WSIs and cloud base height under uniform wind conditions.

Acknowledgments

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Longwave Scattering Effects on Fluxes in Broken Cloud Fields

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The importance of clouds in the radiative energy balance is well known. Since broken cloud fields are prevalent, both the geometry and optical properties of clouds are important parameters. The flat black plate approximation, used by general circulation models for clouds in the longwave, neglects both cloud geometry and optical properties. The error in neglecting geometry has been shown in many works (Ellingson 1982; Harshvardhan and Weinman 1982; Killen and Ellingson 1994).

The single scattering albedo (ω_0) of clouds in the shortwave is 0.99 or higher (Hu and Stamnes 1993). Therefore, most of the previous works on the effects of cloud scattering have been in the shortwave (Welch and Wielicke 1984; Evans 1993; Breon 1992). But the longwave (ω_0) can be as high as 0.75 (Hu and Stamnes 1993), so longwave scattering effects can be significant. In this work, the fluxes above and below a single cloud layer are presented, along with the errors in assuming flat black plate clouds or black clouds.

The predicted fluxes are the averaged results of analyses of several fields with the same cloud amount. Each analysis is in two parts: generating a cloud field and then calculating the fluxes. The cloud size distribution of Cahalan and Joseph (1989) was adapted:

$$\frac{n(a)}{n(a_0)} = \left[\frac{a}{a_0} \right]^{-1.5} \quad (1)$$

The exponent, -1.5, was changed from the original -1.6 in order to simplify the calculation. The clouds were in five cloud size groups (1, 4, 9, 16, and 25 times a_0 non-overlapping and randomly distributed in space. Figure 1 shows the top/bottom view of a cuboidal field; $N = 0.15$. The fluxes were computed using the Monte Carlo method. To simulate an infinite cloud field, cyclical/specular horizontal boundaries were assumed.

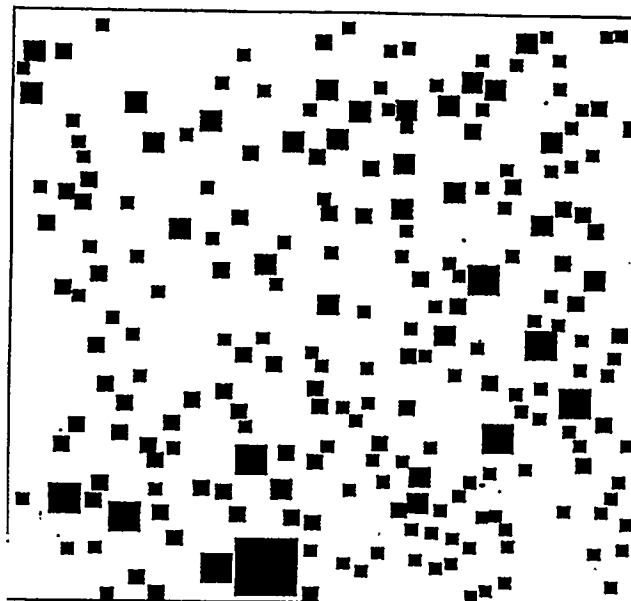


Figure 1. Random cloud field with 15% cloud cover.

Four assumptions were made in order to emphasize the effect of scattering in the atmospheric window as well as simplifying the calculation. First, the gaseous absorption is neglected. Second, the cloud optical properties are constant over the 800 cm^{-1} to 1200 cm^{-1} window region. Third, the cloud field is a single layer with clouds of the same base height, aspect ratio, and either cuboidal or cylindrical. Last, the optical properties and temperature were assumed to be constant over all clouds in the field.

The cloud temperatures were set at 15°C , -10°C , and -50°C , corresponding to low, middle, and high clouds at 0, 4, and 10 km, respectively. Another measure of cloud height is to normalize the integral, over the window region, of the Planck function at the cloud temperature with the same integral at the surface temperature (27°C).

$$\xi = \frac{\int_{8.33 \mu\text{m}}^{12.5 \mu\text{m}} E_{\lambda, b}^{\text{cloud}}(10 \mu\text{m})}{\int_{8.33 \mu\text{m}}^{12.5 \mu\text{m}} E_{\lambda, b}^{\text{ground}}(10 \mu\text{m})} \quad (2)$$

At the cloud temperatures considered, the ξ are 0.822, 0.516, and 0.200.

The above cloud upward flux and the below cloud downward flux were computed for cuboidal and cylindrical cloud fields of various aspect ratio (α), base cloud fraction (N), smallest cloud horizontal optical thickness (L), and single scattering albedo (ω_0). For a given N , results were averaged until the deviation in the cloud absorption ratio was under 2% for a minimum of five fields. Figure 2 illustrates the geometry and relevant variables.

The upward and downward fluxes for cuboidal and cylindrical fields are shown in Figures 3a,b,c,d and 4a,b,c,d. The fluxes at $N=1$ were computed using the independent pixel approximation. The fluxes for black clouds are not shown because the fluxes for absorbing clouds ($\omega_0 = 0$) with $L=100$ are nearly identical. There is little variation between the fluxes for $L=100$ and $L=10$, or between cylindrical and cuboidal cloud fields. The similarity between cylindrical and cuboidal clouds is not too surprising. Both have constant vertical cross section and are indistinguishable when seen from large horizontal distances.

In general, the fluxes are grouped by aspect ratio α for small cloud amounts N . As cloud amounts increase, individual clouds become indistinct. So, at large N , the fluxes are grouped by ω_0 . The variation in the above cloud upward flux with ω_0 for large N decreases with elevation. The below cloud flux variation increases with elevation. The

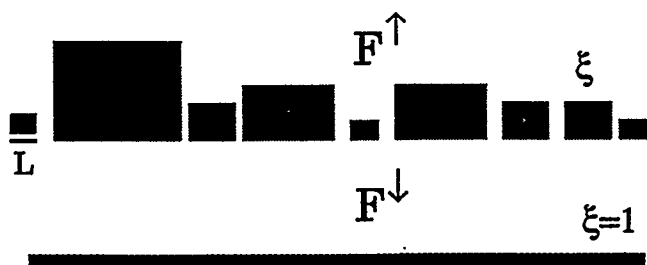


Figure 2. Cloud field parameters and vertical geometry.

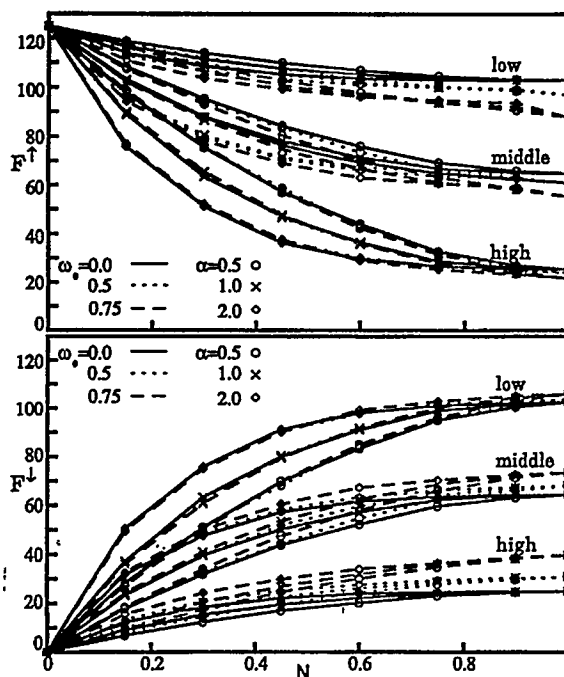


Figure 3a,b. Fluxes (W/m^2) for cuboidal clouds; $L=100$.

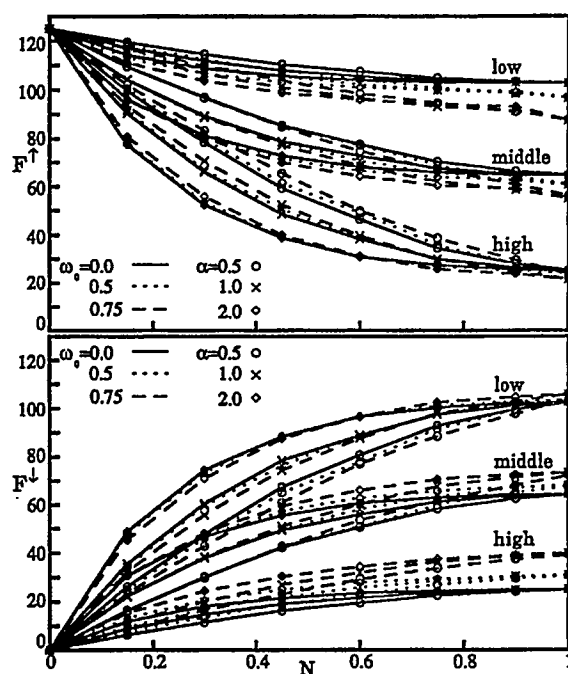


Figure 3c,d. Fluxes (W/m^2) for cuboidal clouds; $L=10$.

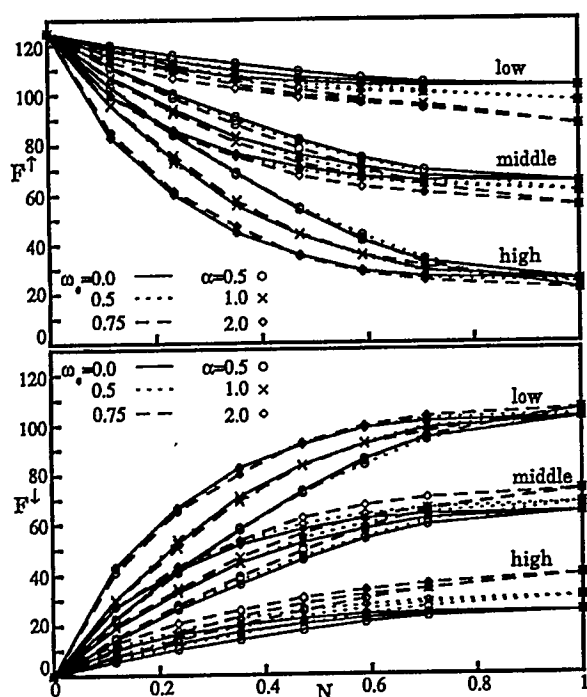


Figure 4a,b. Fluxes (W/m^2) for cylindrical clouds; $L=100$.

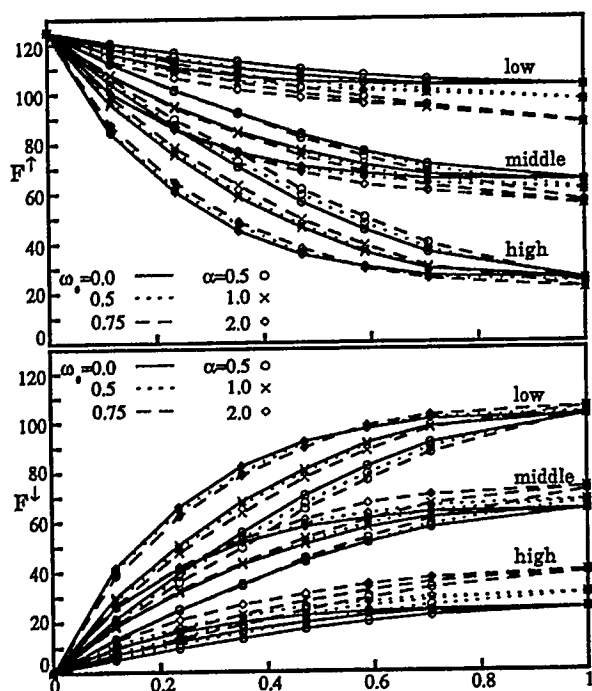


Figure 4c,d. Fluxes (W/m^2) for cylindrical clouds; $L=10$.

magnitude of the variation is about the same; but on a percentage basis, highest for high cloud Ψ_{down} .

In the flat black cloud approximation, the cloud field is replaced by a field of black plates with the same N . In the geometric black cloud approximation, the cloud field geometry remains the same, but the individual clouds become black. Fluxes computed using the flat black plate and geometric black cloud approximations were compared with the computed fluxes. The absolute and percentage flux errors are

$$\delta F = F^{\text{approx}} - F ; \% F = 100 \times \left[\frac{\delta F}{F} \right] \quad (3)$$

The flat black plate maximum absolute and percentage errors for upward and downward fluxes are shown in Figure 5a,b,c,d. The absolute and percentage errors in the upward flux are largest for high clouds. These errors are solely due to neglecting geometry, scattering does not have a significant effect on high cloud upward flux. The error peaks at water vapor, reduces the scattering effects $N = 0.3$, and then decreases. This peak is characteristic of flat plate approximations. Note that the error for the tall clouds ($\alpha=2$) may not be important since high clouds tend to have

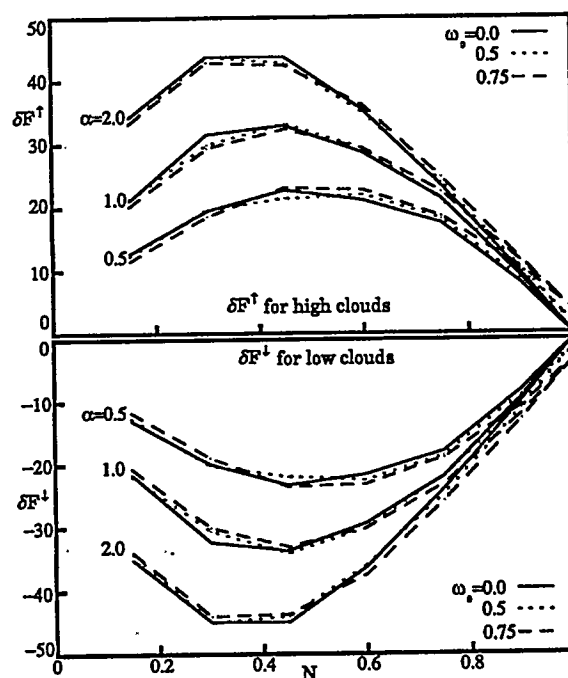


Figure 5a,b. Maximum absolute flux error (W/m^2) for the flat black plate approximation.

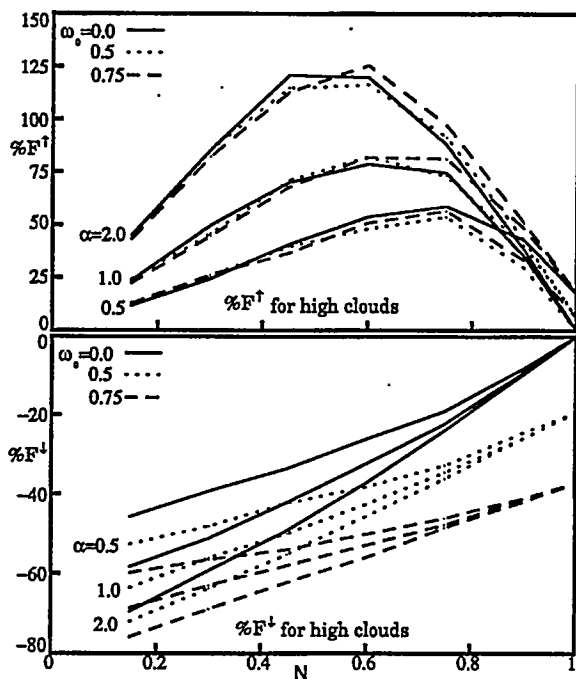


Figure 5c,d. Maximum percentage flux error for the flat black plate approximation.

low aspect ratio. The downward flux absolute error is largest for low clouds and decreases with elevation. The range of the percentage error for the downward fluxes is relatively constant (~80%) for all cloud elevations, high clouds were chosen because they show the most variation with ω_0 .

The geometric black cloud maximum absolute and percentage errors for upward and downward fluxes are shown in Figure 6a,b,c,d. The absolute error in the upward flux is much smaller than the flat black error and has more linear behavior, steadily increasing to a maximum at $N = 1$. The upward flux percentage error has roughly the same range for all cloud heights and is much less than the percentage error from the flat black plate approximation. The maximum downward flux errors are for high clouds; this is where scattering is most important. While the error is large, it is still smaller than the error from the flat black plate approximation.

The results show that scattering can have a significant effect on the fluxes. This is seen in low cloud upward flux and high cloud downward flux. To attain high percentage accuracy, scattering must not be neglected. The addition of absorbing gases, especially water vapor, will reduce the scattering effects presented, with the additional complication that water vapor amounts are dependent on latitude.

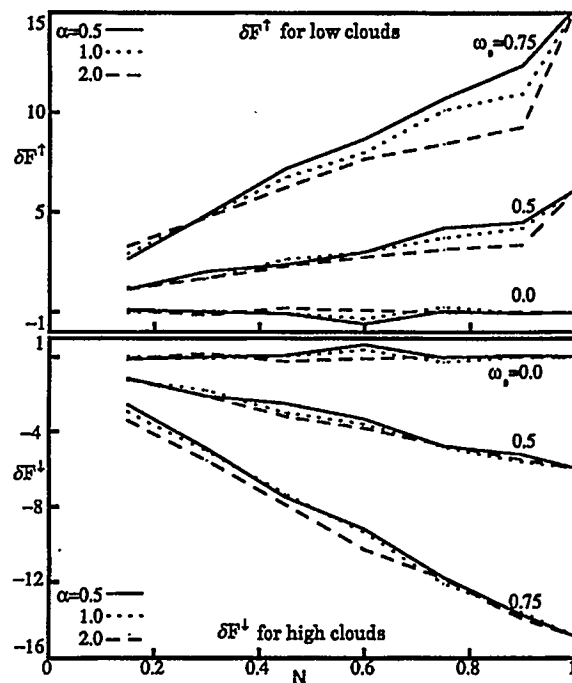


Figure 6a,b. Maximum absolute flux error (W/m^2) for the geometric black cloud approximation.

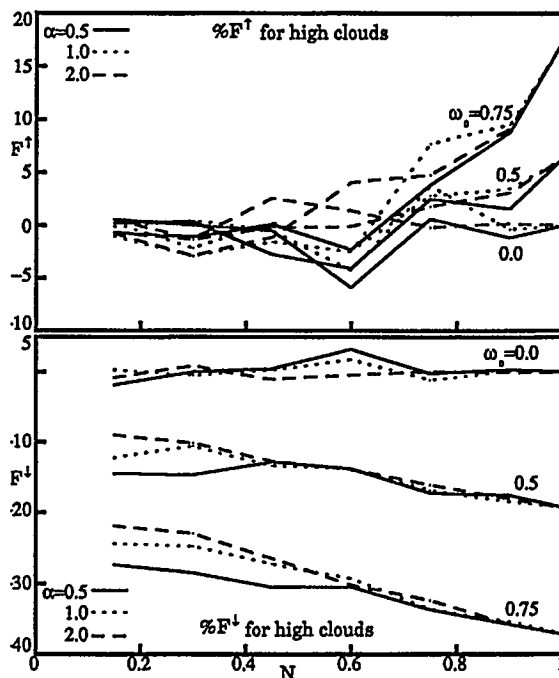


Figure 6c,d. Maximum percentage flux error for the geometric black cloud approximation.

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Comparison of Seasonal and Zonal Patterns of the Direct and Indirect Radiative Forcing of Climate by Aerosols

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Aerosols have a climatically significant influence on the absorption of shortwave radiant energy in the troposphere, both directly through enhanced scattering of sunlight in the absence of clouds and indirectly through their fundamental role in controlling cloud microphysical properties. Quantitative estimates of the magnitudes of both effects are very uncertain (Charlson et al. 1992). In order to possibly distinguish these effects, we calculated the monthly and zonally averaged forcing for two idealized cases representing the direct and indirect forcing for a spatially uniform and seasonally constant perturbation in aerosol and cloud optical properties.

The model atmosphere was divided into two layers: a surface layer with a Lambert albedo of 0.15 and a homogeneous atmospheric layer consisting of molecular Rayleigh scattering and Mie scattering on either sub-micron size particles (direct effect) or cloud droplets (indirect effect). The doubling and adding radiative transfer was calculated at seven wavelengths (from 0.34 nm to 0.99 nm) with the weight of each wavelength interval proportional to the solar flux. The sum of the weights (0.7) is less than 1 to account for the fraction of the solar flux that is absorbed shortward and longward of the above wavelength range (Coakley et al. 1983). The microphysical properties of the aerosol were chosen as effective particle radius $r_e = 0.25 \mu\text{m}$ and variance $v_r = 0.01$ in a modified gamma distribution (Hansen and Travis 1974), refractive index $n_r = 1.4$ [corresponding to $(\text{NH}_4)_2\text{SO}_4$ particles at a relative humidity $\text{RH} = 0.8$; Nemesure et al. 1995]. The direct forcing was determined as the difference in the net downward flux at the top of the atmosphere between a case with an aerosol optical depth of $\tau_a = 0.1$ and one with $\tau_a = 0.05$, corresponding to the background aerosol burden. The hemispheric mean of the direct forcing is -1.1 Wm^{-2} for a 40% cloud free planet.

The cloud layer consists of droplets (refractive index 1.33) with a narrow size distribution having an effective radius

$r_c = 8 \mu\text{m}$ and an optical depth $\tau_c = 6$ for the unperturbed case and $r_c = 7.3 \mu\text{m}$ and $\tau_c = 6.6$ for the anthropogenically perturbed case, corresponding to a cloud droplet number density increase by a factor of 1.3 and a constant liquid water path. The geometric albedo increased by 0.02 from 0.46 for the unperturbed case to 0.48 for the perturbed case. The hemispheric mean of the indirect forcing is -1.8 Wm^{-2} for 20% perturbed cloud cover.

The resulting zonal and seasonal direct and indirect forcings, normalized to a mean of -1 Wm^{-2} , are presented in Figure 1. The seasonal variability and the peak of the direct aerosol forcing are confined to the high latitudes, whereas the indirect forcing extends to much lower latitudes.

The marked difference between Figure 1a and 1b is a direct consequence of the fact that the direct effect operates in the optically thin regime, whereas the cloudy atmosphere of the indirect effect is optically thick. To illustrate this point, Figure 2 shows normalized direct and indirect forcing vs. μ_0 , the $\cos(\text{solar zenith angle})$. Also shown are the "results" of the two extreme optical thickness regimes in the radiative transfer. In the optically thin limit (dashed line marked "direct") the forcing is proportional to the upscatter fraction (the integral of the phase function over the upward hemisphere) because the decrease of downward flux due to the projection of the solar irradiance onto the earth's surface ($\propto \mu_0$) is compensated by the increase in path length through the aerosol layer ($\propto \mu_0^{-1}$). Since the effective path length approaches a finite value as μ_0 goes to 0, the forcing should go to 0 for $\mu_0 = 0$ as indicated by the doubling and adding model result.

The direct and indirect forcing patterns of Figure 1a and 1b may be compared to the seasonal pattern in observed monthly mean surface temperature anomaly trends (Hunter et al. 1993) shown in Figure 1c. This pattern lies between the two patterns exhibited by the direct and indirect forcing

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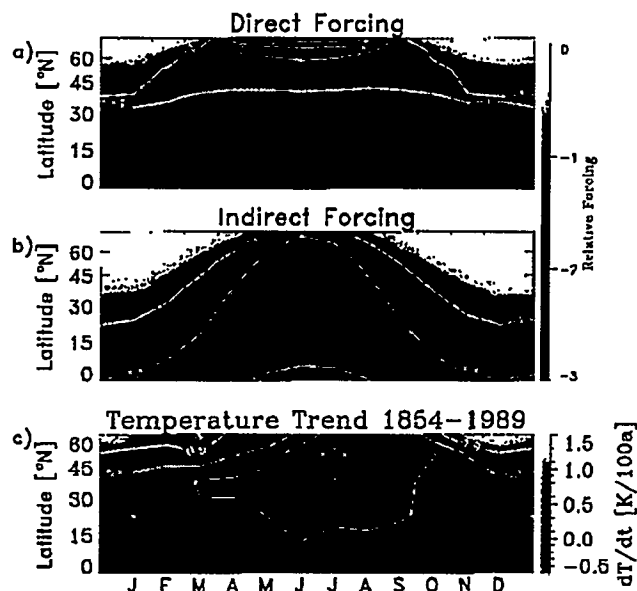


Figure 1. a) Direct aerosol forcing due to a uniform increase of aerosol optical depth over a uniform surface; b) Indirect aerosol forcing due to a uniform increase in cloud optical depth and decrease of cloud droplet radius; c) Zonal average of the monthly mean surface temperature anomaly trends (1854-1989) in K/century.

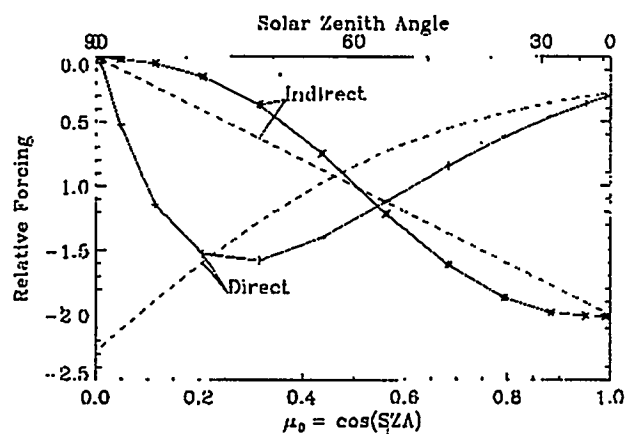


Figure 2. Relative forcing, normalized to a global average of -1, vs $\cos(\text{solar zenith angle})$. Solid lines: Direct (plus symbols) and indirect (asterisks) forcing as computed with the doubling and adding model (see text). Dashed lines: heuristic "models" explaining the main differences between the direct and indirect forcing functional dependence on μ_0 (see text).

and can be matched very well through a linear combination of the two, time lagged by one month. The indirect effect is statistically more significantly correlated to the temperature anomaly trends than the direct effect, indicating that the indirect effect is more important in establishing the seasonal and zonal forcing patterns and that its magnitude is equal to or greater than the direct forcing.

Such a match would only be expected, however, if dynamical effects can be ignored in the climatic surface temperature response to the aerosol forcings on a zonal and monthly mean basis. Quantitative attribution also requires consideration of the seasonal and zonal distributions of low-altitude clouds, cloud free areas, relative humidity, and anthropogenic sulfate aerosol burden and particle size distribution.

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Treatment of Cloud Radiative Effects in General Circulation Models

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We participate in the Atmospheric Radiation Measurement (ARM) program with two objectives: 1) to improve the general circulation model (GCM) cloud/radiation treatment with a focus on cloud vertical overlapping and layer cloud optical properties, and 2) to study the effects of cloud/radiation-climate interaction on GCM climate simulations. This report summarizes the project progress since the Fourth ARM Science Team meeting February 28-March 4, 1994, in Charleston, South Carolina. During this period, four graduate students continue to participate in the research.

Longwave and Solar Parameterization

We evaluate the longwave and solar radiative parameterization of the National Center for Atmospheric Research (NCAR)-Global Environmental and Ecological Simulation of Interactive Systems (GENESIS) (Wang et al. 1995) using concurrent satellite and ground measurements at the Southern Great Plains (SGP) site during the April 1994 Intensive Observation Period (IOP). The comparison results are shown in Figure 1. Note that no aerosols are used for longwave radiation while it is included in solar radiation with visible optical thickness 0.07 and surface albedo 0.16 derived from observations. For longwave radiation, the model calculates larger outgoing flux at the top of the atmosphere (TOA) but smaller downward flux reaching the surface, both suggesting smaller model atmospheric opacity. For solar radiation, the model systematically overestimates the downward flux, again an indication of the smaller model atmospheric opacity, while the agreements for reflected

radiation are much better mainly because it is dominated by the surface albedo. We are currently evaluating the causes for such differences, including the instrumental errors.

High-Level Cloud Parameterization with Interactive Microphysics

A parameterization for solar radiative effect of high-level clouds with interactive microphysics, developed for use in GENESIS, is employed to study the absorption of solar radiation by high clouds. In the parameterization, the cloud particles are assumed to be composed of randomly oriented hexagonal crystals, and the broad band transmittance, reflectance, and absorptance are expressed as a function of single scattering albedo, asymmetry factor, and optical depth which depends on cloud ice water content and effective radius (Ebert and Curry 1992).

The ratio of solar cloud radiative forcing on the surface to that at the top of the atmosphere is used as an indicator for high-cloud absorption. Note that Cess et al. (1995) used observations to derive a value of 1.5 for the ratio, while we calculated 1.3 using the SGP April 1994 IOP data. In the model calculations, the existing GENESIS cloud radiative scheme gives 1.2; this smaller value can be attributed to high-cloud treatment. Indeed, the newly developed high-cloud radiative scheme calculates a value of 1.5, which however, is very sensitive to the effective radius used; for example, the calculated value is 1.2, 1.4, 1.8, and 2.1, respectively, for effective radius 10, 30, 40, and 50 μm .

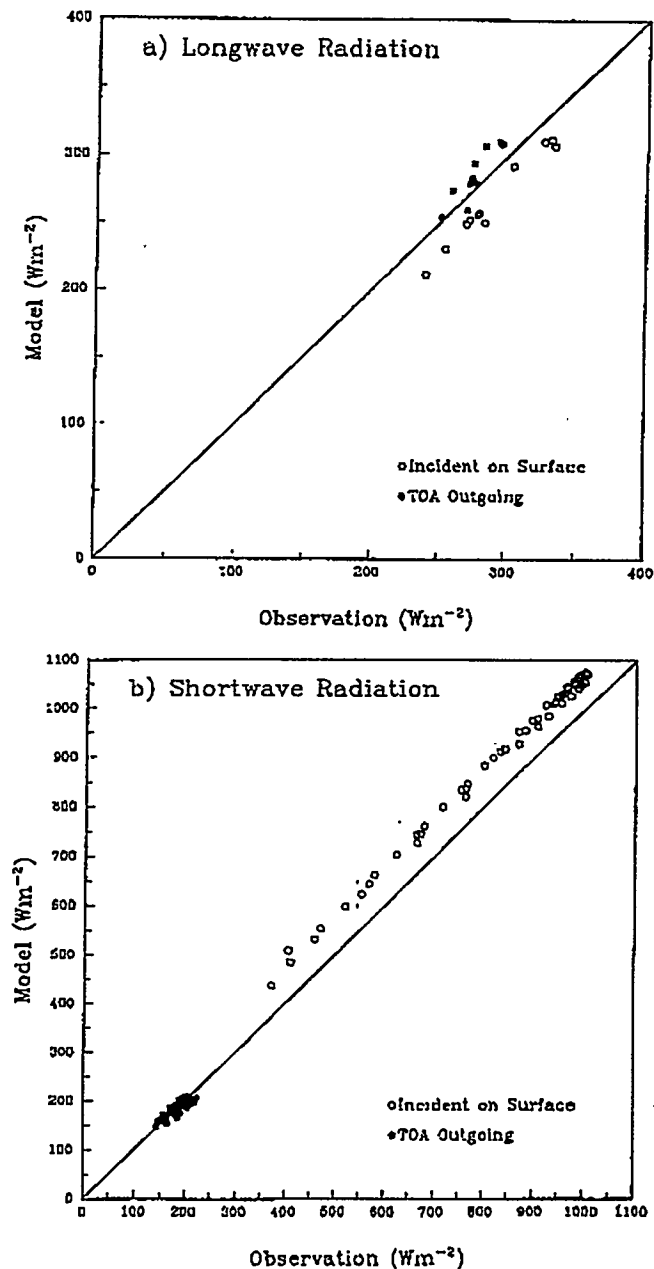


Figure 1. Comparisons between model calculated and observed clear sky of (a) longwave and (b) solar radiative flux (Wm^{-2}) at the SGP central facility during April 1994 IOP (see Figure 2). Open circles are for the downward fluxes reaching the surface while solid circles correspond to values at the top of the atmosphere. The aerosols are only included in the solar radiation calculations with $\tau = 0.07$ and surface albedo 0.16.

Scale-Dependence of GCM Parameterizations for Cloudiness and Radiation

The scale-dependence of cloud-radiation interaction associated with the parameterizations for fractional cloudiness and radiation used in GENESIS is studied by examining the averages, for different spatial scales, of detailed structure of cloudiness and radiation simulated from a regional climate model (ReCM, based on MM5) which incorporates these parameterizations. The regional model simulation is conducted over an area about $(360 \text{ km})^2$ located on the SGP for the period April 10-17, 1994, during which both satellite and surface measurements of radiation fluxes and clouds are available (Figure 2).

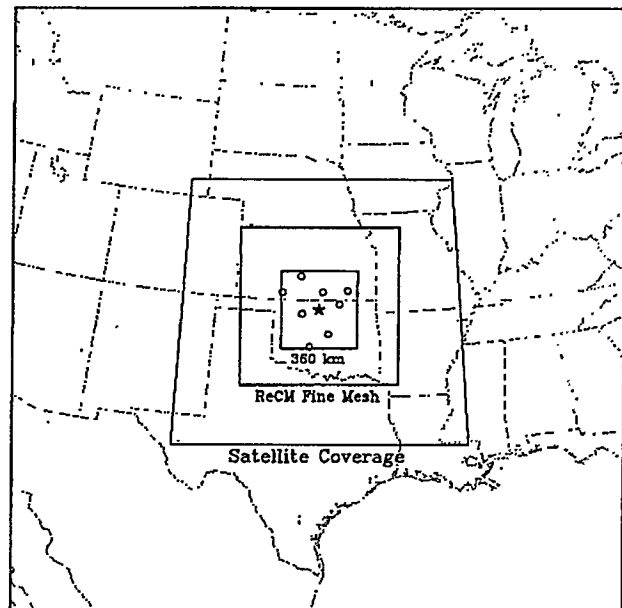


Figure 2. The physical domains for the regional climate model (ReCM) simulations over the SGP ARM site. The ReCM domain has two levels of nesting: a coarse mesh of 45 km and an inner fine mesh of 15 km. The smaller inner square covering part of Oklahoma and Kansas with an area of $(360 \text{ km})^2$ is used for the calculations. The ARM central facility is indicated by a star, while the open circles indicate other locations of the energy balance Bowen ratio (EBBR) measuring stations. The outer box encircles the region where the satellite data are available.

The ReCM simulates well the overall cloud and radiation temporal features when averaged over the entire region. However, specific biases exist in the spatial patterns such as the high clouds, the TOA upwelling solar radiation under cloudy conditions, and the net longwave surface flux under clear conditions at night. The cloud and radiation parameterizations are found to be sensitive to the spatial scale of the computation. The diagnosed total cloudiness shows a strong horizontal resolution dependence which leads to large changes in the surface and TOA radiation budgets. An additional experiment, in which the diagnosed cloud at each level is held constant while the radiation parameterization is recalculated, still produces a substantial sensitivity to spatial scale in the calculated radiation quantities. This is because the nature of the cloud vertical overlapping assumption changes as the horizontal scale of the computation varies.

Small-scale horizontal inhomogeneities have been found to greatly influence the interaction between the cloud and radiation parameterizations used in GCMs and make the results very sensitive to horizontal scale of the computation. This indicates that when the horizontal resolution of a model is changed, particular attention should be paid to the nature of the interaction between clouds and radiation, both in total cloud amount and in the nature of the vertical overlap assumption used by the radiation scheme. When developing and testing physical parameterizations for GCMs, it is important to compare consistent horizontal scales between the observations and the model. High resolution measurements must be averaged to a scale comparable to the model in order to successfully use measurements to develop physically based parameterizations. Details of this study are documented in Dudek et al. (1995).

Stochastic Cloud-Radiation Treatment

As found in Dudek et al (1995), within a typical GCM grid ($\sim 5^\circ\text{C}$ resolution), there exists substantial sub-grid scale variability in cloud radiative forcing (CRF). To incorporate this variability into the cloud-radiation parameterization of a global climate model, we adopt a stochastic approach based on cloud probability distribution function (PDF) which describes the statistical characteristics of cloud variability.

We have examined the observations to define the characteristics of cloud variability. The individual high and total cloud cover PDFs estimated from 0.5°C (sub-cell) satellite measurements at the SGP during the April 1994 IOP (Figure 2) indicate that more than 50% of sub-cells in this domain are either total clear sky or overcast, while partial cloudy conditions occur randomly and much less frequently. We also find that the PDFs are symmetric around 50% cloud cover; for example, the PDF at 70% is quite the same as that for 30% cloud cover. These results suggest that the total and high cloud PDFs change gradually with grid mean cloud cover. These features exist for observed low- and middle-level clouds and are also reproduced (except for high clouds) by the ReCM simulations (Dudek et al. 1995). Consequently, the horizontal variation of vertical cloud distribution, the cloud cover PDF, can be used to parameterize the CRF sub-grid variability.

Given these characteristics, we have designed a new treatment for cloud vertical overlapping, which divides a GCM grid into N sub-cells where column radiation calculations are performed. It can accurately and easily handle the layer cloud optical properties and vertical cloud overlapping because the individual layers within given sub-cells are more likely to be either total clear or overcast condition (binary clouds). The sub-cell cloud covers, determined randomly using observed PDFs, can be treated either as random overlapping or constrained between non-adjacent cloud layers to produce a mixed overlapping. The treatment incorporates the observed sub-grid scale cloud cover variability, and therefore the CRF spatial variability.

Figure 3 gives an example of the effect of cloud overlapping treatment on CRFs. For all calculations, the same vertical profiles of temperature, moisture, ozone corresponding to mid-latitude summer conditions are used; the surface albedo is set at 0.16, a value representative of the SGP, while the solar insolation is fixed at July 15, 1:30 pm SGP local time. These results suggest that given an identical GCM grid mean cloud vertical distribution, surface or TOA CRF can differ by $\pm 30 \text{ Wm}^{-2}$, and the cloud vertical overlapping can affect the cloud absorption by $\pm 10 \text{ Wm}^{-2}$.

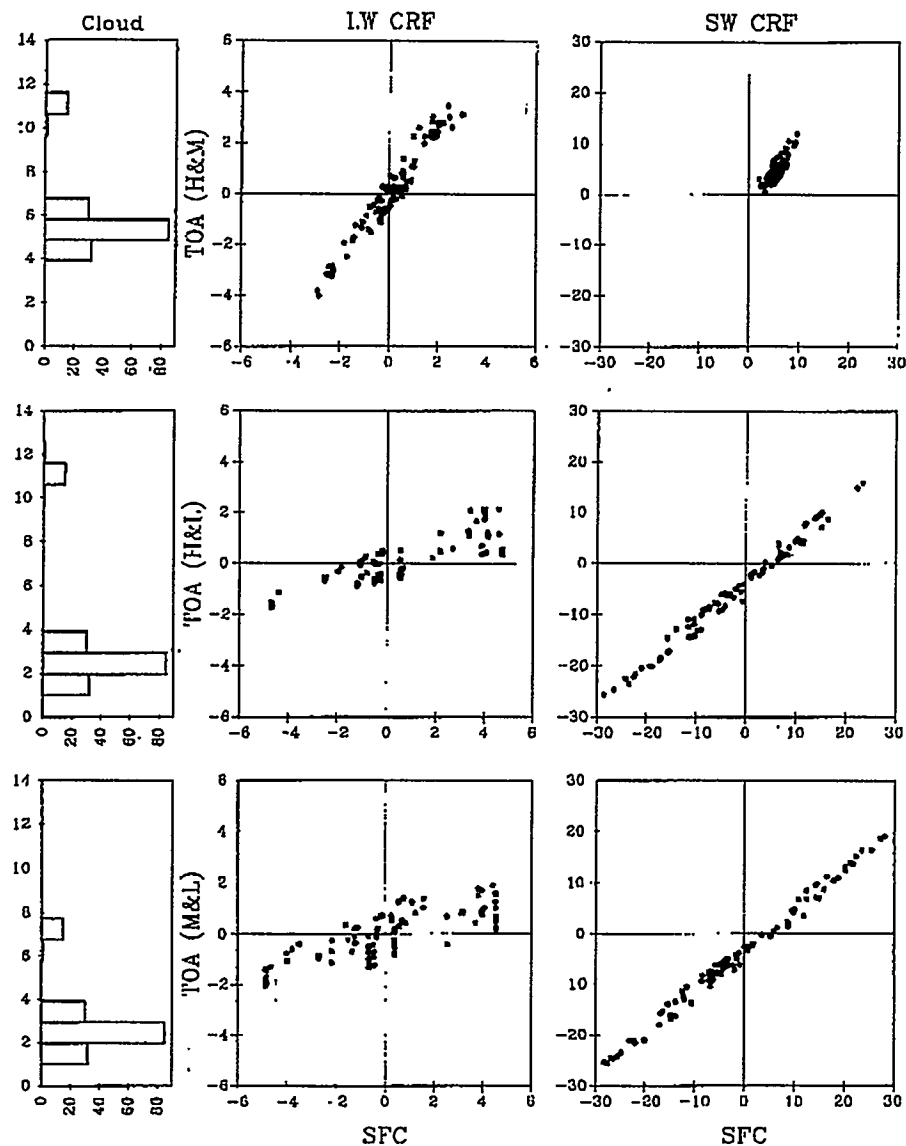


Figure 3. The effect of stochastic treatment of cloud vertical overlapping on the cloud radiative forcing at the top of the atmosphere (TOA) and on the surface (SFC). For each of three idealized cloud vertical distributions (high+middle, high+low, middle+low), 100 sets of full radiation calculations are performed on 12 sub-cells where binary cloud covers occur randomly at individual layers. Each scatter point represents the difference between the mean over the 12 sub-cells and a single grid calculation of the TOA versus surface CRF.

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Clouds, Radiation, and the Diurnal Cycle of Sea Surface Temperature in the Tropical Western Pacific

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Introduction

In the tropical Western Pacific (TWP) Ocean, the clouds and the cloud-radiation feedback can only be understood in the context of air/sea interactions and the ocean mixed layer.

Considerable interest has been shown in attempting to explain why sea surface temperature (SST) rarely rises above 30°C, and gradients of the SST. For the most part, observational studies that address this issue have been conducted using monthly cloud and SST data, and the focus has been on intraseasonal and interannual time scales. For the unstable tropical atmosphere, using monthly averaged data misses a key feedback between clouds and SST that occurs on the cloud-SST coupling time scale, which was estimated to be 3-6 days for the unstable tropical atmosphere (Chu et al. 1990). This time scale is the time needed for a change in cloud properties, due to the change of ocean surface evaporation caused by SST variation, to feed back to the SST through its effect on the surface heat flux.

To clarify the SST, we must distinguish between the "skin" SST, which is the radiometric temperature of the sea surface, and the "bulk" SST. The true bulk SST is defined to be the temperature within the upper few centimeters of the ocean surface. The bulk SST determined from buoy measurements is typically obtained at a depth of 0.5 m, while the bulk SST determined from ship measurements may be obtained from depths as large as 5 m. The radiative, latent, and sensible heat exchanges between the atmospheric and oceanic boundary layers depend on the actual skin temperature of the ocean, making the skin temperature the critical SST for examining air/sea interactions. The skin temperature can differ from the bulk water temperature in the tropics by values as large as 1-3 K (e.g., Cechet 1993), although the magnitude of the bulk-skin temperature difference is usually < 1 K (e.g., Wick et al. 1992).

Table 1 shows the changes in surface heat flux components associated with a 1°C change in SST for average conditions during the Tropical Ocean Global Atmosphere (TOGA)

Table 1. Changes in surface heat flux components associated with a 1°C change in SST for average conditions during the TOGA COARE IOP.

Component	Flux Change	
	(Wm ⁻²)	(%)
Upwelling Longwave	6.3	1.3
Sensible Heat	2.4	23.3
Latent Heat	18.7	16.2

Coupled Ocean Atmospheric Response Experiment (COARE) Intensive Observation Period (IOP). Substantial changes are seen particularly in the surface latent heat flux. All of the changes are of the same sign (i.e., none of the changes cancel if the net surface heat flux is being evaluated). Therefore, a 1°C change (or error) in sea surface skin temperature would result in a change (or error) of 27 W m⁻² in the net surface heat flux. In many instances, an error of this magnitude would be large enough to change even the sign of the net surface heat flux and could significantly modulate atmospheric boundary layer and convective processes. Therefore, care must be taken in using the correct skin SST to evaluate the surface fluxes and in determining the diurnal variation of SST in the TWP.

In this paper, we use data obtained from the TOGA COARE IOP along with an ocean mixed layer model to address the relationship between clouds, surface radiation flux, and SST of the TWP Ocean over the diurnal cycle. A blending of the model and observations are used for this investigation as this analysis would not be possible using only the observational data. By examining the relationship between clouds, surface fluxes, and SST over the diurnal cycle, we can interpret the details of the direct influence of clouds on SST. We examine the effects not only of the surface heat fluxes, but the fresh water and momentum fluxes as well. We also interpret these relationships in the context of the temperature variations at various depths in the ocean that are commonly used

to describe "surface" temperature. Sensitivity studies are conducted with the model to investigate the roles of surface wind speed, precipitation, ocean turbidity, and ocean initial state in modulating the radiation-induced diurnal cycle in SST.

TOGA COARE Data

The TOGA COARE IOP occurred from November 1992 through February 1993 in the Western Pacific region bordered by 10°N, 10°S, 140°E, and 180°W. Oceanographic and meteorological data were gathered from ships, buoys, aircraft, and satellites.

In this study, we use the measurements obtained from the Research Vessel *Moana Wave*, which provided a co-located dataset of surface fluxes and ocean microstructure measurements. Further details of these observations are described by Webster et al. (1995).

A summary of the surface meteorology, surface fluxes, and ocean mixed layer depth from the Research Vessel *Moana Wave* is given in Table 2, with the maximum and minimum values presented for hourly averages.

Model Description

The 1-D ocean mixed layer model that we use here used a second-order turbulence closure scheme described by Kantha and Clayson (1994). Penetration of shortwave radiation into the upper ocean is modelled following Morel and Antoine (1994). A parameterization for skin SST has been added to the Kantha-Clayson model, following Wick (1995).

Table 2. Average values of surface parameters and fluxes measured by the R/V *Moana Wave* during the TOGA COARE IOP (surface flux is positive into the ocean).

Parameter	Average	Min	Max
SST, 1 cm (°C)	29.1	25.7	31.5
Wind speed, surface (m s ⁻¹)	4.9	0.1	15.9
Peak insolation (Wm ⁻²)	732.9	180.0	958.0
Longwave, down (Wm ⁻²)	413.0	370.0	441.0
Latent heat flux (Wm ⁻²)	-115.6	-7.0	-347.0
Sensible heat flux (Wm ⁻²)	-10.3	8.0	-69.0
Precipitation (mm hr ⁻¹)	0.45	0	37.1

The surface fluxes of radiation, sensible and latent heat, fresh water, and momentum are specified using hourly values measured on the *Moana Wave*. Also included are additional surface fluxes associated with precipitation. The surface momentum flux associated with rain is included following Kantha and Clayson (1994). The surface sensible heat flux associated with rain is parameterized following Gosnell et al. (1995).

The performance of the model during TOGA COARE is discussed in Webster et al. (1995). After an initial spin up period of 3 days, the modeled 1-cm temperature is within 0.1°C of the observed value, the model reproducing much of the fine structure in the thermal and salinity profiles of the upper ocean.

Diurnal Cycle

The joint time series of SST, ocean mixed-layer characteristics, surface heat flux components, surface wind speed, and rainfall rate are now examined using a blend of observations and model calculations. In this analysis, we consider data and model calculations for the fourth day of a four-day model integration.

Table 3 presents a summary of seven cases representing a range of conditions encountered, for which we could obtain a reasonably homogeneous four-day time series (as determined by the surface fluxes) and good comparison of the model with observations by the third or fourth day of model integration. The ocean mixed layer depth was determined from the modeled density field. The difference between the skin and true bulk SST is seen to range between 0.3 and 0.43°C, with the skin always cooler than the true bulk SST. Differences between the daily average skin SST and the buoy (0.5 m) and ship (4.5 m) bulk SST range from 0.3 to 0.6°C. Nearly clear-sky conditions were present for cases 1, 4, 5; isolated clouds were present for cases 3 and 6; and cases 2 and 7 represent disturbed conditions. Table 3 shows that there is no strong correlation between the daily peak insolation and average skin SST. The diurnal temperature range was determined by the skin SST difference between the predawn minimum and the maximum value that occurs shortly after the peak insolation. The negative values of diurnal temperature range shown in Table 3 for cases 2 and 7 reflect the general absence of a diurnal cycle and an overall cooling of the SST. The highest values of skin SST and greatest diurnal range in skin SST occur for cases 1 and 5. These two cases have the lowest windspeeds and the shallowest mixed layer depths.

Table 3. Summary of case study statistics.

Case	Date	Peak Insolation (Wm^{-2})	Wind Speed (Ms^{-1})	Total Precip (mm)	Average Skin SST ($^{\circ}\text{C}$)	Diurnal SST Range ($^{\circ}\text{C}$)	Average 1cm SST ($^{\circ}\text{C}$)	Average 0.5m SST ($^{\circ}\text{C}$)	Average 4.5m SST ($^{\circ}\text{C}$)	OML Depth (m)
1	11/16-11/17	929	2.7	3.3	29.2	1.31	29.6	29.8	29.6	3.8
2	12/23-12/24	368	6.0	6.9	28.7	-0.14	29/1	29.1	29.1	24.3
3	12/26-12/27	629	5.5	70.5	28.5	0.30	28.8	28.8	28.9	41.5
4	12/31-1/1	857	9.5	2.2	28.3	0.10	28.6	28.6	28.6	35.2
5	1/9-1/10	910	1.8	0.2	28.8	1.50	29.2	29.3	20.1	4.7
6	2/3-2/4	852	5.3	10.3	28.7	0.48	29.1	29.1	29.1	28.4
7	2/9-2/10	557	4.9	51.4	28.7	-0.09	29.0	29.0	29.0	23.9

The shallow mixed layer depth increases the response of the mixed-layer temperature to surface heating. Cases 3 and 4 occur just after a westerly wind burst and are accompanied by high wind speeds and a deepened ocean mixed layer, resulting in a very low diurnal range in skin SST, even though the peak insolation for case 4 is fairly high. Although there is only a few tenths of a degree difference between the different determinations of the daily mean "surface" temperature, there are large differences in the amplitude of the diurnal surface temperature range, from 0.36°C at a depth of 4.5 m to 1.22°C at a depth of 1 cm. The peak heating at a depth of 4.5 m occurs approximately 8 hours after the maximum value of skin SST. At the time of peak insolation, the mixed layer is above 4.5 m. The daytime heating is only reflected at depths greater than 4 m following the nocturnal deepening of the mixed layer.

A comparison of this case with case 1 illustrates the importance of wind speed and ocean mixed layer depth on sea surface temperature under clear sky conditions. The deeper ocean mixed layer and increased evaporative cooling result in lower surface temperatures, a reduced diurnal amplitude in SST, and a reduced phase lag of temperature with depth.

Sensitivity Studies

Simple correlation between variables is not sufficient to reveal causality. This is especially true for a complicated, nonlinear system such as the upper ocean. Therefore, diagnostic studies cannot unambiguously identify feedbacks. However, carefully designed model experiments can be used to quantify the relative importance of different physical processes and to isolate causal mechanisms, provided that the models accurately simulate the relevant

climate processes and sensitivities. The Kantha and Clayson (1994) ocean mixed layer (OML) model is now used to explore some of the characteristics of the relationships between solar radiation, wind speed, precipitation, and oceanic state in influencing the diurnal SST variability. The model is initialized with average ocean conditions for the TOGA COARE period as listed in Table 2. The model is integrated for periods of four days. The results presented in Figure 1 are for the diurnal cycle at 1 cm of the last day of the integration.

Solar radiation is responsible for the diurnal cycle in SST and surface solar radiation flux is primarily modulated by cloud characteristics. The amplitude of the diurnal cycle is largest for the largest insolation, the amplitude of the

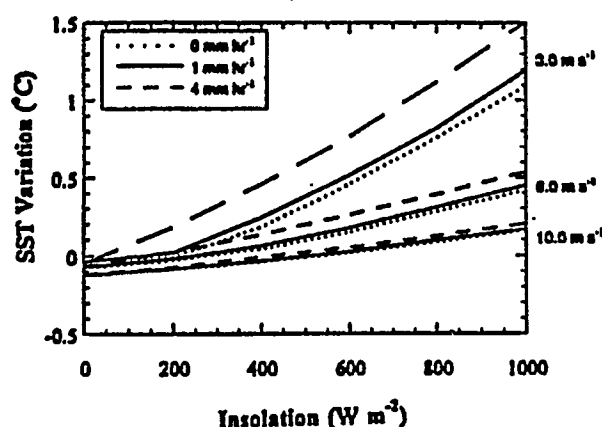


Figure 1. The amplitude of the diurnal variation of the SST as a function of peak insolation for different values of wind speed and rainfall following the rainfall. The coldest SST is for the night rain and the warmest SST is for the a.m. rain.

diurnal cycle being virtually zero for peak insolation values less than 300 Wm^{-2} (Figure 1). A phase lag exists between the time of the peak insolation and the SST maximum.

Surface wind speed modulates the amplitude of the diurnal cycle by influencing the surface heat flux (through the sensible and latent heat flux) and by determining the momentum flux. In general, wind speed is negatively correlated with SST, with increased wind speeds decreasing the net surface heat flux and by increasing the upper ocean mixing, thus distributing the surface heating throughout a greater depth of the ocean and diminishing the surface warming. Figure 1 shows that the diurnal amplitude of the surface temperature decreases nonlinearly with increasing wind speed. This nonlinearity is explained by the work being done on the mixed layer, is proportional to the cube of the surface wind speed (Webster and Lukas 1992), and depends on whether the presence of buoyancy barriers at different depths are overcome by the mixing.

The effects of precipitation on SST are very complex, since precipitation influences the surface fluxes of heat, freshwater, and momentum. Precipitation results in a decrease in surface salinity, cooling of the surface, and increased mixing. Figure 1 shows that the magnitude of the diurnal SST amplitude is not influenced by rainfall rates less than 0.5 mm hr^{-1} ; small amounts of fresh water are rapidly mixed downward. At larger rainfall rates, the diurnal amplitude of SST increases as the formation of a fresh and stable layer at the surface allows greater surface heating. For higher wind speeds, the diurnal amplitudes are much smaller and much less influenced by precipitation. Furthermore, the impact of precipitation may depend on when the precipitation occurs in the diurnal cycle. At 5 a.m. of the day following the rainfall, the coldest SST is for the night rain, and the warmest SST is for the a.m. rain.

A number of experiments were run to assess the sensitivity of the diurnal amplitude in SST to the initial profile of temperature, salinity and chlorophyll in the ocean, and the depth of the mixed layer. An evolving ocean mixed layer will certainly have an influence on the SST, and a deepening OML will have a cooling effect on SST as cold water from below is entrained into the mixed layer. However, whereas the effect of the ocean initial conditions on the absolute SST is important, the effect on the diurnal amplitude of the skin SST is relatively small. Over the range observed during the TOGA COARE IOP (Siegel et al. 1995), increased chlorophyll concentrations increase the diurnal amplitude of SST only slightly. Absorption by

chlorophyll occurs only in the visible part of the spectrum, and the visible radiation penetrates into the ocean, thus influencing surface temperature relatively little.

To summarize the results shown in Figure 1, we find that the magnitude of the diurnal variation of the SST increases with increasing insolation, decreases with increasing wind speed, and increases with increasing precipitation. For peak insolation values of less than 200 Wm^{-2} , the diurnal amplitude is near zero or slightly less than zero for all values of wind speed and precipitation that were considered. Increasing wind speed diminishes the impact of precipitation on the diurnal SST amplitude.

Discussion

The discussion presented here has focused on sea surface temperatures over a range of 3°C , with a diurnal amplitude of 1.5°C or less. Are variations of SST on the order of 1°C important in the TWP? To put these temperature variations into context, the difference in the TWP SST from El Niño to La Niña is only about 1°C . The importance of surface temperature variations (or errors) of 1°C at these high surface temperatures arises from the exponential relationship between saturation vapor pressure and temperature. As pointed out in Table 1, a 1°C error in SST in the TWP will result, on average, in an error of 27 Wm^{-2} to the surface energy balance. The effect of such an error on the ocean surface buoyancy flux is further enhanced by the associated error to the evaporation component to the fresh water flux, which acts in the same direction as the errors to the surface heat balance.

Correct observation and simulation of the diurnal cycle of SST is undoubtedly essential to understanding and modeling the correct feedback between clouds and the SST. Because of the nonlinear relationships between clouds and the surface fluxes, incorrect simulation of the interactions between clouds and the surface on the diurnal cycle may result in errors to the simulation of larger-scale cloud systems and feedback to the large-scale atmospheric dynamics.

Current SST climatologies are effectively climatologies of the upper ocean at a depth of 5 m (corresponding to the level at which ships routinely measure "surface" temperature). Temperatures at 5 m have been shown to differ substantially from the skin SST, particularly under conditions of low winds and strong insolation. The data on the diurnal amplitude of skin SST presented here can provide the foundation for adding a diurnal cycle to

satellite-derived skin SST values, by including ancillary information regarding surface insolation, wind speed, and precipitation that is also available from satellite (e.g., Curry et al. 1993).

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Observations of Regional and Local Variability in the Optical Properties of Maritime Clouds

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Introduction

White and Fairall (1995) calculated the optical properties of the marine boundary layer (MBL) clouds observed during the Atlantic Stratocumulus Transition Experiment (ASTEX) and compared their results with the results obtained by Fairall et al. (1990) for the MBL clouds observed during the First International Satellite Climatology Program (ISSCP) Regional Experiment (FIRE). They found a factor of two difference in the optical depth versus liquid water relationship that applies to the clouds observed in each case. In the present study, we present evidence to support this difference. We also investigate the local variability exhibited in the ASTEX optical properties using measurements of the boundary layer aerosol concentration.

Optical Properties Algorithm

We used solar irradiance measurements from a pyranometer to calculate the transmission coefficient, T_r , as a function of solar zenith angle, θ . We then used the plane-parallel radiative transfer parameterization of Stephens et al. (1984) to obtain values of albedo, R_e , and optical depth, τ . Details are given by Fairall et al. (1990). A laser ceilometer provided measurements of the overhead cloud fraction, f . Microwave radiometric data were used to calculate the liquid water path (LWP).

The plane-parallel assumption is reasonable for the FIRE transmission data since $f \approx 1$ was observed much of the time. However, the bulk of the data collected during ASTEX pertains to broken clouds. To complicate matters further, the values of f and LWP were measured using upward-looking, narrow field-of-view instruments (i.e., the ceilometer and radiometer), whereas the values of T_r were

based on pyranometer measurements that are hemispheric. Therefore, we modified the ASTEX transmission data using the algorithm of Chertock et al. (1993) [their Equations (7) and (8)]. The algorithm accounts for cloud fraction (i.e., $f \neq 1$) and the zenith angle dependence of cloud fraction, $f(\theta)$.

We investigated the relationship between optical depth and liquid water because this relationship is often used to parameterize the radiative effects of clouds in large-scale models. For example, Stephens (1978) has shown that $\tau = (3/2)LWP/(\rho_w r_e)$, where r_e is the effective radius and ρ_w is the water density. Alternatively, Fairall et al. (1990) have shown that for a lognormal droplet size distribution and a linear profile of liquid water $\tau \propto N^{1/3}$, where N is the cloud droplet number concentration.

Results

Figure 1 shows τ versus LWP for ASTEX and FIRE. We used diurnal averages for the ASTEX data to compensate for the sampling inadequacies mentioned above. The lower bounds used for the plot are consistent with the approximate lower limits on the accuracy of our values for τ and LWP. White and Fairall (1995) placed more confidence in the ASTEX optical properties obtained from measurements taken near solar noon because then the instruments are matched to the upward looking case. If we consider only these points in Figure 1, we can place a line through the ASTEX data that lies roughly a factor of two beneath the FIRE line.

A possible explanation for this difference would be that r_e and N were different for the two cases. For example, FIRE took place much closer to a densely populated coastline than ASTEX. Because continental aerosols are generally smaller than maritime aerosols, we might expect FIRE to have a

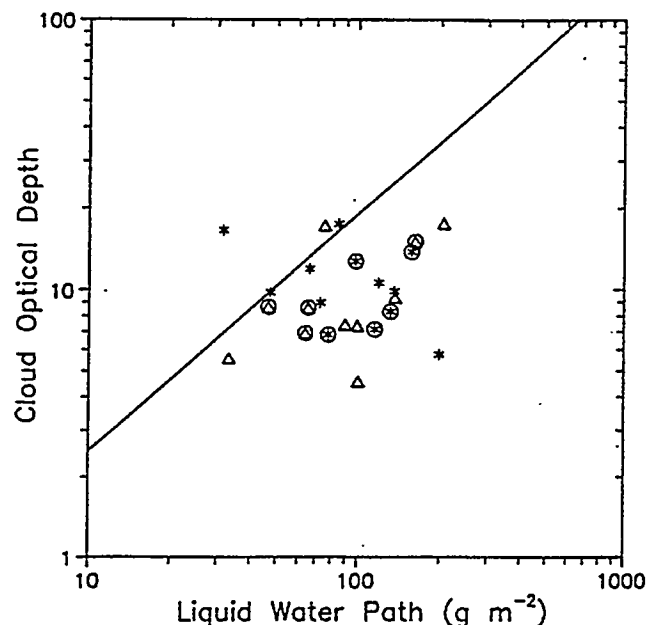


Figure 1. Cloud optical depth versus liquid water path. The symbols plot diurnal averages derived from ASTEX ship data (asterisks) and ASTEX island data (triangles). The line shows the relationship for FIRE (Fairall et al. 1990). The circles classify data derived from measurements taken near solar noon ($\theta \leq 15^\circ$).

higher mean concentration of the smaller aerosols, which would lead to a smaller mean r_e and a larger optical depth for a given LWP. Satellite observations do indicate a higher cloud albedo near the California coast (Minnis et al. 1992), consistent with greater concentration of cloud condensation nuclei (CCN) associated with continental aerosol sources.

There is additional evidence suggesting the mean r_e was different for FIRE and ASTEX. Han et al. (1995) have computed cloud droplet sizes for FIRE using the ISSCP data analysis. Their results show $r_e = 8 \mu\text{m}$ for FIRE. This is an estimate of the r_e near cloud top. Frisch et al. (1995) show that profiles of modal radius, r_m , can be retrieved from vertical profiles of radar reflectivity and microwave radiometer measurements of the LWP by assuming a cloud droplet distribution shape and characteristic width. They retrieved $r_m = 8\text{--}12 \mu\text{m}$ near cloud top during ASTEX. By taking the ratio of the third to the second moment of their assumed lognormal distribution, it can be shown that $r_e = 1.36 r_m$. The observed values of r_m then translate into values of r_e ranging from $11\text{--}16 \mu\text{m}$, which is consistent with a factor-of-two difference in optical depth.

We investigated the impact of aerosols on the optical depth of the ASTEX clouds using shipboard measurements of condensation nuclei (CN). Here, we must make the assumption that the particle concentration measured at the surface is directly proportional to the number of activated CCN. Figure 2 shows 10-min values of ASTEX optical depth data plotted for two ranges of aerosol concentration. Although there is a significant amount of scatter, most of the ASTEX data with "low" CN concentrations (i.e., less than 200 cm^{-3}) lie below the FIRE line, while much of the ASTEX data with "high" CN concentrations (i.e., greater than 800 cm^{-3}) lie above the FIRE line. The low CN concentration data have an average concentration of 131 cm^{-3} and on average lie a factor of 0.7 below the FIRE line. The high CN concentration data have an average concentration of 1128 cm^{-3} and, on average, lie a factor of 1.4 above the FIRE line. If we assume $\tau \propto N^{1/3}$, then the average ratio of aerosol concentrations would change the optical depth by a factor of $(1128/131)^{1/3} = 2.0$. This value agrees with the observed change in optical depth (i.e., $1.4/0.7 = 2.0$).

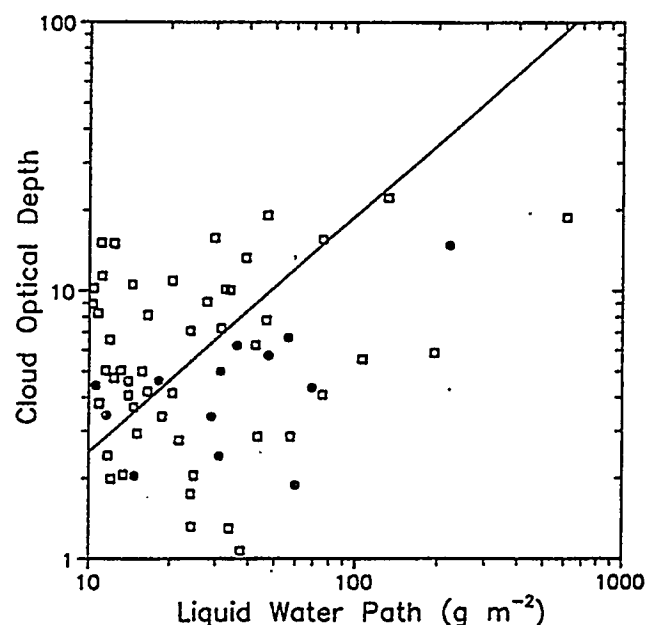


Figure 2. Cloud optical depth versus liquid water path. The ASTEX 10-min, ship-based data are plotted for instances of condensation nuclei concentration greater than 800 cm^{-3} (squares) and less than 200 cm^{-3} (solid circles). The line is for the FIRE relationship (Fairall et al. 1990).

Summary

Using cloud optical properties deduced from surface-based measurements of the transmission coefficient and a radiative transfer model, we investigated the relationship between optical depth and liquid water for ASTEX and compared our result with the result obtained for FIRE. Our comparison indicated the ASTEX clouds formed in significantly cleaner air masses on average than the FIRE clouds. Independent droplet-size information deduced from radar and satellite observations also suggest that the ASTEX cloud droplets were on average a factor-of-two larger than the FIRE cloud droplets. An investigation into the impact of boundary layer aerosol concentration on cloud optical depth implied a factor-of-two change in the τ versus LWP relationship associated with a factor-of-eight variation in CN concentration observed just within the ASTEX data set, a direct demonstration of the so-called Twomey effect.

Acknowledgments

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Solar Semidiurnal Tidal Wind Oscillations Above the CART Site

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Harmonic analysis of wintertime data from 915- and 404-MHz radar wind profilers at four sites in North America has identified coherent semidiurnal wind oscillations through the entire depth of the troposphere. These winds are readily apparent above the Cloud and Radiation Testbed (CART) site, as evidenced from analyses of data from the HaviLand, Kansas, radar profiler. The characteristics of this wind system match the characteristics of solar semidiurnal atmospheric tides, as predicted by a simple dynamic model.

Characteristics of the Observed Wind System

Our recent analyses of radar profiler wind data (Whiteman and Bian 1994, 1995a, 1995b) have identified semidiurnal wind oscillations in the troposphere that are consistent from site to site at four sites that are widely separated across the United States (Table 1 and Figure 1). For these analyses, the series of hourly wind data at each range gate and site were averaged vectorially for each hour of the day to

produce a mean daily time-height cross section of vector winds. A harmonic analysis was then performed for the u- and v-components of the 24 vectors in the cross section for each range gate. The semidiurnal wind amplitudes and phases obtained from this analysis were consistent from site to site, despite widely differing geography and topography at the four sites. The coherent semidiurnal wind system first became apparent at altitudes above a 500- to 1000-m-deep frictional boundary layer. Tidal amplitudes were ~0.2 m/s there, but increased with elevation to attain 0.7 m/s at 5-7 km above mean sea level (MSL). The phases were consistent from site to site when calculated in local mean solar time and changed little with altitude at the four sites investigated, with the maximum northward (eastward) wind speeds attained at about 0100 (0400) apparent local solar time. The winds at a fixed height turned clockwise with time. This wind system behavior is shown in a hodograph representation in Figure 2. The characteristics of this semidiurnal circulation were compared with the results of a dynamic model used to simulate the characteristics of solar semidiurnal thermal atmospheric tides.

Table 1. Information on radar profiler sites and data sets.

Site	ID	Latitude (°N)	Longitude (°W)	Elevation (m MSL)	Period of Record
NOAA Wind Profiler Demonstration Network (404 MHz)					
Hudson, Massachusetts	HUD	42.41	71.48	93	JFD 91 and JF 92
HaviLand, Kansas	HAV	37.65	99.09	648	JFD 91 and JF 92
Platteville, Colorado	PLT	40.18	104.72	1524	JFD 91 and JF 92
Project MOHAVE (915 MHz)					
Tempe Bar, Arizona	TBR	36.02	114.33	460	01/03/92-04/19/92

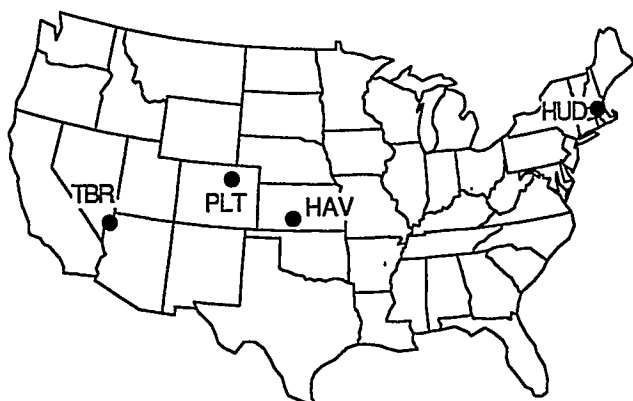


Figure 1. Radar profiler locations.

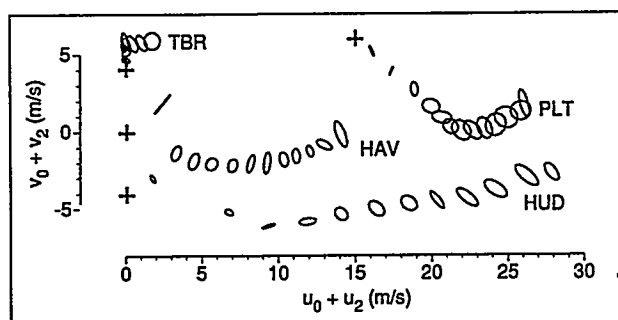


Figure 2. Hodograms of the sum of the mean daily (subscript 0) and mean hourly semidiurnal (subscript 2) wind components at the four sites. Twenty-four hourly points make up the elliptical wind figures for each range gate. Each of the hourly points can be thought of as defining the tip of a wind vector drawn from the coordinate origin, indicated for each site by a plus sign. This vector rotates clockwise twice per day around each of the hodograms. For use, the u_1 and v_1 -component scales must be shifted to the origin for each of the sites. The first range gates are at 2024, 593, 637 and 1148 m MSL for HAV, HUD, PLT and TBR, respectively. Additional plotted range gates are at 500-m intervals for HAV, HUD and PLT, and at 406-m intervals for TBR, so that the highest plotted range gates are at 8024, 6093, 3885, and 6648 m MSL, respectively.

Dynamical Model

Semidiurnal tidal wind characteristics are illustrated by solving the momentum equations for a frictionless atmosphere on a rotating earth:

$$\begin{aligned} \frac{\partial u}{\partial t} - 2 \Omega v \sin \Phi &= -\frac{1}{\rho R \cos \Phi} \frac{\partial p}{\partial \lambda} \\ \frac{\partial v}{\partial t} + 2 \Omega u \sin \Phi &= -\frac{1}{\rho R} \frac{\partial p}{\partial \Phi} \end{aligned} \quad (1)$$

where u and v denote, respectively, the eastward- and northward-directed semidiurnal wind components, Φ is latitude, λ is longitude, Ω is the angular velocity of the earth, ρ is air density, R is the earth's radius, and p is the pressure perturbation obtained from Haurwitz's (1956) global fit to observed semidiurnal surface pressure perturbation data

$$p = p_m \cos^3 \Phi \sin[2(t' + \lambda) + \sigma] \quad (2)$$

where p_m is the amplitude of the semidiurnal pressure oscillation at the equator (1.16 hPa), t' is Universal Time (UT) reckoned in angle at the rate of 360° per mean solar day from lower transit (i.e., midnight), and σ is the initial phase of the pressure perturbation (158°). The solutions to Equation 2 for the semidiurnal winds, from Chapman and Lindzen (1970), are

$$u = C_s (1 + 1.5 \sin^2 \Phi) \sin(2t'' + 180^\circ + \sigma) \quad (3)$$

$$v = 2.5 C_s \sin \Phi \sin(2t'' + 90^\circ + \sigma) \quad (4)$$

where $C_s = p_m / \rho R = 0.2$ m/s, and $t'' = t' + \lambda$ is apparent local solar time. The maximum amplitudes of the u and v components at the mean latitude of the four sites is 0.3 m/s, and the maxima in the u and v components occur at 0344 (and 1544) and 0044 (and 1244) apparent local solar time, respectively. Thus, the modeled wind system characteristics match those of the observations. The modeled winds occur over the entire globe, turn clockwise with time at a given height, and have amplitudes and phases similar to those obtained from the radar profiler analyses.

Conclusion

Since the winds observed with radar profilers have characteristics that match those of the solar thermal tides, as predicted by a simple dynamical model using observed surface semidiurnal pressure perturbations, we conclude that the observed winds are solar semidiurnal tides. This global tidal wind system is apparent in data collected near the CART site. Further investigations are under way to determine whether tidal convergence fields influence cloudiness and precipitation in the CART region.

Acknowledgments

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A Characteristic Scale in Radiation Fields of Fractal Clouds

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The wavenumber spectrum of Landsat imagery for marine stratocumulus clouds shows a scale break when plotted on a double log plot; the $\approx -5/3$ slope is interrupted at a scale around 200 m and followed by a much steeper slope (Cahalan and Snider 1989). We offer an explanation of this scale break in terms of a smoothing by horizontal radiative fluxes (Davis et al. 1995), which is parameterized and incorporated into an improved independent pixel approximation (Marshak et al. 1995).

We compute the radiation fields emerging from cloud models with horizontally variable optical depth fields given by fractal models developed by Schertzer and Lovejoy (1987), Cahalan et al. (1994), and Marshak et al. (1994). These models were tuned to have spectral exponents around $5/3$, as observed by Cahalan and Snider (1989) using ground-based microwave radiometry. Two approaches were used: 1) the independent pixel approximation (IPA) which treats the radiation properties of each pixel as a 1-D plane-parallel layer, ignoring any net horizontal transport; and 2) forward Monte Carlo (MC) simulation with variance reduction. For marine stratocumulus, IPA and MC have close domain-average albedoes (Cahalan et al. 1994), but there are dramatic differences in individual pixel values. Figure 1 illustrates the random optical depth field and the two albedo fields calculated by IPA and MC methods.

We use comparative spectral and multifractal analyses to qualify the validity of the IPA at the largest scales and demonstrate its shortcomings on the smallest scales. We show that there is a characteristic scale η_r which we call the "radiative smoothing scale." For scales smaller than η_r , the IPA follows the fluctuations of the optical depth field whereas the MC albedo has a much smoother behavior with a spectral exponent around 3 (Figure 2). For scales larger than η_r , IPA and MC albedoes and nadir radiances have similar spectral properties which resemble those of the optical depth field. Multifractal properties differ only for the higher moments (Davis et al. 1994) implying that, the bigger the jumps in optical depth, the more powerful the radiative smoothing.

The location η_r of the scale break is determined entirely by the characteristic size $\langle \xi \rangle$ of the spot resulting from localized illumination. The size of this spot is shown to be proportional to the harmonic mean of the product of the transport mean free path and the geometrical cloud thickness, h , in the frame of diffusion theory for homogeneous media. Equivalently,

$$\langle \xi \rangle \approx \frac{h}{\sqrt{(1-g)\tau}} \quad (1)$$

where τ is the cloud optical thickness and g is the asymmetry factor. Numerical calculations show that Equation (1) remains true for fractal cloud models with τ replaced by its average value $\langle \tau \rangle$. Typical values for marine Sc are $h = 0.3$ km, $\langle \tau \rangle = 13$, and $g = 0.85$, so that $\langle \xi \rangle \approx 0.2$ km.

If the shape and size of the spot are known, then the effect of the horizontal photon transport can be estimated. More precisely, IPA fields can be corrected by convolution with the probability density function describing the spot. The shape of the spot, meaning the distribution of the distance ξ between photon entry- and exit-points, is well approximated by a gamma-type distribution for both homogeneous and fractal models:

$$\rho(\alpha, \langle \xi \rangle; \xi) = \frac{1}{\gamma(\alpha)} \left(\frac{\alpha}{\langle \xi \rangle} \right)^\alpha \xi^{\alpha-1} e^{-\alpha \xi / \langle \xi \rangle} \quad (2)$$

with $\alpha = \left(\frac{\langle \xi^2 \rangle}{\langle \xi \rangle^2} - 1 \right)^{-1}$

The improved "nonlocal" IPA albedo,

$$R_{NIP}(\chi) = \frac{1}{2} \int_{-\infty}^{\infty} R_{IP}(\chi') \rho(\alpha, \langle \xi \rangle; |\chi - \chi'|) d\chi' \quad (3)$$

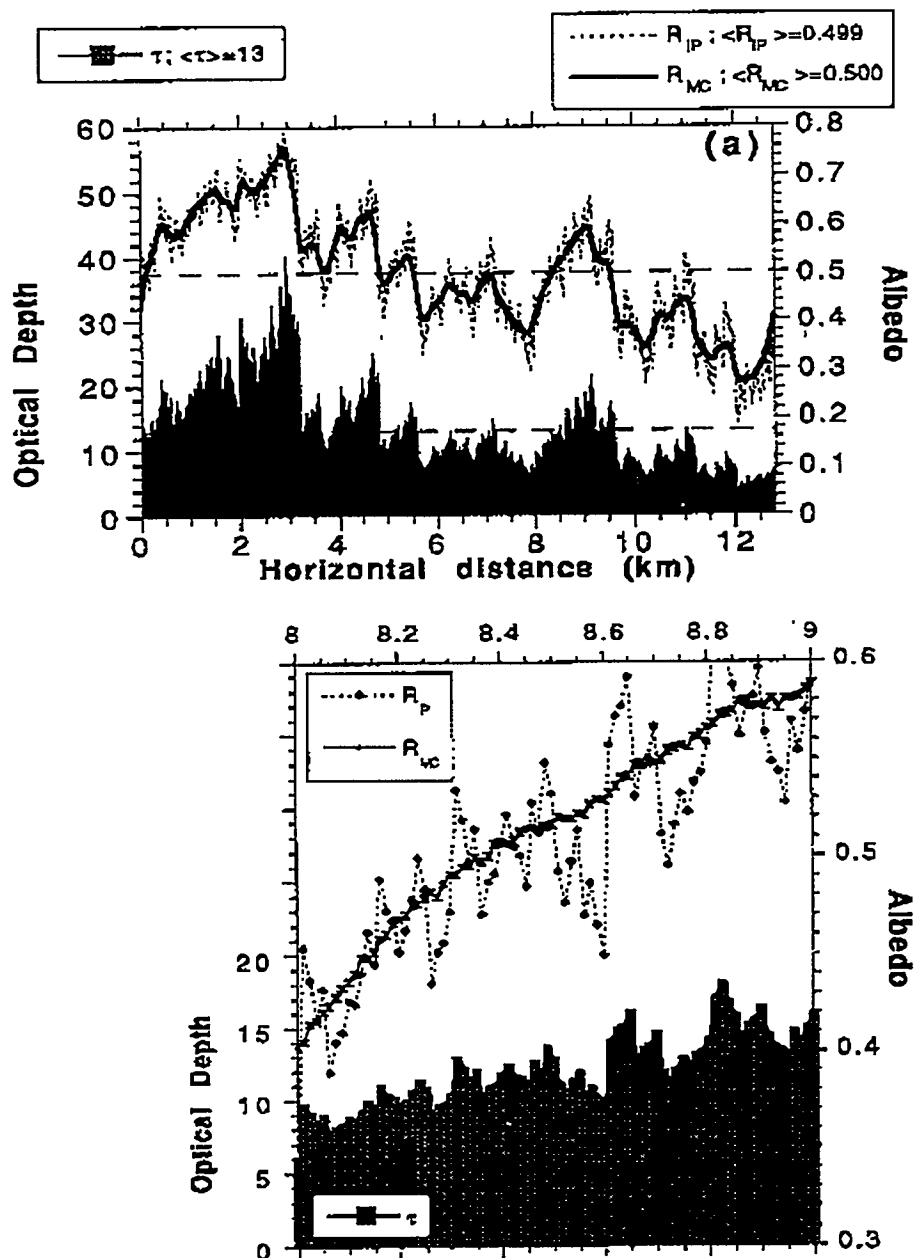


Figure 1. Horizontal variations of optical depth, IPA and MC albedos. (a) The lower curve is optical depth simulated by a fractal bounded cascade model with 10 cascade steps and $\langle \tau \rangle = 13$ (left axis); the two upper curves are the IPA and MC albedo fields (sun angle $\theta = 22.5^\circ$, asymmetry factor $g = 0.85$). Geometrical thickness $h = 300$ m, horizontal pixel size is 12.5 m. The basic cloud element is 0.3 km thick and $2^{10} \times 12.5 \cdot 10^{-3} \approx 12.8$ km long (there is no variability in the y direction). MC and IPA yield almost equal domain-average albedo 0.5 but they show radically different degrees of smoothness (the thickness of the MC curve shows the level of its numerical noise). (b) A one km zoom of panel (a) where the MC noise is now visible.

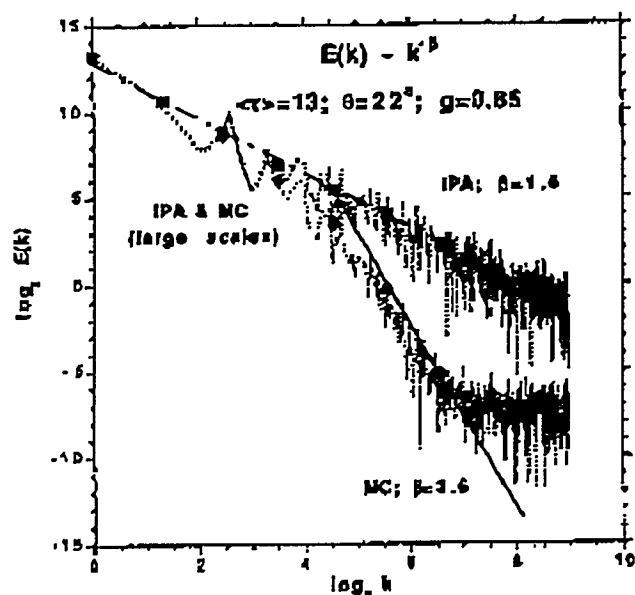


Figure 2. Energy spectra of IPA- and MC-simulated albedo fields. Parameters are the same as in Figure 1.

generates smoother fields at small scales and leaves large-scale properties unchanged (Marshak et al. 1995). To approximate the MC albedo results in Figure 1 starting with the IPA, we take $p(\alpha, \xi; \xi)$ with $\alpha = 1/2$ for simplicity and $\langle \xi \rangle = 0.215$ km from Equation (1). In Figure 3 we plotted the three albedo fields (R_{IP} , R_{MC} and R_{NIP}) highlighting the small residuals between R_{MC} and R_{NIP} .

If the wavenumber spectrum $E(k)$ of cloud liquid water (and, hence, that of the associated IP) follows the $k^{-5/3}$ power law, convolution with the spot shape yields a field with a spectrum

$$E(k) \sim k^{-5/3} \frac{\cos^2[\alpha \tan^{-1}(k\langle \xi \rangle/\alpha)]}{[1 + (k\langle \xi \rangle/\alpha)^2]^\alpha} \quad (4)$$

For scales larger than $\langle \xi \rangle$, we find both cloud liquid water and albedo have the same scaling behavior, while for scales less than $\langle \xi \rangle$, the albedo is much smoother, in agreement with analyses of Landsat observations of marine Sc by Cahalan and Snider (1989) and with our accurate MC results.

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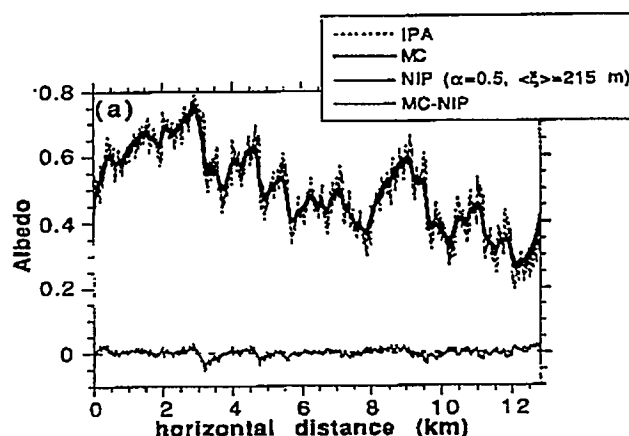


Figure 3. Nonlocal independent pixel approximation. Fractal model for optical depth is the same as in Figure 1. Three albedo fields calculated by the IPA, MC, and the improved IPA in Equation (3) are presented. The first two fields are the same as in Figure 1. The lower curve shows the differences between MC and the new “nonlocal” IPA.

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The Great Plains Low-Level Jet (LLJ) During the Atmospheric Radiation Measurement (ARM) Intensive Observation Period (IOP)-4 and Simulations of Land Use Pattern Effect on the LLJ

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Introduction

The Great Plains low-level jet (LLJ) is an important element of the low-level atmospheric circulation. It transports water vapor from the Gulf of Mexico, which in turn affects the development of weather over the Great Plains of the central United States. The LLJ is generally recognized as a complex response of the atmospheric boundary layer to the diurnal cycle of thermal forcing. Early studies have attributed the Great Plains LLJ to the diurnal oscillations of frictional effect (Blackadar 1957), buoyancy over sloping terrain (Holton 1966), and the blocking effects of the Rocky Mountains (Wexler 1961). Recent investigations show that the speed of the LLJ is also affected by the soil type and soil moisture (Fast and McCorcle 1989). Some studies also suggest that synoptic patterns may play an important role in the development of the LLJ (Uccellini 1980; Chen and Kpaeyeh 1992).

Land surface heterogeneities significantly affect mesoscale circulations by generating strong contrasts in surface thermal

fluxes (Mahfouf et al. 1987; Segal et al. 1989). Thus one would expect that the land use pattern should have effects on the LLJ's development and structure. In this study, we try to determine the relative roles of the synoptic forcing, planetary boundary layers (PBL) processes, and the land use pattern in the formation of the LLJ using the observations from the Atmospheric Radiation Measurement (ARM) Intensive Operation Period (IOP)-4 and numerical sensitivity tests.

Analysis of the Observations

During ARM IOP-4 (June 15-25, 1993), LLJs were well observed on June 16, 17, 21, 22, 23, and 24 (Figure 1). During the period, powerful storms swept through the nation's midsection, accompanied by tornadoes, high wind, large hail, and heavy rain. According to the National Oceanic and Atmospheric Administration (NOAA) weekly weather summary for the two weeks of June 13-19 and June 20-26, 1993, the LLJs played an important role in the weather events.

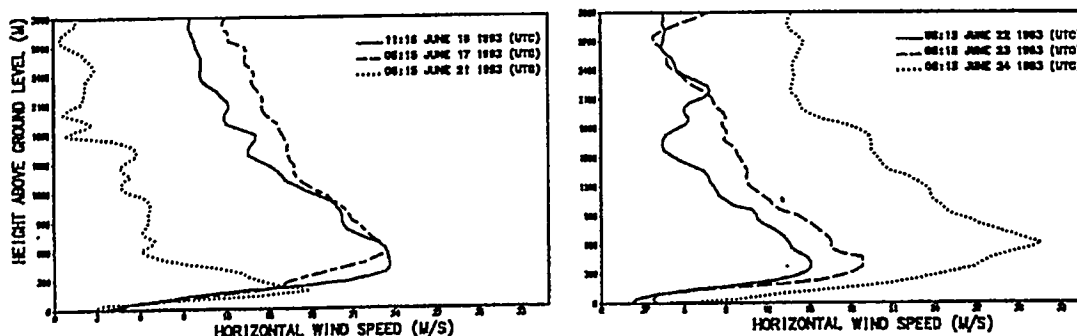


Figure 1. The low-level jets during ARM IOP-4, June 15-25 1993 (soundings at Kingman Station, Kansas).

The analysis of 300-mb maps at 0000 UTC for the six LLJs shows that two basic synoptic circulation patterns existed. The synoptic patterns for the LLJs on June 16, 17, 23, and 24 consisted of a strong trough over the Rockies and a ridge located in the eastern part of the country. The synoptic patterns for LLJs on June 21 and 22 consisted of a strong ridge over the Great Plains and a trough over the eastern United States. These two synoptic patterns are shown in the 300-mb maps at 0000 UTC for two typical days (June 21 and 24) (Figure 2). For June 21 and 22, the LLJs were at a relatively lower position with a weaker speed. The jet on June 22 had the lowest speed (15.1 ms^{-1}), while the jet on June 24 had the strongest speed (31.5 ms^{-1}). One jet (June 23) reached its peak at 0500 UTC, three jets (June 21, 22, and 24) reached their peaks at 0800 UTC, and the other two at 1100 UTC. Thus, there are some similarities in the evolutions of the six LLJs: each LLJ was associated with nocturnal temperature inversion and had strong diurnal oscillation, formed in the late afternoon or early evening, reached its peak in the next day early morning, and lasted for about 9–12 hours. Each LLJ was found in the ambient southerly flow over the central and southern Great Plains. The sounding profiles for the six LLJs indicate that they were mainly established by boundary layer processes and their position and strength were significantly affected by synoptic patterns.

Simulations of Land Use Pattern Effect on the LLJ

Model Description

This study uses the North Carolina State University mesoscale model which consists of three systems: the atmosphere, the vegetation, and the soil. The atmospheric portion of the model is three dimensional, anelastic, and

hydrostatic with a terrain-following sigma (σ) coordinate system. The surface layer is based on the similarity relations of Businger et al. (1971). Above the surface layer, prognostic turbulent kinetic energy and turbulent dissipation equations are used. A two-time-level fourth-order Crowley scheme is used for horizontal advection, while a quadratic upstream interpolation scheme is used for vertical advection. The model is linked with the soil-vegetation system to investigate the effects of the surface forcings on the LLJs. The parameterization of the soil-vegetation system is similar to that of Deardorff (1978). A single layer of vegetation with negligible heat capacity is assumed to be present and is characterized by its physical height, canopy density (shielding factor), optical properties (albedo and emissivity), and stomatal resistance. The foliage surface temperature is determined by the energy budget for the foliage layer. The air temperature and moisture within the canopy are represented by corresponding properties of the air immediately above the canopy, the soil, and the vegetation itself. The soil system consists of a shallow top layer and a deep bottom layer. The force restore method is used to calculate the soil surface temperature (T_g) which depends on the energy budget at the surface and a restoring term containing the deep soil temperature. The soil surface moisture is estimated in the same way.

Numerical Experiments

The model was run in its two-dimensional version in this study. The model domain is 450 km long and 10 km high with 45 grids in the horizontal and 25 in the vertical. The vertical grid spacing varies from 30 m to 1000 m with higher resolution at lower levels. The time step is 10 seconds.

Two ideal land use patterns dependent on vegetation type, soil type, soil moisture, and their distribution were chosen for this study. One is of oasis type which is an area of vegetation within arid surroundings. The other is with

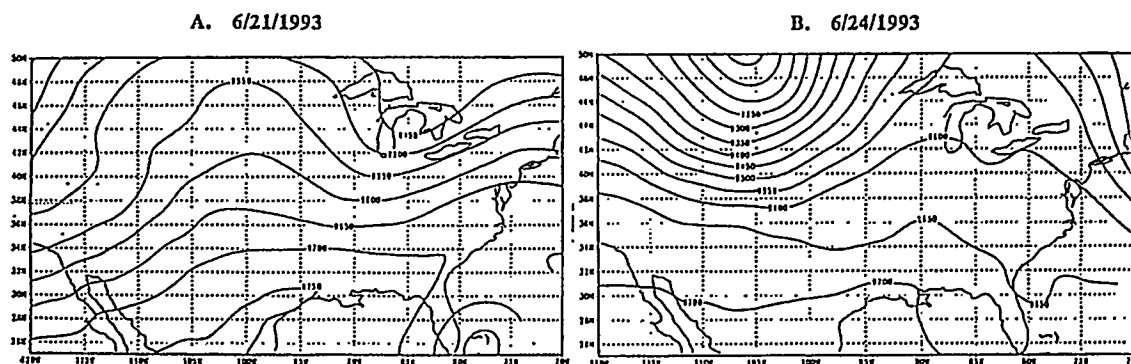


Figure 2. The 300 mb maps for 00 UTC 21 and 24 June 1993.

contrasting bare soil and vegetated areas. The soil and vegetation input parameters are given in Table 1. Two simulations were run for each pattern with and without initial wind. The initial surface temperature is 297 K with a potential temperature lapse rate of $+3.5 \text{ K km}^{-1}$ below 3000 m and $+3.0 \text{ K km}^{-1}$ above 3000 m. The specific humidity corresponds to sounding at 1100 UTC June 21, 1993. The initial wind is southerly and uniform with a magnitude of 10 ms^{-1} . All the simulations commenced at 0800 local solar time (LST) which corresponds to time when the sensible heat fluxes become effective in the development of the convective PBL. The latitude used in the model is 37.4°N which is the latitude of Kingman station, Kansas. In order to show the diurnal variations of the LLJ structure, the model was run for 48 hours for each simulation.

Results

The results from the 24-48 hour simulation were analyzed. For the oasis simulations, the oasis consisting of dense forests with a height of 10 m and a shielding factor of 0.99 is located at the center of the model domain with a width of 110 km. For the simulation without initial wind, the simulated fields of temperature and wind are very similar to the results by Segal et al. (1988). Horizontal gradient of sensible heat flux is evident as a large temperature difference exists between the oasis and its surroundings. There are two

symmetric circulation cells. At 0200 LST, the magnitude of the V component is about five times that of the U component. Two pair of low and high centers of V component developed below 3000 m. Normally, if there is a synoptic southerly (northerly) wind, the high centers (the low centers) would develop into two LLJs. The results from the simulation with initial wind at 0200 LST are given in Figure 3. Two LLJs are clearly displayed in the field of the V component, one located at a height of 450 m and the other at 1800 m. The vertical motion in this case is about two times stronger than the motion in the previous case. The evolution of the LLJ with lower position is shown in Figure 3 by the vertical profile of the horizontal wind speed at grid point 10 at different times. Both the jets formed around 2200 LST and reached their peak (17.5 ms^{-1}) around 0200 LST.

For the contrasting bare soil and vegetated area, the contrast consists of a relatively dry bare soil region of 210 km width in the west and a wet soil region of 240 km width in the east covered by dense forest. For the simulation without initial wind, a large difference in potential temperature exists between the dry bare soil region and the dense forest with wet soil. The circulation generated in this case is stronger than in the oasis case. Two regions of maximum winds are clearly displayed in the V wind field at 0020 LST. The simulation fields of V and W components of the wind and potential temperature at 0200 LST for the simulation with initial wind are provided in Figure 4. A narrow jet structure is at the 600-m level in the V field. The evolution of the LLJ is shown in Figure 4. The LLJ formed around 2200 LST and reached its peak (21 ms^{-1}) around 0200 LST.

The simulated LLJs are generated by a strong horizontal temperature gradient caused by land surface heterogeneities and a southerly synoptic wind. Although the LLJs are under ideal conditions, they have some similarities to the observed ones: 1) They have significant diurnal oscillation, forming at midnight and reaching their peak in the early morning; 2) they are generated in the ambient southerly flow; and 3) the heights of the LLJs near the surface are consistent with the observations.

Summary

Analysis of the soundings during ARM IOP-4 indicated that the synoptic circulation patterns for the LLJs can be classified into two categories. The LLJs under the more favorable synoptic pattern have relatively higher position and stronger speed. However, each of the six LLJs is associated with a temperature inversion and has significant diurnal oscillation. The simulated LLJs also have significant diurnal oscillation,

Table 1. Soil and Vegetation Input Parameters

1. Soil	Sand		
Cv	$1.256 \times 10^6 \text{ J m}^{-2} \text{ K}^{-1}$	k	$0.293 \text{ J m}^{-1} \text{ K}^{-1} \text{ s}^{-1}$
Dh	$0.233 \times 10^6 \text{ m}^2 \text{ s}^{-1}$	z_0	0.01 m
T	23.8°C		
2. Vegetation	Forest		
Z_a	10.0 m	of	0.99
R_{min}	100.0 s^{-1}	α_f	0.15
E_f	0.98	z_0	1.0 m

Cv = soil volumetric heat capacity; k = soil thermal conductivity; D_h = soil thermal diffusivity; z_0 = roughness; T = soil temperature; Z_a = vegetation height; of = vegetation shielding factor; R_{min} = minimum stomatal resistance; α_f = albedo of the vegetation; E_f = emissivity of the vegetation.

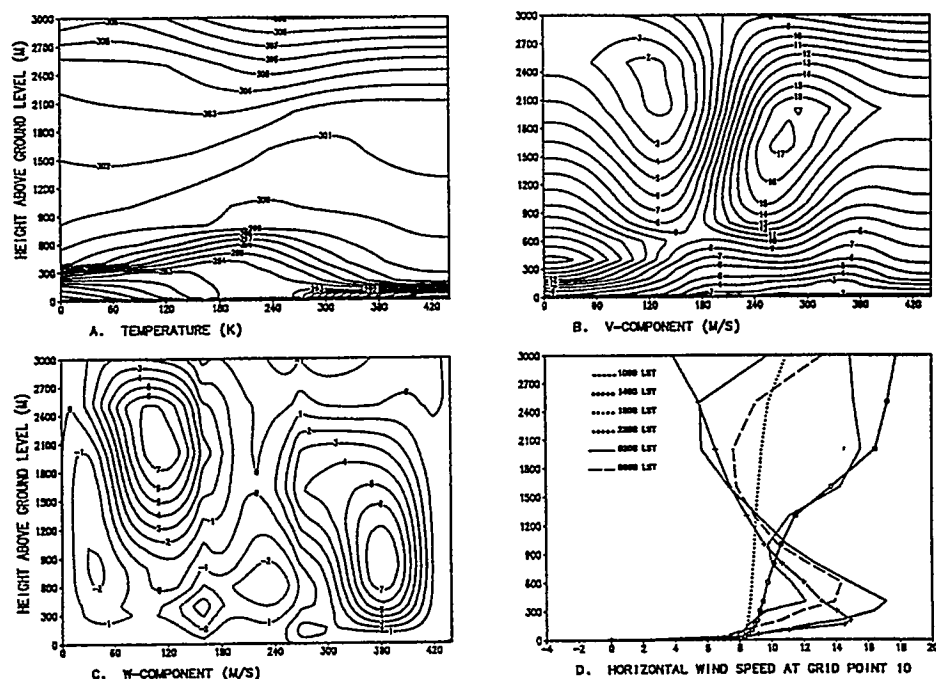


Figure 3. Simulation results from the Oasis Case.

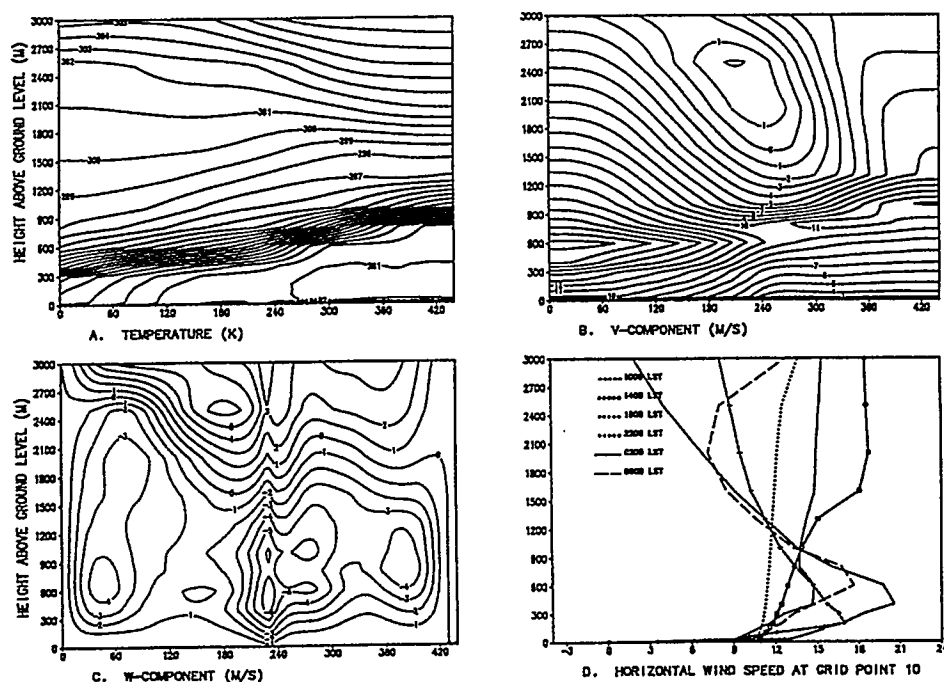


Figure 4. Simulation results from the bare soil-forest contrast.

form around 2200 LST, and reach their peaks around 0200 LST. Both the soundings and the numerical tests indicated that each of the LLJs is mainly established by the boundary layer processes. The mesoscale circulations generated by the surface heterogeneities strengthen the LLJs.

Acknowledgments

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Clouds and Snowmelt on the North Slope of Alaska

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Introduction

Clouds have a large effect on the radiation field. Consequently, possible changes in cloud properties may have a very substantial impact on climate. Of all natural surfaces, seasonal snow cover has the highest surface albedo, which is one of the most important components of the climatic system. Interactions between clouds and seasonal snow cover are expected to have a significant effect on climate and its change at high latitudes.

The purpose of this paper is to investigate the sensitivity of the surface cloud-radiative forcing during the period of snowmelt at high latitudes. The primary variables investigated are cloud liquid water path (LWP) and droplet equivalent radius (r_e). We will also examine the sensitivity of the surface radiative fluxes to cloud base height and cloud base temperature.

Model Description

A one-dimensional atmospheric radiative transfer model combined with the surface energy balance equation was employed in these investigations. The surface radiation fluxes were calculated using a comprehensive atmospheric radiative transfer model developed by Tsay et al. (1989). The atmospheric radiative transfer model uses a discrete-ordinate approximation to solve the radiative transfer equation (Stamnes et al. 1988) for both solar and terrestrial radiation. The model has detailed routines for including the effects of clouds, arctic haze, CO_2 , H_2O , O_3 , and snow layers on the atmospheric radiative transfer under arctic conditions.

Sensible and latent heat fluxes are calculated using bulk aerodynamic formulae following Louis (1979) and Ebert and Curry (1993). The ground-soil heat flux is computed using a snow-active layer-permafrost model (Zhang 1993). The lower boundary was set at 1 m below the snow-soil interface, and a constant temperature lower boundary condition was used in this study.

Cloud properties were specified based on the measurements for Arctic stratus by Herman and Curry (1984) and Tsay and

Jayaweera (1984). The geometrical thickness is 291.7 m with cloud base at 700 m for low Arctic clouds. Liquid water content ranges from 0.01 g m^{-3} to 0.40 g m^{-3} for low Arctic clouds, with the droplet (r_e) ranging from $4 \text{ }\mu\text{m}$ to $11 \text{ }\mu\text{m}$. Within the cloud layer, the relative humidity of the atmosphere was set at 100% to calculate the atmospheric water vapor content.

Results and Conclusions

Figures 1A-C show the variations of surface radiation fluxes with LWP and (r_e). Compared with clear sky conditions, clouds reduce the surface net solar radiation significantly, up to 80% (Figure 1A). Clouds increase the incoming longwave radiation from the atmosphere to the snow surface by as much as 30% or more (Figure 1B). The net surface radiation flux (R_N) is greater under cloudy sky conditions than under clear sky conditions (Figure 1C).

As can be seen from Figure 1C, the rapid increase in incoming longwave radiation with LWP dominates for small LWP. As the cloud thickens to the point of high opacity in the thermal infrared so that the incoming longwave flux levels off, the continued decrease in net solar radiation with increasing LWP comes to dominate the net radiation flux, leaving a maximum at moderate LWP.

Plots of shortwave, longwave, and total cloud radiative forcing against LWP simply involve relabeling the axes, and are not shown here. Instead, Figure 1D shows contours of the total cloud radiative forcing as a function of LWP and the droplet (r_e). As can readily be seen, the maximum value of C occurs for (r_e) approximately equal to $9 \text{ }\mu\text{m}$ and a LWP of about 29 g m^{-2} .

Figure 2 illustrates the effect of low Arctic clouds on the surface radiative fluxes and snowmelt at high latitudes. The net surface solar irradiance increases with time but decreases significantly with increase of LWP (Figure 2A). The net surface longwave radiation increases rapidly for lower LWP values (Figure 2B). Changes to the net surface longwave radiation are very limited when cloud LWP is higher than about 20 g m^{-2} . Figure 2C shows the net radiative balance as a function of LWP and time. The maximum values of R_N at a given time occur at LWP

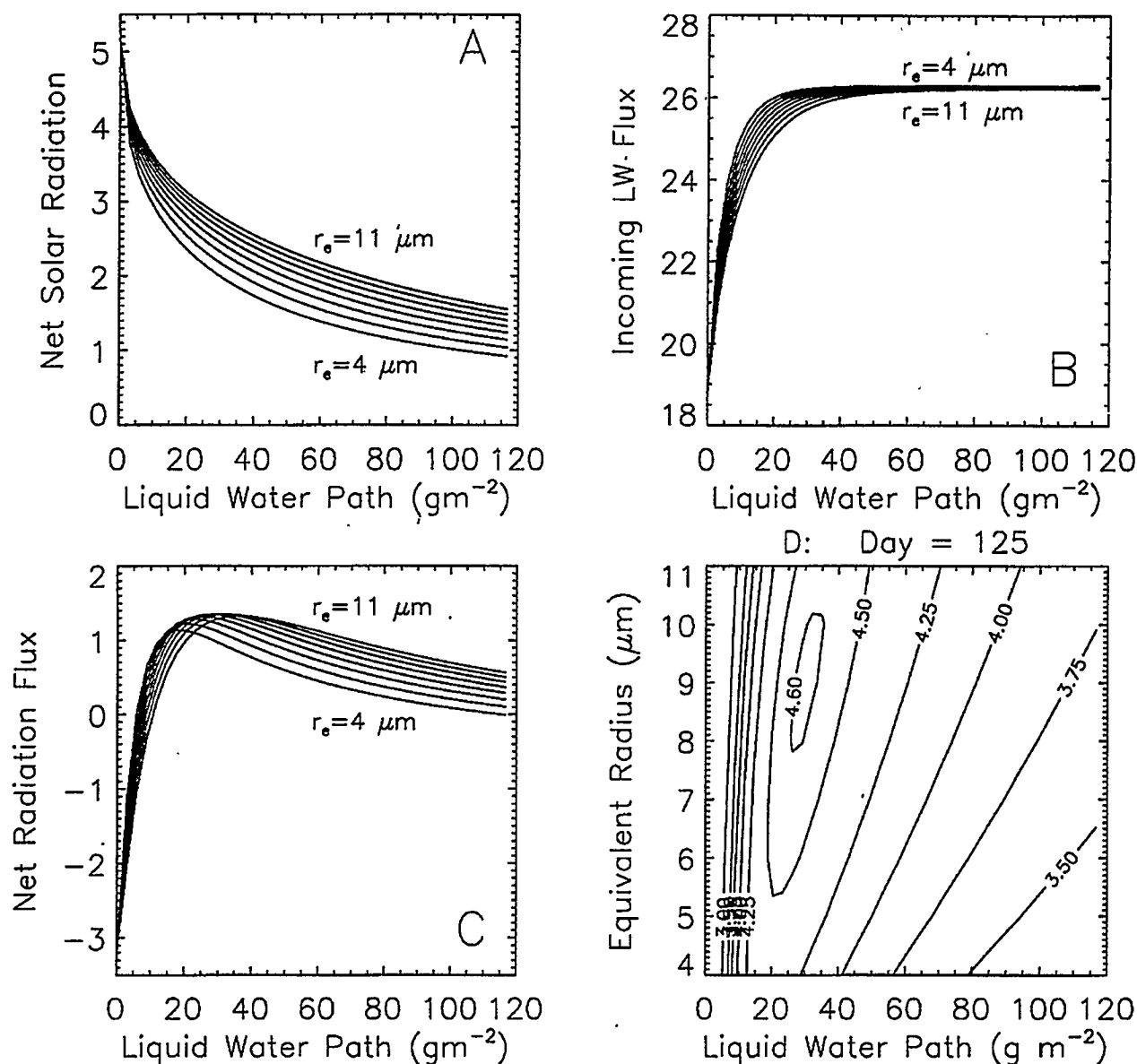


Figure 1. Effect of cloud liquid water path and equivalent radius on the surface radiative fluxes (A: net daily solar radiation; B: net daily incoming longwave radiation; C: net daily radiation balance) and the surface cloud-radiative forcing (D). Units are in MJ m² day⁻¹.

~25 g m⁻² during the early period of snowmelt and at LWP ~ 20 g m⁻² during the late period of snowmelt. As a result, the onset of snowmelt follows a similar pattern (Figure 2D). The onset of snowmelt under cloudy sky conditions happens earlier than under the clear sky conditions (LWP = 0). The timing of the earlier onset of snowmelt due to the effect of low clouds could range from a few days to over a month depending upon the cloud properties (Figure 2D).

This study confirms observational evidence gathered over the last three decades that the net daily radiation balance over a snow surface at high latitudes is larger under cloudy sky conditions than under clear sky conditions. Ambach (1974) suggested that the enhancement of the net surface radiation balance under overcast conditions is due to the significant influence of the incoming longwave radiation and the high albedo of the snow surface. This study supports that suggestion and further indicates that

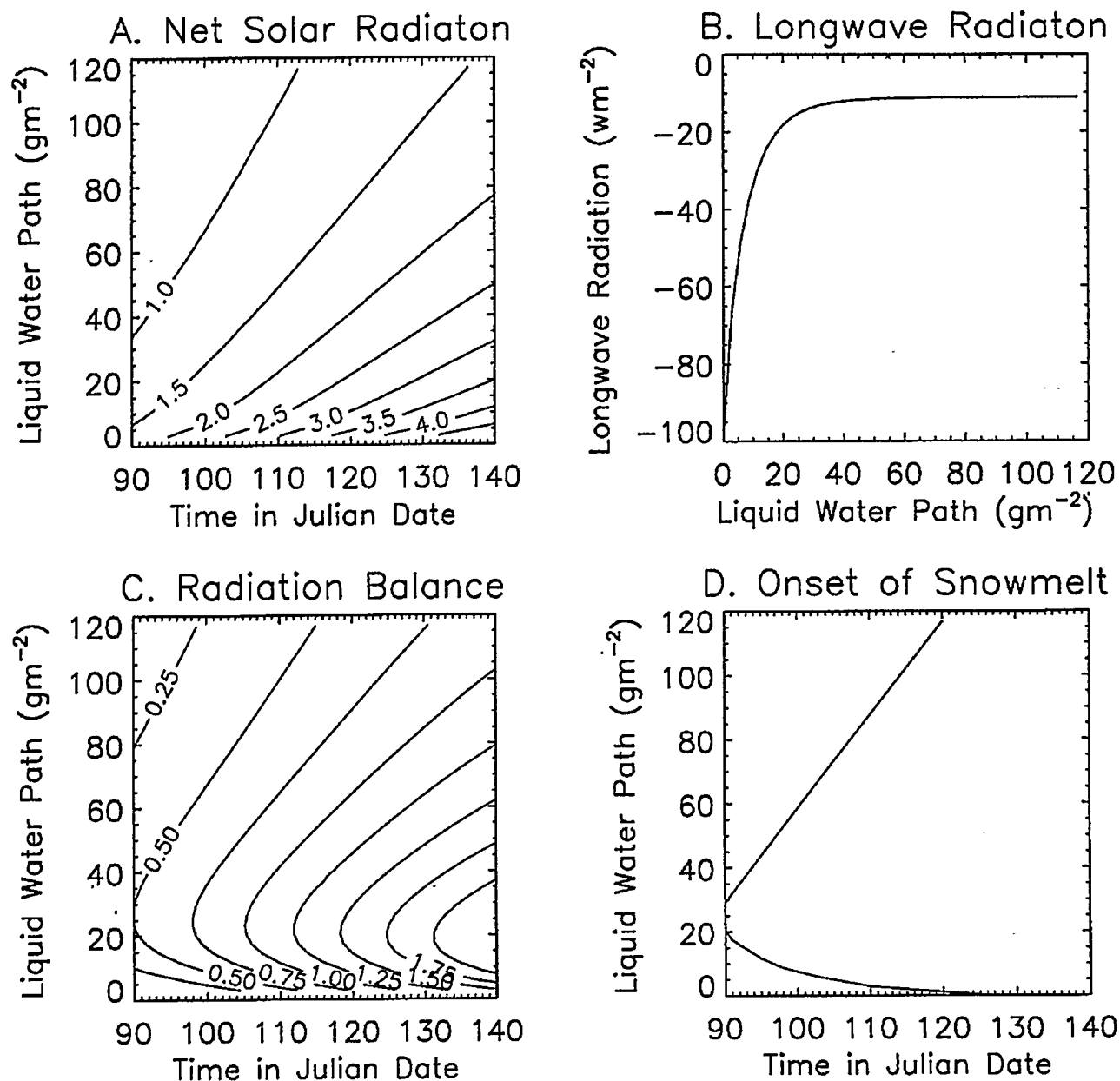


Figure 2. Effect of low clouds on the surface radiative fluxes and the onset of snowmelt at high latitudes. Units for the contours in (A) and (C) are in $\text{MJ m}^{-2} \text{ day}^{-1}$.

clouds exert a strong control on the surface radiation balance by reducing the incident solar radiation and increasing the incoming longwave radiation. However, the warming effect due to surface longwave cloud-radiative forcing is indeed greater than the cooling effect due to surface shortwave cloud-radiative forcing,

which results in an enhancement of the net radiation balance under cloudy sky conditions that warms the snow surface and the lower atmosphere. The magnitude of this enhancement under overcast conditions depends strongly upon cloud microphysical properties, cloud base height, cloud base temperature, and cloud thickness.

This study shows that the impact of changes in cloud conditions, such as cloud microphysical properties (r_c and LWP), cloud base height and temperature, on the onset of snowmelt could range from a few days to over a month, which is sufficient to account for the observed inter-annual variations in the timing of snowmelt at high latitudes.

This study also illustrates that clouds have a positive feedback on the climate system at high latitudes, at least during the period of snowmelt. This feedback is caused by positive surface cloud-radiative forcing leading to earlier snowmelt. Some general circulation models predict that temperature and thereby evaporation would increase due to elevation of the CO_2 content in the atmosphere. A warmer atmosphere has a greater moisture-bearing capacity, and enhanced evaporation implies an increase in cloudiness unless the precipitation is significantly changed. In the Arctic, the abundance of cloud condensation nuclei may prevent clouds from precipitating more in a warming climate. Thus the end result may be an enhanced moisture-cloud radiative warming and a positive feedback loop. Earlier snowmelt would reduce the surface albedo and allow the surface to absorb more solar radiation and warm the atmosphere. The coupling between these two positive feedback loops (moisture-cloud radiative warming and earlier snowmelt) could result in an amplified feedback to the climate system in the Arctic.

Finally, this study suggests that comprehensive field measurements of cloud physical properties (including cloud liquid or ice water content, cloud particle size and shape, cloud height and morphology), atmospheric temperature and moisture conditions, surface and near surface radiation, wind, humidity, and surface optical properties, are required to improve our physical understanding of the transfer of energy, momentum, and mass between the atmosphere and the surface. Better knowledge of these interactions over land as well as over the ocean in the Arctic is a prerequisite for improved comprehension of the Arctic climate system.

Acknowledgments

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A Case Study of the Great Plains Low-Level Jet Using Wind Profiler Network Data and a High Resolution Mesoscale Model

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Introduction

The Great Plains low-level jet (LLJ) has important effects on the life cycle of clouds and on radiative and surface heat and moisture fluxes at the Atmospheric Radiation Measurement (ARM) Southern Great Plains (SGP) Cloud and Radiation Testbed (CART) site. This diurnal phenomenon governs the transport and convergence of low-level moisture into the region and often leads to the development of clouds and precipitation. A full understanding of the life cycle of clouds at the SGP CART site and their proper representation in single column and global climate models cannot be obtained without an improved understanding of this important phenomenon.

A detailed case study of one complete episode of a typical summertime LLJ using data collected by the National Oceanic and Atmospheric Administration (NOAA) wind profiler demonstration network and soundings released at the CART central facility is presented. A two-day period beginning on the morning of 11 July 1992 and ending on the morning of 13 July 1992 was chosen for this study because a typical summertime LLJ was evident over many of the observational sites. The high temporal and spatial resolution of the data permits a much more detailed picture of the Great Plains LLJ and its associated mesoscale convective systems than is possible from previous studies. A three-dimensional mesoscale numerical model is also used to simulate the episode and to provide information on the physical mechanisms responsible for the initiation, evolution, maintenance, and decay of the LLJ. The model is also used to examine the sensitivity of the LLJ to various factors such as the beta effect, soil moisture content and distribution, and the possible feedbacks from cloud and precipitation processes.

Data

The data used in this study are vertical wind profiles from the wind profiler demonstration network and temperature

and moisture soundings at the central facility. The profiler network consists of 31 stations mostly scattered across the southern and central Great Plains. Each wind profiler provides hourly vertical profiles of horizontal wind components. The geographic location of this profiler network (Figure 1) and the high temporal resolution of the wind profile measurements make it ideal for depicting the spatial variation and the diurnal evolution of the LLJs.

Model

The model used for the case study is the Colorado State University Regional Atmospheric Modeling System (RAMS) (Pielke et al. 1992). A nested grid configuration was used with a coarse grid of 100-km resolution covering most of the United States, including the Gulf of Mexico, and a fine grid of 25-km resolution covering most of the Great Plains. Thirty vertical levels were used for both grids. The

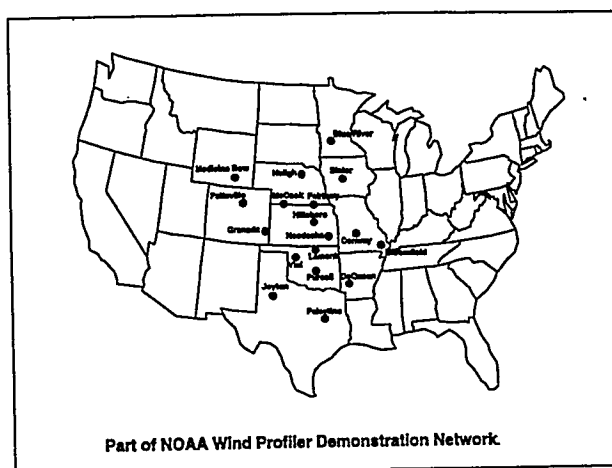


Figure 1. Locations of the wind profilers used for the present study in the NOAA wind profiler demonstration network. The SGP CART central facility is located at Lamont, Oklahoma.

simulation was initialized at 0600 CST, 11 July, using data from the National Meteorological Center global analysis, standard rawinsonde observations, wind profiles from the profiler network, and special soundings from the central facility. The simulation continued for 48 hours during which the lateral and the top boundary conditions were allowed to change by "nudging" the outermost nodes at these boundaries towards the objectively analyzed observational fields that were available every 12 hours.

Model Verification

The model's performance in simulating the LLJ in this case is evaluated by the observations. Figure 2 shows the comparison of the horizontal wind vectors at the height of the LLJ maximum at the time of maximum LLJ development. Very good agreement was obtained between the simulated and observed position of the jet core, the width of the jet core, the values of the wind maximum, as well as the associated wind shear and vorticity fields. A good agreement is also reached between simulated and observed precipitation fields, as shown in Figure 3. The maximum precipitation occurs at the leading edge of the jet due to low-level moisture convergence and strong upward motion produced by the LLJ. These comparisons suggest that the model is able to capture the principal features of the LLJ and associated vertical motion and precipitation fields, which justifies the use of the model for an in-depth analysis as presented below.

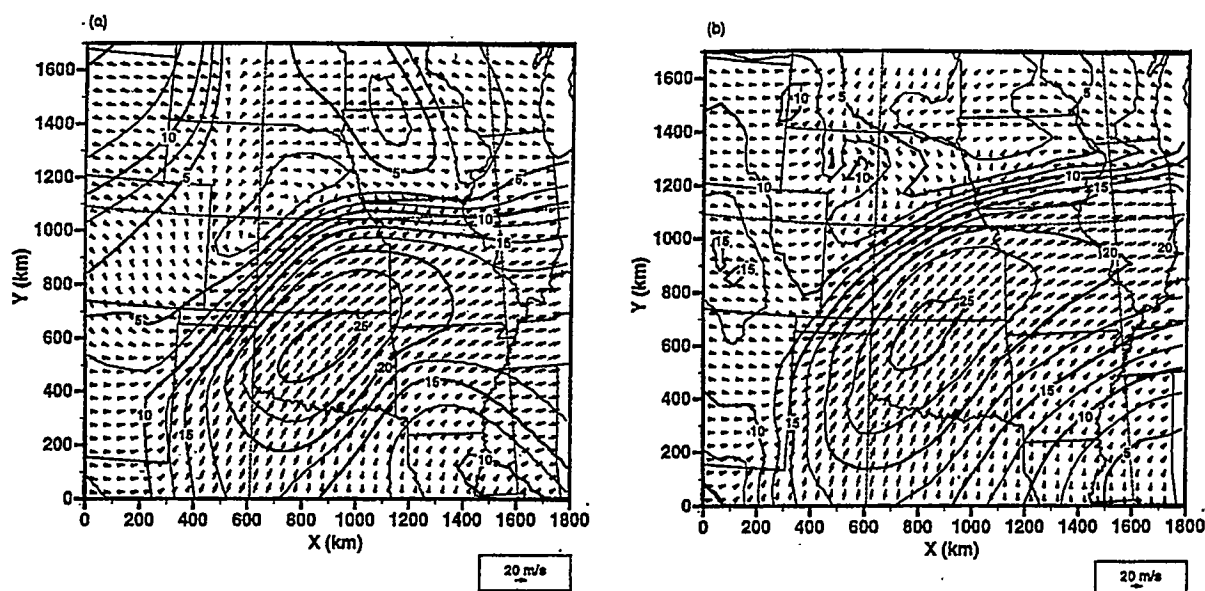


Figure 2. Observed (left) and simulated (right) wind speed contours and vector fields at 867 m at 0200 CST, 13 July 1992.

Momentum Balance Analysis

Because the model produces a dynamically balanced data set, it can be used to examine the forces that make up the individual components of the momentum equation. These forces control the momentum balance and, hence, account for spatial and temporal variations of the wind field.

The horizontal momentum equation can be written as

$$\frac{\partial \vec{v}}{\partial t} = \underbrace{-\vec{v} \cdot \nabla \vec{v}}_{\text{I}} - \underbrace{\alpha \nabla p}_{\text{II}} - \underbrace{2\vec{\Omega} \times \vec{v}}_{\text{III}} + \underbrace{\alpha \vec{F}}_{\text{IV}} \quad (1)$$

This equation states that wind accelerations are driven by the balance between horizontal advection (term I), pressure gradient force (term II), Coriolis force (term III), and friction (term IV).

The diurnal evolution of the individual momentum equation terms shows similar patterns at stations within the jet core. An example is shown in Figure 4 for Lamont, Oklahoma. Clearly, within the jet core, the momentum balance is determined predominantly by the pressure gradient force, Coriolis force, and friction during daytime, and by the pressure gradient and Coriolis forces during nighttime. The Coriolis force shows a strong diurnal oscillation, which,

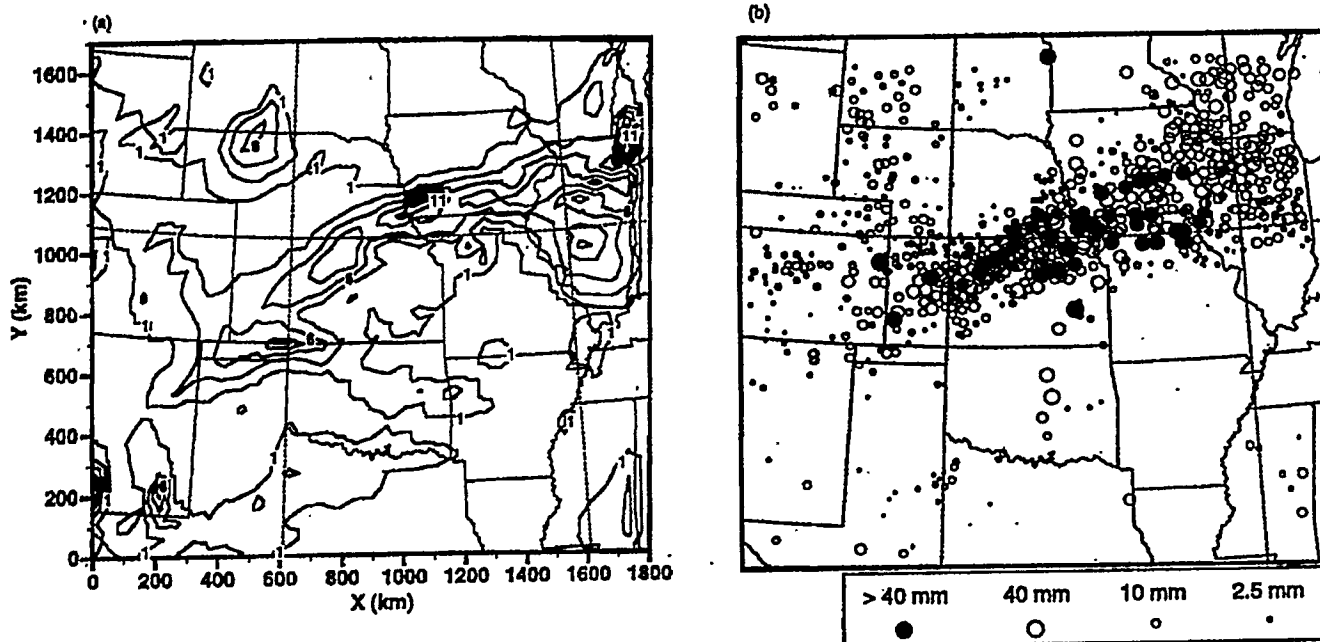


Figure 3. Simulated (a) and observed (b) 24-h total surface precipitation (in mm) ending at 0600 CST, 13 July. The solid circles indicate total precipitation greater or equal to 10 mm, the open circles indicate total precipitation between 1 mm to 10 mm, and the x indicates precipitation amount less than 1 mm.

together with the diurnal oscillation of friction, accounts for most of the diurnal variation of total acceleration in the jet core. The pressure gradient force, on the other hand, shows a relatively small diurnal variation. This implies that frictional decoupling and inertial oscillation, as proposed by Blackadar (1957), are the primary mechanisms responsible for the diurnal variation of the LLJ. The terrain-induced pressure gradient oscillation over the sloping Great Plains appears to be less significant than was suggested in previous studies (Holton 1967). This is consistent with the analysis of ageostrophic wind components, which indicates that the isallobaric components arising from the changes in horizontal pressure gradients are secondary to the inertial components associated with the turning of the wind by the Coriolis force in the total ageostrophic wind.

In contrast to the stations within the jet core, stations west of the jet core, as represented by Platteville, Colorado in Figure 4, display strong diurnal oscillations of the pressure gradient force, especially in the u -momentum component, reflecting the effect of diabatic cooling and heating above the sloping terrain. Nevertheless, the pressure gradient force, or the geostrophic wind, is much smaller than at locations within the jet core. By examining the differences in pressure gradient force between the two stations in Figure 4, one can see that the pressure gradient force is considerably

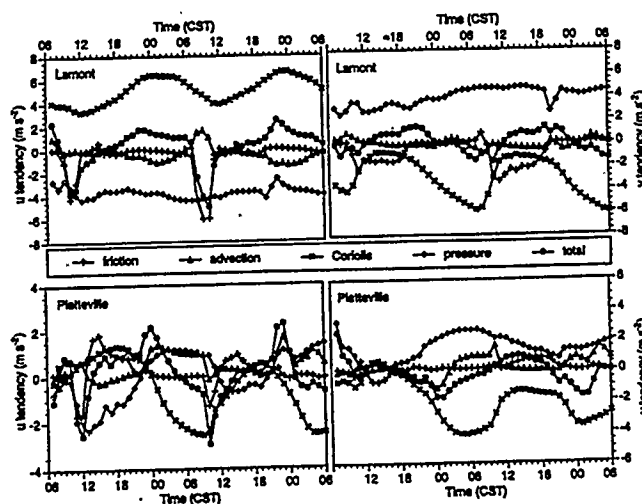


Figure 4. Time variation of the individual terms in the horizontal momentum equation at 867-m mode level at two stations. The time starts from 0700 CST, 11 July and ends at 0600 CST, 13 July.

smaller outside the jet core than inside. This suggests that a strong synoptic-scale pressure gradient is crucial for the formation of a strong LLJ. Without such a

large-scale pressure gradient, or the "basic flow" on which the frictional stress could operate diurnally, there would not be the strong diurnal LLJ that is so characteristic of the Great Plains.

Sensitivity Studies

Sensitivity tests were conducted to study the relative influence of various physical factors on LLJ characteristics. Simulations with a constant Coriolis parameter in the entire domain, as compared to the actual situation in which the Coriolis parameter varies meridionally, result in a weaker jet, confirming Wexler's (1961) prediction that the increase of Coriolis parameter with latitude enhances the LLJ. Simulations with different soil moisture values show that changes in soil moisture have little effect on the phase of the diurnal oscillation of the jet, but modify the amplitude of the oscillation with larger amplitudes associated with drier soil (Figure 5). However, considerably larger vertical velocities at the leading edge of the jet are found to be associated with wet soil than with dry soil. This enhancement of rising motions appears to be associated with latent heat release from precipitation; the simulation without cloud and precipitation processes displays much weaker rising motions ahead of the jet than the experiment retaining the processes.

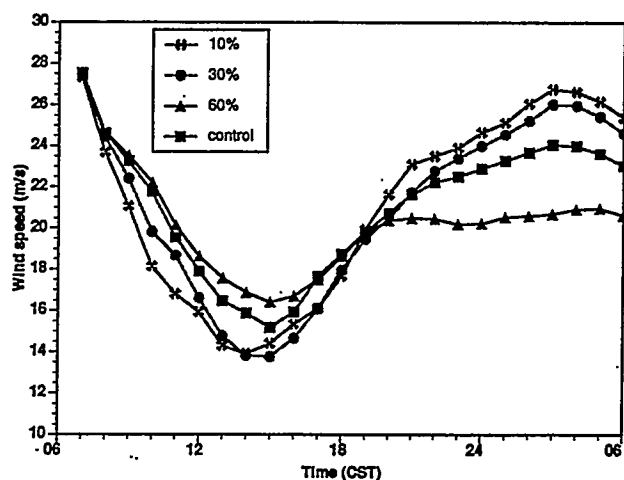


Figure 5. Time variation of maximum wind speeds averaged over the 698-, 867-, 1071-m model vertical levels from 0700 CST, 12 July, to 0600 CST, 13 July, for simulations with different soil moisture contents. The control simulation uses a realistic soil moisture distribution derived from the crop moisture index.

Summary

A combined analysis and modeling approach has produced a detailed description of the three-dimensional structure of the LLJ and its time evolution. Relative contributions to the boundary-layer wind oscillation from different forces were diagnosed by reconstructing tendency terms in the momentum equations. The analyses suggest that the diurnal oscillation of horizontal pressure gradient over sloping terrain is secondary to the inertial oscillation mechanism resulting from the release of frictional constraint in the evening and throughout the night in driving the LLJ. Sensitivities of the LLJ to the beta effect, soil moisture content and distribution, and cloud and precipitation feedbacks were also investigated. The meridional variation of the Coriolis parameter as air moves northward appears to enhance the strength of the jet. A larger amplitude of the diurnal oscillation of the jet speed is found to be associated with drier soil, while rising motion at the leading edge of the jet is stronger for wetter soil. This enhanced vertical motion appears to be associated with latent heat release due to precipitation.

The results suggest that an improved turbulence treatment and an incorporation of information such as soil moisture are likely to improve the ability of forecast models and global climate models in simulating the LLJ and the associated nocturnal convection. Research has also demonstrated the ability of radar wind profilers to improve our understanding of a well-studied boundary-layer phenomenon. More wind profilers are now equipped with radio-acoustic sounding systems which allow vertical profiles of virtual temperatures to be measured as well. Observations from these profilers will provide a great opportunity to verify previous conceptions regarding the forcing of the LLJ.

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Some Results of an Experimental Study of the Atmospheric Aerosol in Tomsk: A Combined Approach

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As widely accepted, the atmospheric aerosol strongly contributes to the formation of the earth's radiation balance through the absorption and scattering of solar radiation. In addition to this direct influence on the radiation balance, aerosols, being active condensation nuclei, also have a role in the cloud formation processes. Consequently, we pay special attention to the block of aerosol studies in our Tomsk Radiation Experiment in Siberia (TRES) Project, proposed for the Atmospheric Radiation Measurement (ARM) Program.

Despite a strong spatio-temporal variability of the aerosol physicochemical parameters, their mean values could be quite characteristic of any specific region.

In this paper we present some results of aerosol studies undertaken on our own initiative at the field measurement sites of the Institute of Atmospheric Optics in Tomsk and the Tomsk region. To a certain extent these results reveal some regional features of the atmospheric aerosol formed over the central part of the Eurasian subcontinent, with its typically midlatitude climate characterized by a long, snowy winter and short, hot, and dry summer.

To keep our discussion within the scope of a single paper, we present here only our studies of aerosol in the near ground atmospheric layer. No results of lidar studies of aerosol in the boundary layer and in the free atmosphere are discussed here, though some of them are mentioned in our proposal on the TRES Project to the ARM Program.

Figure 1 presents the data on annual behavior of the aerosol number density (particles with the diameter $d \geq 0.4 \mu$) recorded in the ground atmospheric layer over Tomsk Akademgorodok in 1993, together with data on the annual behavior of the aerosol column density averaged over many years of airborne observations over West Siberia. As seen in Figure 1, the aerosol number density reaches its maximum during the cold season, while its minimum value is observed during warm periods. Two secondary maxima in July and November are obviously due to regional peculiarities in air

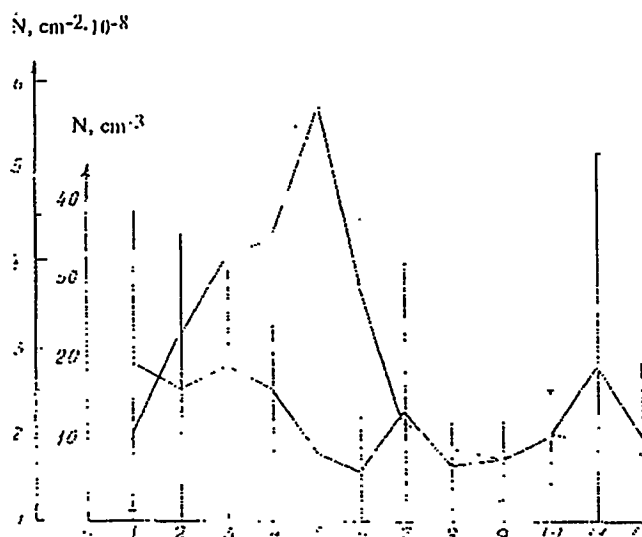


Figure 1. Annual behavior of the aerosol number density in the ground atmospheric layer (—) and of its column density (through the mixing layer) (o—o); vertical bars show rms deviations.

mass circulation normally observed in these periods. Such a behavior of the aerosol number density in the ground atmospheric layer is mainly determined by the mixing layer height, provided that aerosol sources are constant. The mixing layer height, in turn, reaches its maximum in July, while its minimum occurs in December-January. Seasonal behavior of the aerosol column density is, on the other hand, governed by joint action of several processes. Among these, the photochemical processes play the key role in the *in situ* aerosol particles formation during spring time when forests coming to life start to emit a lot of terpenic phytoncides into the atmosphere.

The intensity of various sources contributing to the formation of the aerosol field in the atmosphere varies during the year. The variability of the aerosol sources can result in changes of the particle size distribution and in different

chemical composition of the particulate matter. The trends in the annual behavior of different size fractions of aerosol can be seen in Figure 2.

As shown in Figure 2a, the annual behavior of the submicron fraction is very similar to that of the aerosol column density presented in Figure 1.

According to conventional understanding, size distribution of the atmospheric aerosol has a minimum in the region of 1μ size (that is, between the fine and coarse fractions of aerosol). The annual behavior of the number density of particles of this size is shown in Figure 2b. As shown in Figure 2b, the concentration of this fraction monotonically falls off to the level characteristic of the period from August to December, with the only exception in the May-June period.

The source of a coarse aerosol fraction is, as a rule, the underlying surface and partly the aged aerosol of photochemical origin. From the data presented in Figure 2c, it follows that the main source of coarse aerosol fraction in the

measurement site is the underlying surface because the concentration of particles of this size is practically zero in winter. In fact, it is clear because the underlying surface at that time was closed with snow, which could only be a sink for aerosols of coarse fraction and did not produce the aerosol itself. On the contrary, in the summer, stronger winds and dry underlying surface produce more coarse disperse particles which are more easily transported aloft due to stronger turbulence than is seen in the winter. At the same time, it is worth noting the similarity of the curve depicted in Figure 2c to the annual behavior of the aerosol column density shown in Figure 1. This circumstance allows us to assume that in the period of seasonal change from spring to summer the coarse fraction of aerosol is partly composed of particles of photochemical origin that are aging in the upper layers in the course of natural smog formation. Thus, large particles are then descending onto the underlying surface.

Because of regional siberian peculiarities characterized by long winter when snow covers the land from November till April and temperatures vary from 0°C to -30°C , it would be interesting to analyze data on the optical properties of the atmosphere during the winter. In fact, very little experimental data of field measurements of the atmospheric transmission have been available until recently. To fill this gap, a 3-week field measurement campaign was arranged and conducted at the Institute of Atmospheric Optics in the winter of 1992 to study the spectral transmission of the atmosphere in the spectral range from 0.44 to 12μ . These measurements were conducted along a 1-km horizontal path. We recorded about 180 transmission spectra of the atmosphere, of which we have arranged three data sets on the aerosol extinction coefficients $\alpha(\lambda)$ of a winter haze (59 spectra), an ice fog (35 spectra), and a snowfall (75 spectra). We separated the purely aerosol extinction from the total atmospheric extinction using a statistical method.

Mean $\alpha(\lambda)$ spectra for the three optical weather types are shown in Figure 3. Curve 1 in this figure presents the data on winter haze, Curve 2 the ice fog, and Curve 3 the data obtained in snowfalls. Note that ice fogs occur in the atmosphere under anticyclonic conditions and temperatures below 258 K . As shown in this figure, $\alpha(\lambda)$ spectra in winter haze and ice fog are very similar. This clearly demonstrates the fact that particle size spectra of the haze fraction of aerosol are quite stable in winter. So the ice fog only increases the total extinction by approximately 0.2 to 0.25 km^{-1} , this addition being spectrally neutral. This additional extinction is due to ice crystals that do not affect the size spectra of the fine aerosol fraction.

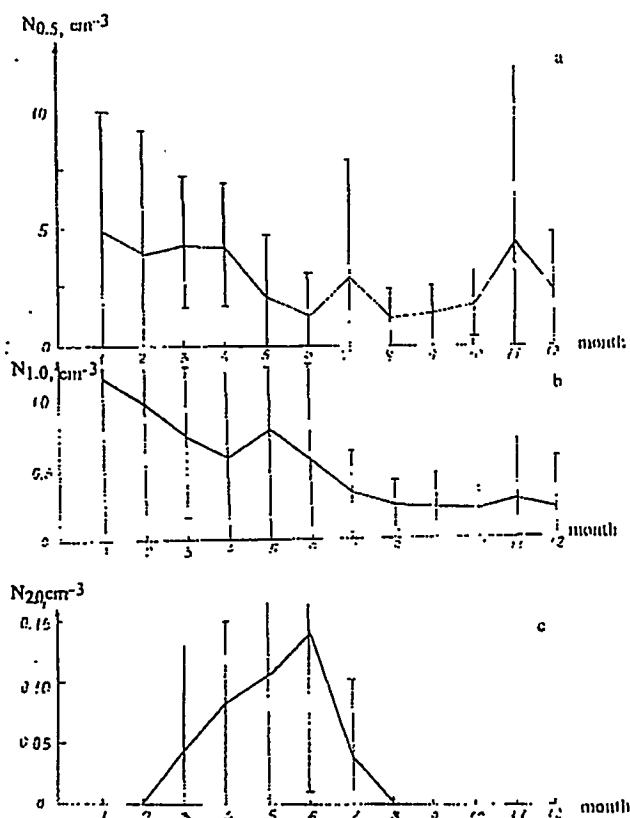


Figure 2. Annual behaviors of the three aerosol fractions.

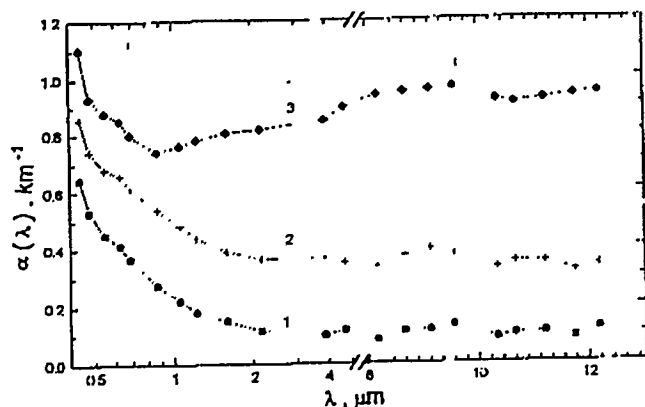


Figure 3. Spectral behavior of the aerosol extinction coefficient measured in a winter haze (1), an ice fog (2), and a snowfall (3).

This conclusion is well confirmed by the data presented in Figure 4. This figure shows the spectra of the parameter $\Delta\alpha(\lambda) = \alpha(\lambda) - \alpha(3.9)$ which characterizes only the contribution from fine aerosol fraction into the total extinction of radiation by atmospheric aerosol. Curves 1 and 2 of Figure 4 are practically coincident.

In contrast to the ice fog, snowfalls essentially modify the $\alpha(\lambda)$ behavior in winter haze and create a significant increase of the extinction over the entire spectral range. This takes place because of washing out of fine aerosol fraction (see Curve 3 in Figure 4). Once $\alpha(\lambda)$ spectra are measured, the distribution of a particle's geometric cross sections over the particle's radii can be reconstructed by solving the system of equations of the form

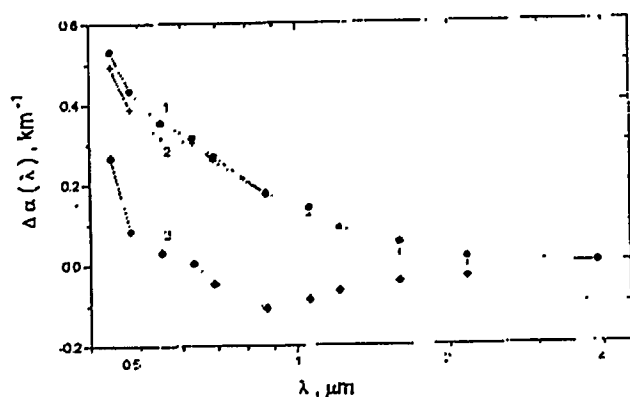


Figure 4. Spectral behavior of the parameter $\Delta\alpha(\lambda) = \alpha(\lambda) - \alpha(3.9)$ for a winter haze (1), an ice fog (2), and a snowfall (3).

$$\int_{R_1}^{R_2} K(r, \lambda_i) s(r) dr = \alpha(\lambda_i)$$

where r is the particle radius and $K(r, \lambda_i)$ is the extinction efficiency factor of a particle for radiation at the wavelength λ_i . The extinction efficiency factor also depends on the refractive index of particulate matter.

Figure 5 presents mean spectra $\alpha(\lambda)$ measured under different visibility conditions in winter hazes ($S_m > 20$ km in Curve 1 and $S_m < 5$ km in Curve 2) at ten wavelengths within the spectral region from 0.44 to 3.9 μ . The results obtained by inverting these data using Tikhonov regularization method are presented in Figure 6. The complex index of refraction was taken in this case according to a model which assumes it to be equal to $m = (1.36 - 10.00i)$ in the visible spectral region. When solving the inverse problem, we also considered the fact that in most situations the relative humidity of the air exceeded 90%. The $s(r)$ distributions reconstructed clearly reveal the cumulative fraction of aerosol in the submicron size range and the coarse fraction of aerosol with the maximum in the size range 2.6 to 2.9 μ .

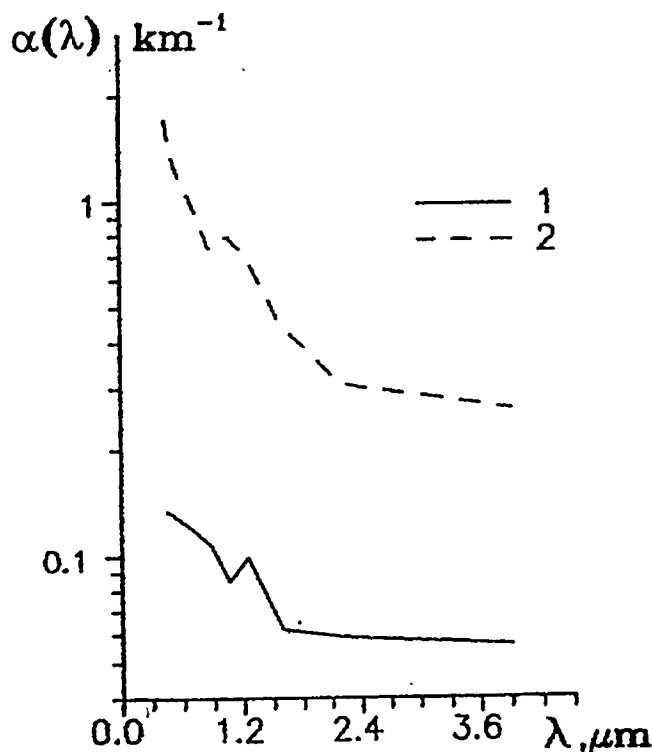


Figure 5. Spectral behaviors of the aerosol extinction coefficient (monthly mean measurement data for December 1992) obtained at meteorological visual ranges $S_m > 20$ km (1) and $S_m < 5$ km (2).

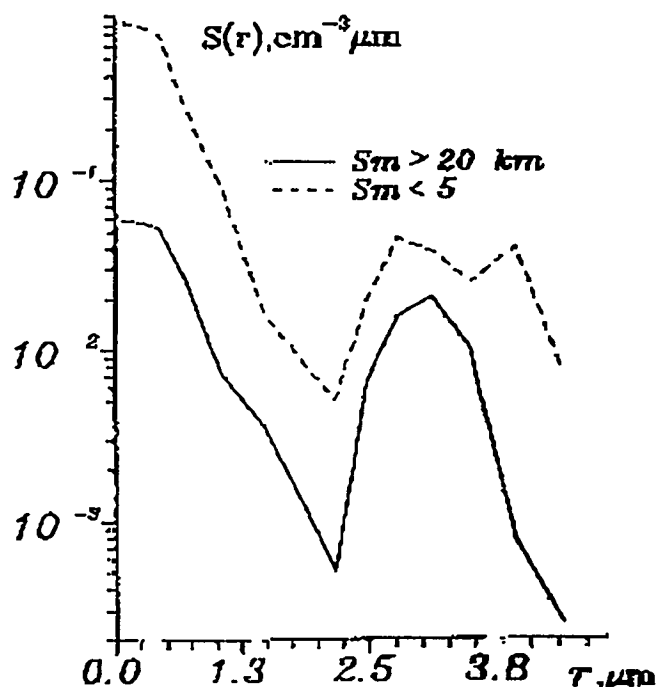


Figure 6. Size spectra of aerosol particles cross sections as reconstructed by inverting the extraction coefficients spectra presented in Figure 5.

The size distributions $s(r)$ and the value of the total geometric cross section corresponding to it are convenient for use in studies of the influence of particle size spectra on the optical properties of the atmosphere. However, in some applied studies like, for instance, studies of aerosol pollutions of air, it is more convenient to use the size distribution functions for a particle's volume [$v(r) = (4/3) \pi r^3 n(r)$] and the total volume of particles contained in 1 cm^3 . These quantities enable one to estimate mass concentration of aerosol per an interval of a size spectrum.

In microstructure studies of the atmospheric aerosol, the data of direct sizing measurements are very often approximated using a combination of lognormal distribution functions according to the following:

$$n(r) = \frac{1}{\sigma r \sqrt{2\pi}} \exp \left[- \left(\ln \frac{r}{r_m} \right)^2 / 2\sigma^2 \right]$$

where r_m is the median radius and σ is the standard deviation. It would be interesting to analyze the applicability of this formula to the description of results obtained by solving an inverse problem. Approximations of the size distribu-

tions $v(r)$ reconstructed by solving the inverse problem with a combination of two lognormal distribution functions are shown in Figure 7 by solid curves. For the distributions $v(r)$ reconstructed from data averaged over the entire data set, we have found that the parameter is $r_{m1} = 0.32 \mu$ and $\sigma = 2.27$ for the cumulative fraction and $r_{m2} = 2.74 \mu$ and $\sigma = 1.19$ for the coarse aerosol fraction. As shown in Figure 7, the main contribution to the atmospheric turbidity in winter haze comes from a submicron aerosol fraction. It is also clear that in winter the coarse fraction of aerosol is mainly localized in a narrow size interval.

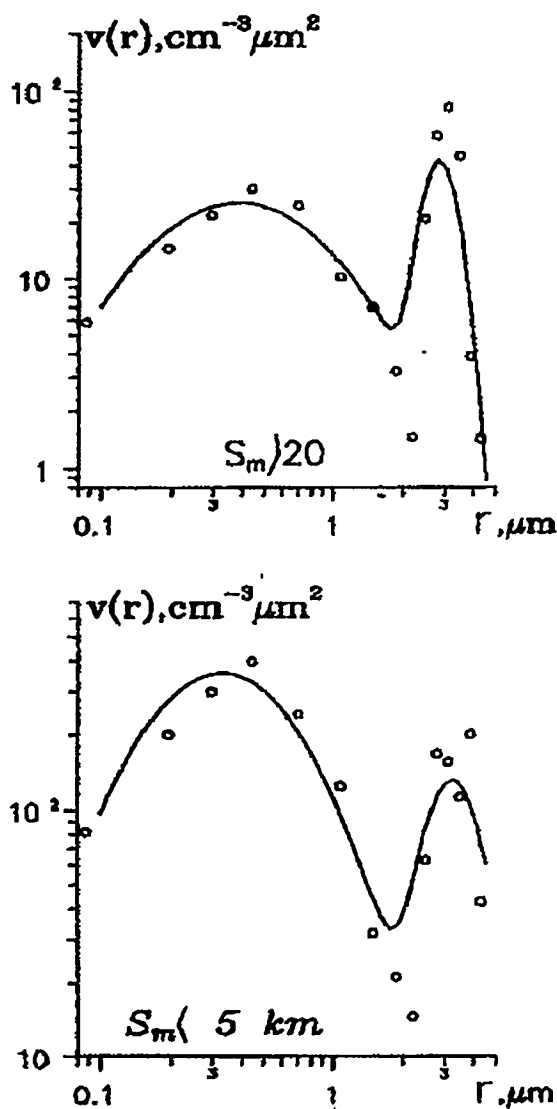


Figure 7. Size spectra of aerosol particles volume reconstructed by solving an inverse problem (circles) and their analytical approximations with superpositions of lognormal size-distributions (solid curves).

In 1994, we conducted a 3-week cycle of field measurements of the atmospheric spectral transmission in summer hazes. These measurements were carried out from June 16 till July 9 under low relative humidity conditions and low turbidity.

The mean spectrum of the aerosol extinction, $\alpha(\lambda)$, measured in a summer haze is shown in Figure 8 (Curve 1). The view of the spectrum presented here is indicative of the fact that the atmospheric turbidity at that time was mostly caused by background factors, which reveals a lack of fine aerosol fraction (the mean meteorological visual range, S_m , was about 23 km). In comparison, the mean spectrum $\alpha(\lambda)$ measured in 1992 in winter hazes is also shown in Figure 8 (Curve 2). These two curves quite clearly demonstrate the difference between the optical properties of the atmosphere observed in winter and in summer. The main difference is due to essentially different fine aerosol fractions in these seasons. This is in good agreement with the data presented in Figure 2.

An important part of the atmospheric aerosol studies concerns the study of its chemical composition. This information is useful both for identifying the sources of aerosol and for the determination of optical constants of the particulate matter necessary for making various calculations for the radiation models. Monthly mean characteristics of the aerosol chemical composition are shown in Figure 9, where all

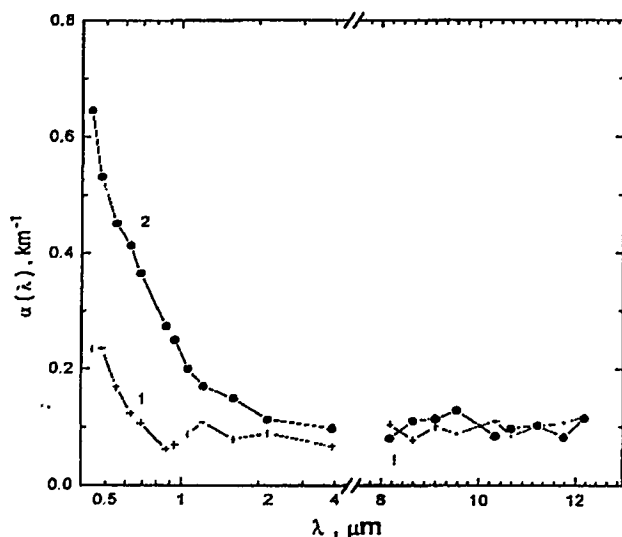


Figure 8. Spectral behavior of the mean aerosol extinction coefficients, measured in summer 1994 (1) and in winter 1992 (2).

the ions and elements identified are conditionally divided into three groups. The upper histogram shows those which are met in particles of natural origin. These components are usually called terragenic ones. The middle histogram presents the elements of anthropogenic and photochemical origin. The bottom histogram presents the microelements of both terragenic and anthropogenic origin. Normally the ratio of the total concentrations of the first group to that of the second one is practically constant over the whole Siberian region, and the total contribution coming from microelements to the chemical composition of aerosol makes only a few percent.

As shown in Figure 9, some changes in composition are observed during the change from spring to summer. Thus the concentration of Na^+ increased twice that which is indicative of more frequent arrival of marine air masses at the measurement site. Figure 9 also shows that the concentration of Al decreased by a factor of three, and the decrease of Ca concentration was even larger (about 40 times). At the same time, the concentration of Si and Mg increased almost ten times that which is indicative of difference in air masses coming there in July and April. In July an essential fall off of the anthropogenic components Cl^- , Br^- , NO^+ , and F^- was observed (if such components as Zn^{2+} , As^{5+} , Cd^{2+} , and Hg^{2+} , whose concentrations in our region are 10 to 100 times lower than the concentrations of other elements, are ignored). This situation may be explained by better dispersal of anthropogenic emissions in the summer compared to the winter. Concentration of microelements also decreased from April till July, except for Cr, B, and Mo.

The above results have been obtained in the ground atmospheric layer over Tomsk Akademgorodok and therefore possible contributions to the atmospheric aerosol from industrial emissions of Tomsk city must be remembered. This possibility is the subject of the next section.

In order to provide for separation of the above mentioned contribution from city of Tomsk, in 1993 we arranged simultaneous measurements at the Institute measurement site in a rural area (Kireevsk village) and in Akademgorodok, Tomsk. In this intercomparison experiment, we measured the same aerosol characteristics as our previous study. We also performed aerosol sampling in equal periods to provide a control of chemical composition of aerosol substance. Note that Kireevsk is 60 km to the West from Tomsk at the bank of the river Ob'. Since there are no big industrial centers near Kireevsk, it is quite reasonable to assume that changes in the physicochemical properties of the air

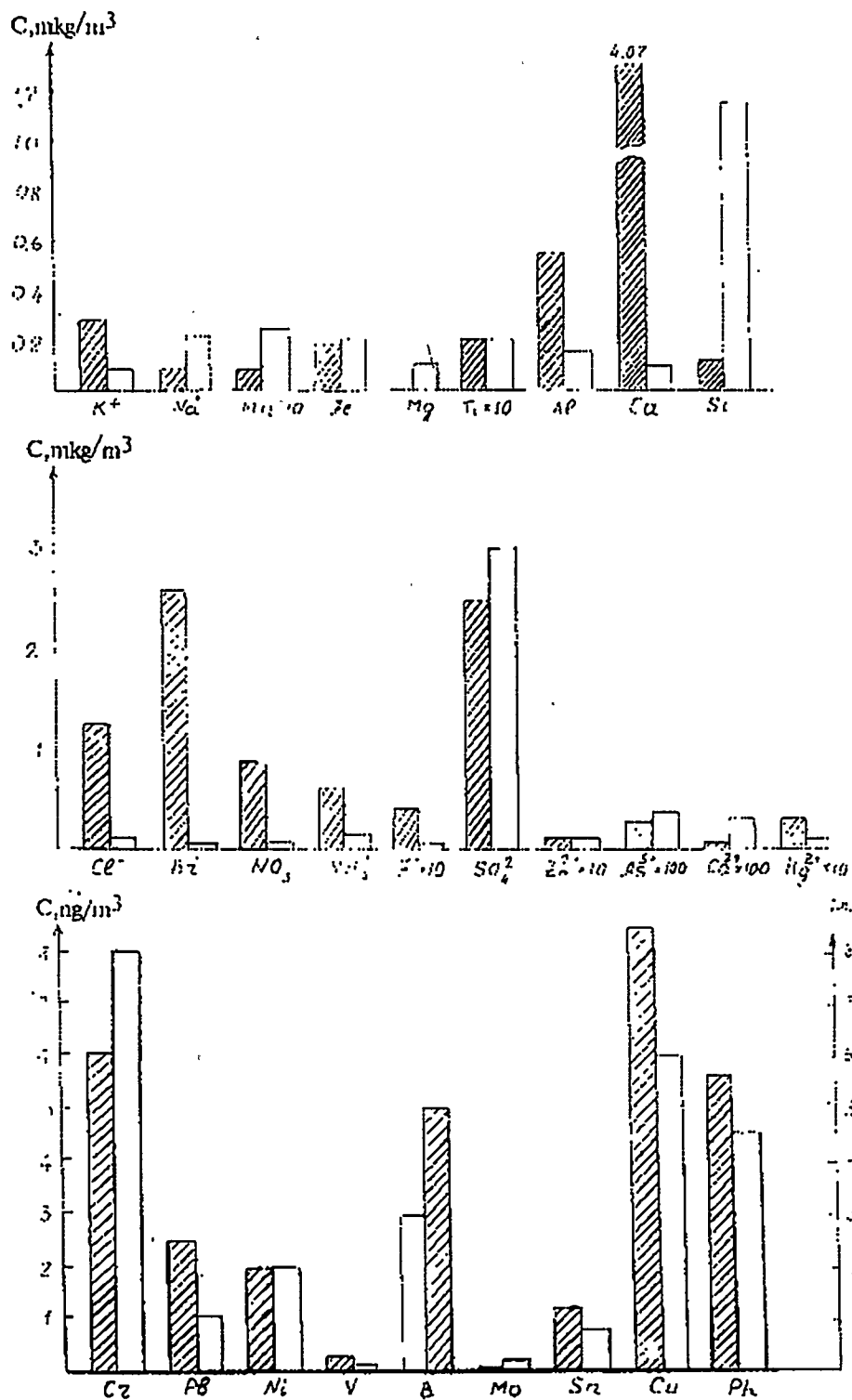


Figure 9. Monthly mean characteristics of the particulate matter chemical composition (shaded columns for April, clear columns for July).

mass coming from Kireevsk to Tomsk as detected in Tomsk could bear information on the contribution from city aerosol sources. The variations of the aerosol number density and of the coefficient of directed scattering $\mu_0(45^\circ)$ observed in Tomsk Akademgorodok and Kireevsk took place synchronously at both sites and had approximately the same amplitudes. It is important to note that measurements of $\mu_0(45^\circ)$ were performed with a nephelometer for dry air. The air coming to the nephelometric measurement cell was dried by heating in the pipe leading to the cell. Time behavior of the daily mean values of $\mu_0(45^\circ)$ measured in Tomsk and Kireevsk are plotted in Figure 10. Good agreement between the curves is indicative of the general character of the processes governing the aerosol variability at both points. These data also confirm the regional nature of the processes of formation and evolution of the main mass of the submicron aerosol in the ground atmospheric layer.

Aerosol samples collected in this study enable us to estimate in which elements the aerosol could be enriched on its way from Kireevsk to Tomsk Akademgorodok. Table 1 shows that the monthly mean number density of aerosol over Akademgorodok is a little bit (1.18 times) higher than in Kireevsk. For the majority of the elements identified, a certain, not random, enrichment can be seen. At the same time, some decrease in the concentrations can be seen for such elements as K^+ , As^{5+} , Si, Mo, and Cu. The concentration of Cl^- , Br^- , SO_4^{2-} , Zn^{2+} , Hg^{2+} , Ti, V, and Ca was constant or almost constant, which is indicative of the absence of sources of these elements on the way from Kireevsk to Tomsk. The marked increase of concentrations

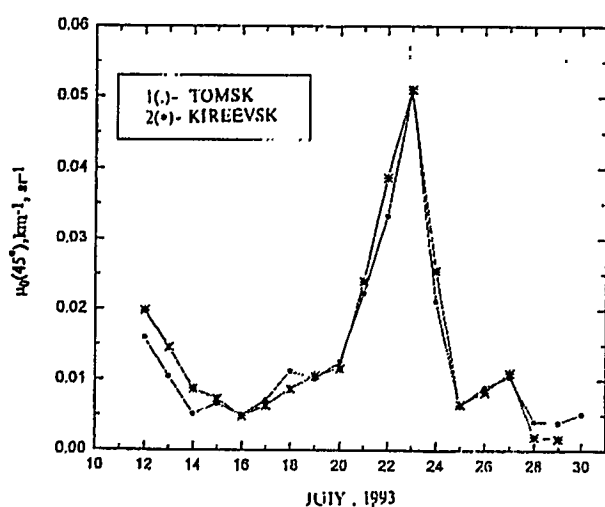


Figure 10. Time behavior of the daily mean values of $\mu_0(45^\circ)$ measured in Tomsk (Curve 1) and in Kireevsk (Curve 2).

Table 1. Chemical composition of aerosol (mkg/m³) in Kireevsk and Tomsk in July 1993.

Element	Tomsk	Kireevsk	Relationship
pH	4.64+0.63	4.89+0.59	0.95
K^+	0.08+0.05	0.24+0.07	0.33
Cl^-	0.01+0.003	0.01+0.01	1.00
Br^-	0.03+0.01	0.03+0.02	1.00
NO_3^-	0.005+0.03	0.02+0.01	2.50
NH_4^+	0.13+0.12	0.03	-
F	0.004	-	-
SO_4^{2-}	2.90+1.19	2.58+1.92	1.12
Zn^{2+}	0.01+0.006	0.009+0.006	1.11
Hg^{2+}	0.001+0.001	0.0008+0.0006	1.25
As^{5+}	0.003+0.003	0.008+0.007	0.38
Mn	0.025+0.01	0.015+0.010	1.67
Fe	0.2+0.14	0.13+0.08	1.54
Cr	0.008+0.001	0.005+0.001	1.60
Pb	0.001+0.001	0.0006+0.0003	1.67
Mg	0.11+0.03	0.08+0.04	1.34
Ni	0.002+0.001	0.001+0.0004	2.00
Ti	0.04+0.02	0.05+0.02	0.80
V	0.0001+0.0001	0.0001+0.0001	1.00
Al	0.15+0.06	0.10+0.03	1.50
B	0.005+0.004	0.0025+0.0020	2.00
Ca	0.09+0.03	0.10+0.06	0.90
Si	0.14+0.5	0.48+0.40	0.29
Mo	0.0001+0.0001	0.0002+0.0001	0.50
Sn	0.0008+0.0001	0.0005+0.0002	1.60
Cu	0.0006+0.0005	0.004+0.002	0.15
Σ	5.5926	3.9477	1.42

of NO_3^- , Ni, and B are evidently due to the influence of the city, though their contribution into the total aerosol mass is inessential.

Simultaneous measurements of the meteorological parameters of the atmosphere in Tomsk and Kireevsk showed their complete coincidence. The time behaviors of temperature and moisture recorded at both these measurement sites confirm the validity of the assumption that the same air mass travels across both sites.

The decrease in concentrations of K^+ , As^{5+} , Si, Mo, and Cu that occurred in the air mass on its way from Kireevsk to Tomsk could be connected with the varying underlying surface. Thus, near Kireevsk, river sand can be a source of Si. Since Si particles mainly contribute to the coarse fraction of aerosol, they rapidly descend to the surface and are therefore being removed from the air.

In summary, at least in July 1993, the city of Tomsk did not essentially contribute to the changes in the atmospheric aerosol composition. At the same time, simultaneous measurements of the meteorological quantities performed in Kireevsk and Akademgorodok showed, especially by their synchronous variations, that mainly the general circulation processes in the atmosphere were important at that time at both measurement sites.

Albedo and Transmittance of Inhomogeneous Stratus Clouds

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A highly important topic of recent concern has been the study of the relationship between the statistical parameters of optical and radiative characteristics of inhomogeneous stratus clouds. This relationship is important not only because it can be used to derive geophysical information (cloud spatial structure, within-cloud turbulence, etc.), but also because the radiation codes of general circulation models (GCMs) need improvement.

A cascade model has been created at the Goddard Space Flight Center (GSFC) to treat stratocumulus clouds with the simplest geometry (plane-parallel layer) and horizontal fluctuations of the liquid water path (optical thickness). These fluctuations were modeled by one-dimensional lognormal distribution and power-law spectrum (Cahalan 1989; Cahalan et al. 1994). It was shown that the area-average albedo depends essentially on the spatial distribution of optical thickness, and not only on its mean value.

Real stratus clouds have odd and irregular upper boundaries. Thus a question arises: How strongly does the stochastic geometry of clouds influence the statistical characteristics of albedo and transmittance of visible solar radiation? In the present work, this question is settled using the GSFC model extended to include the random upper boundary of the cloud layer.

Cloud Model

Current literature on stratus clouds contains no data on simultaneous measurements of the distributions of cloud liquid water and height of the cloud upper boundary, thus the relationship between microphysical and geometrical cloud characteristics has not been experimentally justified. For this reason, we have suggested that the cloud liquid water path and the height of the cloud upper boundary are independent random fields.

The statistical characteristics describing the fluctuations of the upper boundary are obtained from laser sensing of stratus clouds over the Barents and Norwegian seas in October-November 1987. The experimental data processed to date indicate that the stratus cloud upper boundary can be modeled to a first approximation by Gaussian homogeneous isotropic field with exponential correlation function (Andrianov et al. 1994). The correlation radius as determined at e^{-1} level is $2 \div 2.5$ km.

Below we use the following abbreviations:

- LWP model describes the fluctuations of liquid water path (optical thickness) in plane-parallel cloud layer. The LWP fluctuations have one-dimensional lognormal distribution and power-law spectrum.
- CUB model describes the fluctuations of the cloud layer upper boundary. The CUB fluctuations are assumed Gaussian process with exponential correlation function.
- LC model is the superposition of LWP and CUB models and describes both the fluctuations of liquid water content (extinction coefficient) and cloud layer upper boundary.

Notation used for statistical characteristics of optical depth (τ), albedo (R), transmittance (Q), etc., includes subscript (for a model), overbar (for the area average), and symbol Var (for a variance) (e.g., R_{LWP} and $\text{Var}R_{LWP}$ denote the area-average and the variance of albedo in LWP model).

For simplicity and to reduce computer time, we consider one-dimensional models, that is those in which the optical depth, albedo, and transmittance are random processes. The algorithm of simulating LWP, CUB, and LC realizations is based on the method of "spectrum randomization" (Mikhailov 1978, 1983) which consists in essence of the following:

- 1) LWP Model - Plane-parallel layer with thickness $\Delta H = 0.3$ km and length $L = dx \times N = 204.8$ km is divided into $N = 2^{12}$ pixels with equal horizontal sizes $dx = 0.05$ km. For each of the pixels, we simulate the optical thickness τ_{LWP} using one-dimensional lognormal distribution ($\tau_{LWP} = 13$, $\text{Var}\tau_{LWP} = 25.1$) and power-law spectrum (with exponent, $\beta = 5/3$). Each pixel, then, is assigned the extinction coefficient $\sigma(i) = \tau_{LWP}(i) / \Delta H$, $i = 1, N$.
- 2) CUB Model - Height of the upper boundary, $H(i)$, $i = 1, N$, is simulated with one-dimensional Gaussian distribution ($H = 0.3$ km, $\text{Var}H = 0.01$ km²) and exponential correlation function (correlation radius 2.5 km). The optical thickness of each pixel is given by

$$\tau_{CUB}(i) = \frac{\tau_{LWP}}{\Delta H} \times H(i) = \frac{13}{0.3} \times H(i), \quad i = 1, N \quad (1)$$
- 3) LC Model - Each pixel has the extinction coefficient $\sigma(i)$, thickness $H(i)$, and optical thickness $\tau_{LC}(i) = \sigma(i) \times H(i)$, $i = 1, N$. Owing to the independence of the random processes considered, $\tau_{LC} = \tau_{LWP} = 13$.

Examples of the one-dimensional distributions and energy spectra of optical depth are presented in Figure 1. The distribution mode in the LC model is shifted toward smaller optical thicknesses and is about 30% smaller, while the distribution itself is broader ($\text{Var}\tau_{LC} \approx 2 \times \text{Var}\tau_{LWP}$) (Figure 1a).

Figure 1b is a log plot of the energy spectrum of optical thickness, $E(\tau, k)$, versus bin number k , which is related to the spatial frequency f_k as $f_k = k/L = k/(dx \times N)$, with L the length of the realization processed. The smoothing of energy spectrum was performed by sliding summation over 59 points of nonsmoothed spectrum. In the spatial frequency range under consideration, the energy spectrum $E(\tau_{CUB}, k)$ has a near flat (on a log plot) section, which can be approximated by $E(\tau_{CUB}, k) \sim k^{-\beta_{CUB}}$, with $\beta_{CUB} = 2.0$; that is, $\beta_{CUB} > \beta_{LWP}$ and $E(\tau_{CUB}, k)$ falls off faster than $E(\tau_{LWP}, k)$. Since the variance of τ_{LWP} is 1.5 times that of τ_{CUB} , the energy spectrum $E(\tau_{LC}, k)$ differs little from $E(\tau_{LWP}, k)$.

Albedo and Transmittance

Albedo and transmittance are computed in each pixel by the Monte Carlo method, employing, in particular, the Maximal Cross-Section Method (Marchuk et al. 1976). The calculations are made for overhead sun and Henye-Greenstein

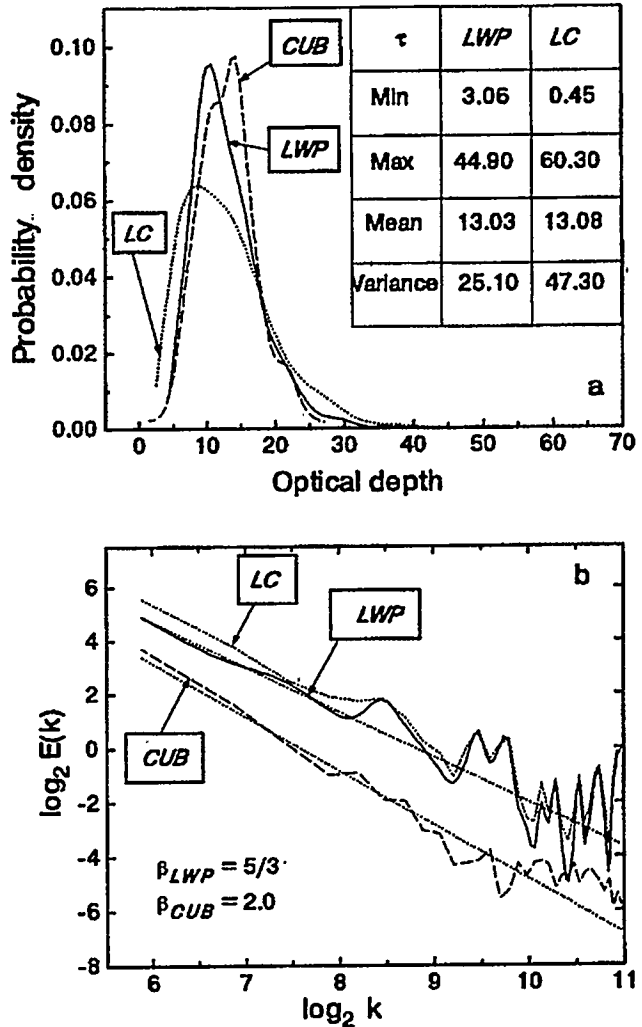


Figure 1. Probability density (a) and energy spectrum (b) of optical depth.

scattering phase function with an asymmetry parameter of 0.843. Impact of the underlying surface and aerosol atmosphere was not considered. In order to evaluate the extremely possible impact of the cloud top stochasticity on solar radiation transfer in inhomogeneous stratus clouds, R_{LC} and Q_{LC} were calculated with $\text{Var} H = (H/3)^2 = 0.01$ km². Minimum and maximum ($H_{\max}$) cloud top heights were, respectively, 0.014 and 0.646 km. The albedo R_{LWP} was determined at height $z = \Delta H$, while R_{LC} was at the maximum height of the cloud upper boundary (plane $z = H_{\max}$). The transmittances Q_{LWP} and Q_{LC} were calculated at the level of the cloud lower boundary (plane $z = 0$). The relative computation error was within 1%.

Below we present the calculation results which illustrate the influence of the effects, caused by the stochastic geometry of stratus cloud upper boundary, upon the statistical characteristics of albedo and transmittance. We note that these effects, missing completely in the LWP model, are most important for cumulus clouds and are sufficiently discussed in Titov (1990) and Zuev and Titov (1995).

The random upper boundary of stratus clouds produces a considerable increase in the variance of optical depth, so that $\text{Var}R_{LC} \approx 1.7 \times \text{Var}R_{LWP}$ and $\text{Var}Q_{LC} \approx 2 \times \text{Var}Q_{LWP}$. The mean albedo and transmittance, however, are almost unaffected ($\sim 3\%$ change). The probability density of R_{LC} , unlike that of R_{LWP} , is distinctly bimodal (Figure 2a), which might be explained by the increase of the variance of optical depth. To verify this suggestion, we have calculated the albedo and transmittance for the LWP model with $\text{Var}\tau_{LWP} = \text{Var}\tau_{CUB}$.

The results are tabulated in the figures in parentheses (Figure 2) and show that, for fixed values of the mean and variance of optical depth, the mean and variance of albedo and transmittance depend on the cloud model only weakly. For such variations of the variance of optical depth, the probability density of R_{LWP} is not distinctly bimodal. This means that the effects caused by the stochastic cloud geometry are responsible for the bimodal behavior of the probability density of R_{LC} .

As is well known, there is a "scaling break" in the energy spectrum of reflected intensity (Barker and Davies 1992) and albedo (Cahalan and Snider 1989) on the space scales of several hundred meters. Our results support this fact and show that the energy spectrum of transmittance is "broken" as well (Figure 3). The explanation is that the multiple scattering significantly smoothes out the radiation field at space scales on the order of ten photon mean-free pathlengths.

The difference is appreciable between energy spectra of albedo, $E(R_{LC}, k)$ and $E(R_{LWP}, k)$, while being negligible between $E(Q_{LWP}, k)$ and $E(Q_{LC}, k)$. The former is attributable to the fact that the radiation field reflected by a given pixel "spreads" in the space before reaching the plane $z = H_{\max}$ to give albedo; in addition, it overlaps with radiation fields reflected by neighboring pixels. The two effects are 1) the stronger the energy spectra of albedo, the larger the difference between $H_{\max}$ and the cell height, and 2) both act to smooth out the spatial distribution of albedo.

In summary, the results presented above show that, for fixed values of the mean and variance of optical depth, the effect of the stochastic geometry of the upper boundary of inhomogeneous stratus clouds upon the mean and the variance of

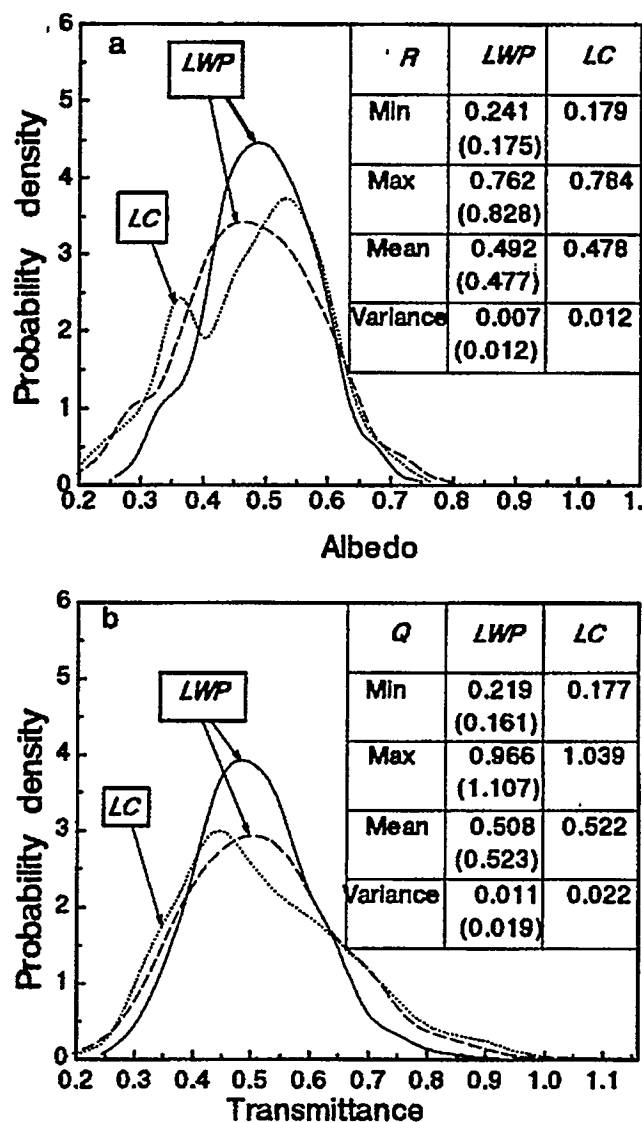


Figure 2. Probability density of albedo (a) and transmittance (b). Solid line shows $\text{Var}\tau_{LWP} = 25.1$, dashed line shows $\text{Var}\tau_{LWP} = 47.3$.

albedo and transmittance can be neglected. The probability density and energy spectrum of albedo are highly sensitive to the variations of the stratus cloud upper boundary.

Current literature on stratus clouds contains no data on the relationship between the statistical characteristics of cloud liquid water and cloud top altitude. For this reason, we suggested that the cloud liquid water path and the height of cloud upper boundary are independent random processes. To develop cloud models which most fully and adequately describe the real inhomogeneous stratus clouds, and to

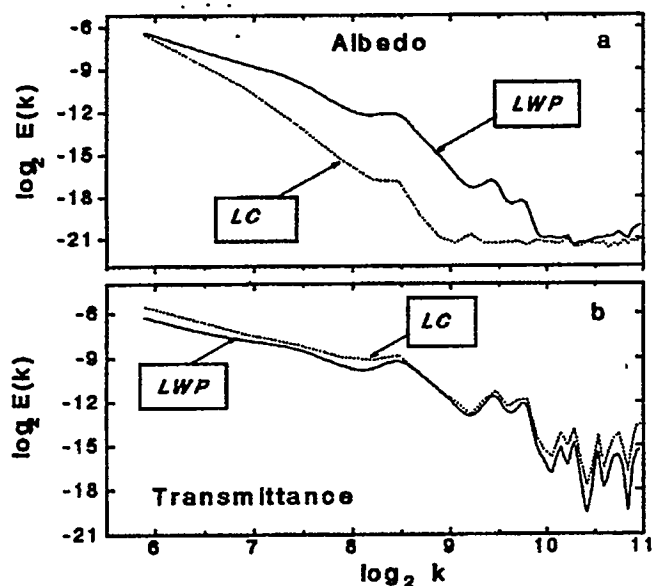


Figure 3. Energy spectrum of albedo (a) and transmittance (b).

elaborate the appropriate solution techniques, will require data on the statistical characteristics of the cloud microphysical and geometrical parameters and the relationship between them. It is hoped that such data will be available from the Atmospheric Radiation Measurement (ARM) Program.

Acknowledgment

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Simulation of Solar Radiative Transfer in Cumulus Clouds

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Current radiation codes of general circulation models (GCMs) are still largely based on plane-parallel models, which unsatisfactorily describes the radiative effects of cumulus clouds because their optical characteristics are highly variable both vertically and horizontally. Updated GCM parameterizations of the interaction of radiation with cumulus clouds will use newer Atmospheric Radiation Measurement (ARM) data and more realistic radiative transfer models. In this work we present a 3-D model of radiative transfer which is used to study the relationship between the spatial distribution of cumulus clouds and fluxes (albedo and transmittance) of visible solar radiation.

In the visible spectrum, the scattering by water droplets can be assumed conservative, and the mean albedo $\bar{R}$ and transmittance $\bar{Q}$ are related by formula $\bar{R} + \bar{Q} = 1$, expressing the law of energy conservation. We let the clouds occupy some spatial domain shaped as parallel piped with thickness ΔH and with a square base of side length Xl . In the horizontal plane we divide the domain into cells (pixels) of equal size Δl , and for each of these calculate the albedo R_{ij} and transmittance Q_{ij} , $i, j = 1, \dots, nx$ where $nx = \frac{Xl}{\Delta l}$ is the number of cells along each of the coordinate axes OX and OY . The total number of cells is nx^2 . If the domain is a part of a plane-parallel cloud, then, on account of the homogeneous boundary conditions, the equality $Q_{ij} + R_{ij} = 1$, $i, j = 1, \dots, nx$, holds in each of the cells. Will the same equality hold in the cumulus cloud case? What consequences will the horizontal inhomogeneity of cumulus clouds have on the spatial distribution of uniform incident solar flux? The mathematical simulation results we present below answer somewhat these questions, as well as make it possible to obtain the probability densities and energy spectra of albedo and transmittance in statistically inhomogeneous cumulus clouds.

Model of Cumulus Clouds and Method of Solution

The stochastic geometry of cumulus clouds is, to date, poorly understood. To avoid expensive computations, we use a simple model of cumulus clouds generated by Poisson

point fluxes in space. Cumulus clouds are approximated by inverted, truncated paraboloids of revolution with height H equal to diameter D . The latter has the exponential distribution function $f(D) \sim \exp(-\alpha D)$, $D_{\min} \leq D \leq D_{\max}$. Optical parameters (extinction coefficient, single scattering albedo, and scattering phase function) are assumed constant within the cloud. The radiation effects of atmospheric aerosol and underlying surface are neglected for simplicity.

A computer realization of cloud field for $\Delta l = 0.1$ km, $nx = 64$ ($nx^2 = 4096$ and $Xl = 6.4$ km), $\alpha = 2$, $D_{\min} = 0.03$ km, $D_{\max} = 1.2$ km, and for a cloud fraction of $N = 0.5$ is presented in Figure 1a. If a cell belongs to a group of (two or more) clouds, its optical thickness is calculated for the largest cloud in the group. In this realization, maximum cloud thickness is $\Delta H = 1.174$ km.

The 3-D transfer equation was solved subject to cyclical boundary conditions assuming that the layer $0 \leq z \leq \Delta H$ is filled by a given cloud field realization. The 2-D albedo and transmittance fields were computed by Monte Carlo (MC) method using one or more hundred million trajectories that ensured no more than 2% relative error. The transmittance was determined at the lower (plane $z = 0$) and the albedo at the upper (plane $z = \Delta H$) cloud boundary. Details of the cumulus cloud model and the MC algorithms are given in Titov (1990) and Zuev and Titov (1995).

2-D Albedo and Transmittance Fields

All calculations below employ Henyey-Greenstein scattering phase function with an asymmetry factor of 0.843 (typical of wavelengths 0.3 - 3.0 μm). The azimuthal solar angle measured from OX -axis is zero.

Multiple scattering plays a dominant role in the formation of radiation field in clouds, leading, in particular, to the fact that the albedo R and diffuse transmittance Q , reach appreciable values even in noncloudy cells (Figures 1b and 1d). In the given cloud field realization, clouds vary in thickness from 0.033 km to 1.174 km. Radiative field,

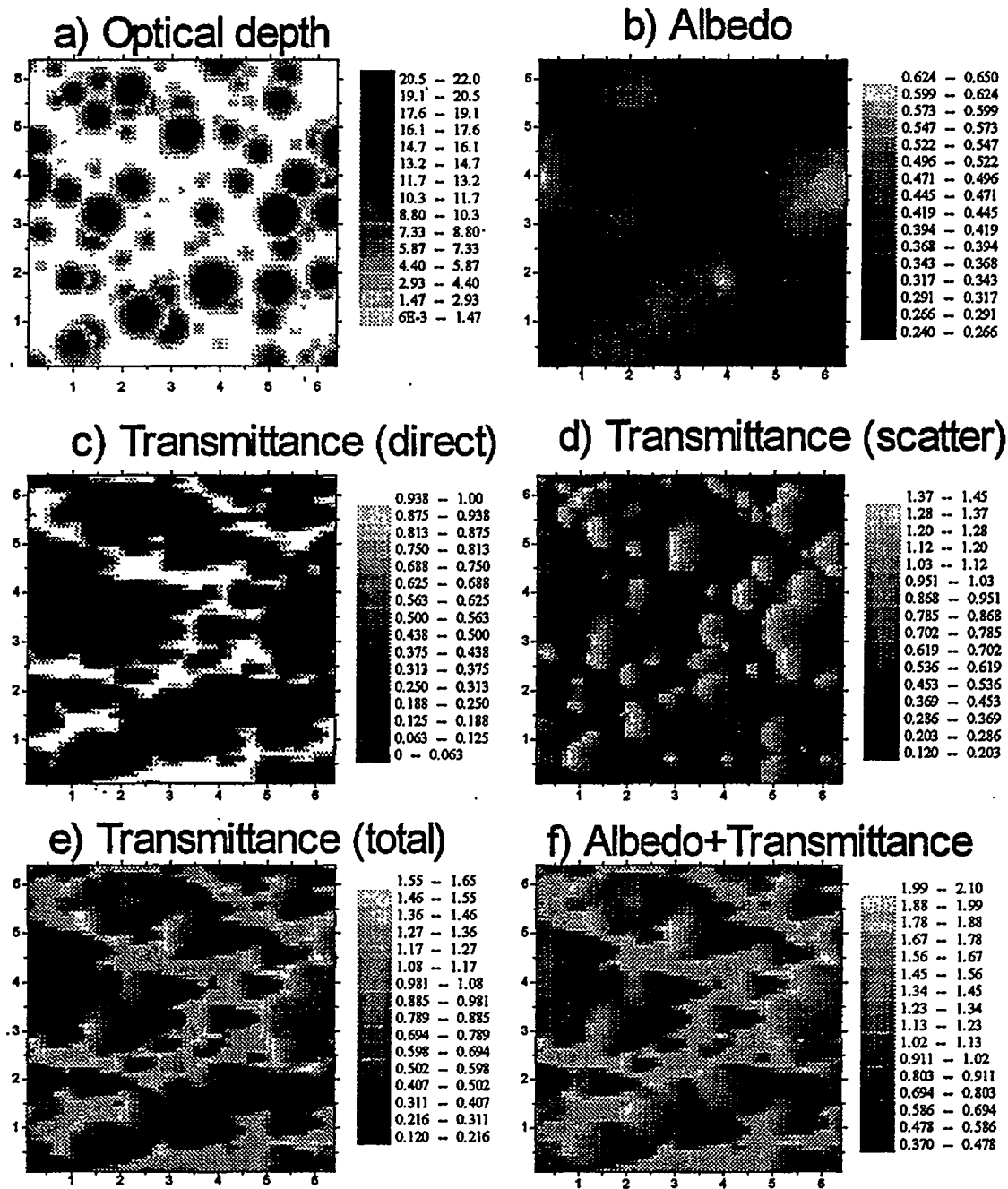


Figure 1. 2-D fields of optical depth (a), albedo (b), transmittance of direct (c), diffuse (d), and total radiation (e), and albedo plus transmittance (f), for the cloud fraction $N = 0.5$, pixel horizontal size of 0.1 km, extinction coefficient $\sigma = 20 \text{ km}^{-1}$, and solar zenith angle $\xi = 60^\circ$.

reflected by a single cloud, spreads in the space and overlaps with the radiative fields of other clouds, before reaching the plane $z = \Delta H$ to give albedo. Owing to the spread and overlap effects, the albedo is essentially smoothed out in the horizontal plane, so that the many details are effectively masked, thus complicating the visual reconstruction of the real pattern of cloud spatial distribution from known albedo values (Figures 1a and 1b). The tops of the most massive clouds are distinctly observed because for these, the effects above are not sufficiently strong. The albedo varies from 0.24 to 0.65, the mean and the variance being 0.33 and 0.0039, respectively.

The transmittance S of direct radiation passing through holes between clouds is unity (the direction of the sun is cloud-free); and at oblique sun angles, the cloud shadows are clearly seen (Figure 1c). Optically thin cells, localized near cloud bottom and directly sunlit, are able to transmit much radiation without scatter, smearing somewhat shadow boundaries. The diffuse transmittance Q_s is maximum for those cells which have small (roughly 2-3) optical thickness and are not screened by surrounding clouds (i.e., which are reached by the incident solar radiation without interaction with the clouds [Figure 1d]). Interestingly, Q_s for such cells can be markedly greater than unity, owing to the radiation striking the neighboring, optically thicker cells. However, through multiple scattering, the Q_s "slides down" to these optically thinner cells, thus adding to their transmittances. Certainly, there exist cells with small (0.1-0.2) Q_s values, which are either far apart from the clouds or fall within the shadow zones of the most massive of them. As a result, the Q_s may vary by more than order of magnitude (from 0.12 to 1.45) with the average being 0.42.

The transmittance of total radiation, $Q = S + Q_s$, is given in Figure 1e and varies from 0.12 to 1.65, with the mean being 0.67 and variance being 0.1839. The smallest Q values occur in the cells shaded by most massive cumulus clouds. For cells located in cloud gaps $S = 1$ and, due to the diffuse contribution, the inequality $Q > 1$ always holds. Clearly, this inequality will also hold for those cells where $Q_s > 1$. The $Q > 1$ values have long been obtained by field measurements (Timanovskaya and Feigelson 1970); however, no such theoretical estimates of Q were previously reported, as far as we know.

The simulation results indicate that the albedo and transmittance of cumulus have large horizontal gradients. As a result, the law of energy conservation holds only on averages, for the whole spatial domain under consideration,

but in each cell the sum $Q_{ij} + R_{ij}$, $i, j = 1, \dots, nx$, may substantially differ from unity, ranging from 0.37 to 2.10 (Figure 1e). The horizontal inhomogeneity of radiant fluxes in cumulus can affect appreciably such atmospheric processes as underlying surface heating, cloud dynamics, and photochemical reactions, among many others.

Independent Pixel Approximation (IPA)

This approximation neglects the net horizontal photon transport and uses the plane-parallel model for each pixel (cell) (Cahalan 1989; Cahalan et al. 1994). In other words, radiative properties of each cloud pixel depend only on its own vertical optical thickness, not on the optical thickness of neighboring pixels. In the case of plane-parallel stratocumulus clouds with nonuniform distribution of optical depth, the IPA has reasonable accuracy if only the horizontal size of the pixel exceeds its thickness and is much larger than photon mean free paths in clouds. Otherwise, mean-albedo biases of up to 10-20% result. In the IPA, the albedo R_{IPA} and transmittance Q_{IPA} can be calculated for each pixel using the approximation derived by King and Harshvardhan (1986) for conservative scattering:

$$\begin{aligned} R_{IPA}(\tau; \xi, g) &= 1 - Q_{IPA}(\tau; \xi, g), \\ Q_{IPA}(\tau; \theta, g) &= \frac{\delta(\xi) + [1 - \delta(\xi)] \exp[-\tau / \alpha(\xi)]}{1 + \gamma(g)\tau} \end{aligned} \quad (1)$$

where τ is the pixel optical thickness, ξ is the solar zenith angle, and g is the asymmetry parameter. Below we use the following values of the functions $\delta(\xi)$, $\alpha(\xi)$, $\gamma(g)$: $\delta(60^\circ) = 0.8$, $\alpha(60^\circ) = 0.8$ and $\gamma(0.43) = 0.11$.

Much larger albedo biases are expected for cumulus clouds whose geometries are fundamentally different from plane-parallel because the effects associated with their stochastic geometry (shading and multiple scattering between clouds) profoundly influence the solar radiative transfer. From (1), the 2-D fields R_{IPA} and Q_{IPA} follow the distribution of optical thickness (Figure 1a) and, therefore, visually they are quite dissimilar to their MC counterparts, R and Q fields (Figures 1b and 1e). In particular, R_{IPA} and $Q_{s, IPA}$ have zero values in the pixels located in gaps (holes) between clouds, and Q_{IPA} never exceeds 1. In addition, IPA fails to describe the shading effects.

The MC versus IPA comparison in cumulus is performed by calculating

$$\Delta R = \frac{R - R_{IPA}}{R} \times 100\%$$

$$\Delta Q = \frac{Q - Q_{IPA}}{Q} \times 100\% \quad (2)$$

The IPA underestimates albedo for pixels located in gaps (holes) between clouds, and vice versa, for the other pixels (Figure 2a). The IPA-calculated mean and variance of albedo are, respectively, 0.25 and 0.0807. Compared with the MC values above, the IPA gives about a factor of 1.5 lower mean albedo and 20 times higher variance. This implies that the radiative interaction of pixels and the effects of stochastic geometry effectively smooth out the reflected field of solar radiation. The transmittance Q exceeds Q_{IPA} in the pixels located in the optically thin cells near cloud base or in cloud gaps, both directly sunlit (i.e., not shaded by clouds [Figure 2b]). As expected, $|\Delta Q|$ assumes maximum values in the pixels located in cloud gaps and shaded by clouds. Values of Q may be far in excess of 1, so the variance of Q is two times that of Q_{IPA} .

Summarizing, the results above clearly demonstrate that IPA unsatisfactorily describes the radiation effects of cumulus clouds.

Currently the lack of experimental data on stochastic cumulus clouds represents a major obstacle for developing 3-D radiative transfer models for cumulus clouds. Hopefully, more data will come from the ARM Program, which will help to improve the existing models and, if necessary, to develop newer, more realistic ones for treating the interaction of solar and thermal radiation with cumulus clouds.

Acknowledgment

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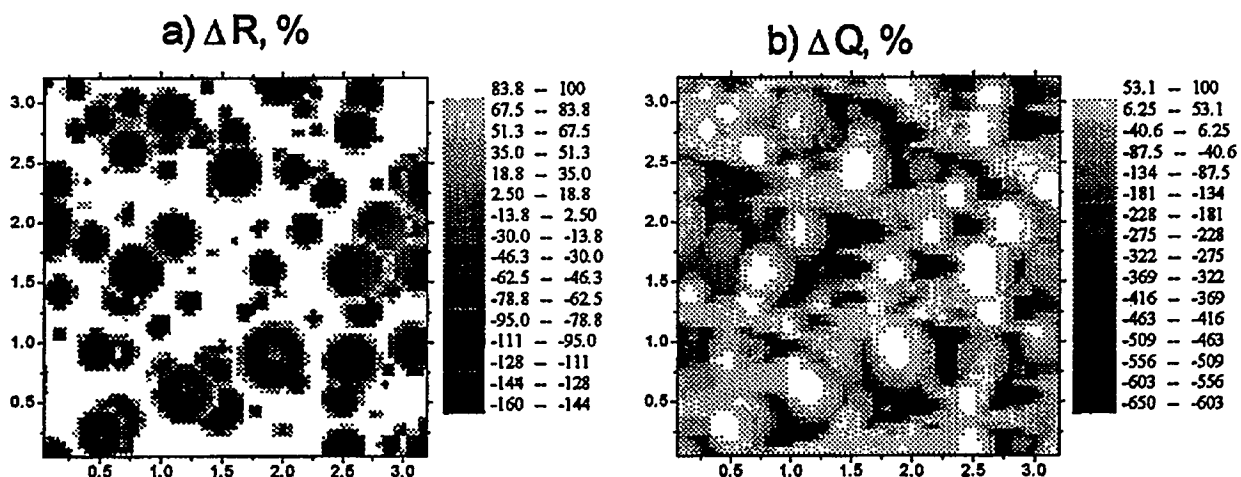


Figure 2. 2-D fields ΔR and ΔQ for cloud fraction $N = 0.5$, pixel horizontal size of 0.1 km, extinction coefficient $\sigma = 20 \text{ km}^{-1}$, and solar zenith angle $\xi = 60^\circ$.

Absorption of Solar Radiation in Broken Clouds

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Introduction

It is recognized now that the plane-parallel model unsatisfactorily describes the transfer of radiation through broken clouds and that, consequently, the radiation codes of general circulation models (GCMs) must be refined. However, before any refinement in a GCM code is made, it is necessary to investigate the dependence of radiative characteristics on the effects caused by the random geometry of cloud fields. Such studies for mean fluxes of downwelling and upwelling solar radiation in the visible and near-infrared (IR) spectral range were performed by Zuev et al. (1994a,b).

In this work, we investigate the mean spectral and integrated absorption of solar radiation by broken clouds (in what follows, the term "mean" will be implied but not used, for convenience). To evaluate the potential effect of stochastic geometry, we will compare the absorption by cumulus ($0.5 \leq \gamma \leq 2$) to that by equivalent stratus ($\gamma \ll 1$) clouds; here $\gamma = H/D$, H is the cloud layer thickness and D the characteristic horizontal cloud size. The equivalent stratus clouds differ from cumulus only in the aspect ratio γ , all the other parameters coinciding. The spectral absorption $P(\lambda)$ of stratus clouds partially covering the sky is calculated with high accuracy by the formula

$$P(\lambda) = N \times P_{\text{lay}}(\lambda) + (1-N) \times P_{\text{cle}}(\lambda) \quad (1)$$

where N is the cloud fraction, and $P_{\text{cle}}(\lambda)$ and $P_{\text{lay}}(\lambda)$ are the spectral absorptions by clear sky and plane-parallel overcast, respectively. Obviously, the same formula applies to the calculation of integrated absorption as well.

Models of the atmosphere and the methods of spectral flux calculations in the near-IR have been sufficiently discussed by Zuev et al. (1994a,b) and Titov et al. (1995). It should only be recalled that the atmosphere is assumed to extend vertically from 0 to 16 km and that the clouds are contained in the 1-1.5 km layer; also that only the absorption by water vapor and carbon dioxide is considered. The spectral absorption is calculated by assuming that the unit solar flux is incident on the top of the atmosphere in direction $\vec{\omega}_0 = (\xi_0, \varphi_0)$, with ξ_0 and φ_0 the zenith and azimuthal solar angles. To change to the absolute values, results for spectral

absorption must be multiplied by $\pi S_\lambda \cos \xi_0$, where πS_λ is the spectral solar constant.

Since the optical thickness of atmospheric aerosol is considered to be much smaller than that of the clouds, any absorption difference between cumulus and stratus clouds will primarily result from multiple scattering in clouds. So we can restrict ourselves to the discussion of absorption by the cloud layer alone. In addition, owing to the small optical thickness of above-cloud atmosphere, the scattering of radiation reflected from the cloud layer can be neglected. This implies that the incident solar radiation on the cloud top is independent of the cloud type.

Spectral Absorption

As is well known, the absorption by cloud particles (water droplets) increases with the increase of the fraction of diffuse radiation and the mean multiplicity of scattering, as well as with the decrease of the single scattering albedo. From the above definition of equivalent stratus clouds, it is obvious that only the first two of these factors depend on cloud type. For fixed pressure, temperature, and concentration, the absorption by atmospheric gases is determined by the value of photon mean free path in clouds. We begin the discussion of calculation results with the simplest case when the albedo of underlying surface, A_s , is zero.

Radiation may escape through the sides of cumulus clouds, so that the mean multiplicity of scattering and the photon mean free path in cumulus are less than in stratus. When the sun is in zenith, the fraction of diffuse radiation is the same for both cloud types, thus the absorption in stratus, $P_{\text{st}}(\lambda)$, exceeds the absorption in cumulus, $P_{\text{cu}}(\lambda)$ (Figure 1).

Except for large zenith solar angles ($\xi_0 > 80^\circ$), the fraction of diffuse radiation in stratus depends very weakly on ξ_0 , whereas in cumulus it rapidly grows with increasing ξ_0 . For cumulus, this leads to an increase in absorption, which may compensate for the effect caused by the decrease of the mean multiplicity of scattering and photon mean free path.

Thus, there are two opposite effects whose net result can be determined by expanding the spectral absorption in series in scattering multiplicities (orders). Obviously, the n th term of

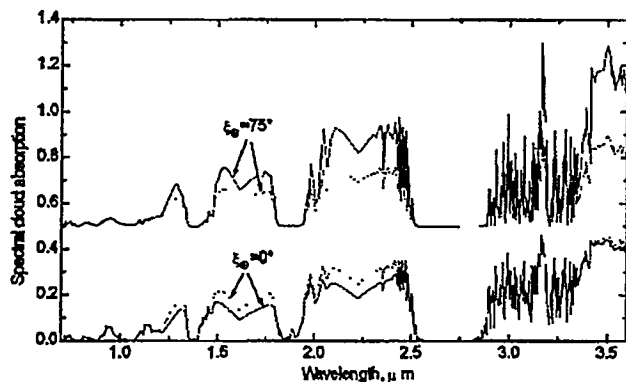


Figure 1. The dependence of spectral absorption in cumulus (solid) and stratus (dots) on solar zenith angle with $N = 0.5$, $\sigma_{0.71 \mu m} = 30 \text{ km}^{-1}$, $D = 0.25 \text{ km}$, $A_s = 0.0$. For convenience, values of cloud absorption for $\xi_0 = 75^\circ$ are exaggerated by 0.5.

the expansion is proportional to $w_\lambda^{n-1}(1-w_\lambda)$. When the single scattering albedo approaches unity, a major contribution to the spectral absorption will come from high scattering orders. Low scattering orders are more important contributors to the absorption by cumulus, while higher orders are of more importance in stratus. At wavelengths $\lambda \leq 1.2 \mu m$, water droplets absorb weakly ($w_p \geq 0.999$). For such w_λ values, the increasing diffuse fraction may not compensate for the effect of decreasing mean scattering order and photon mean free path; as a result, for $\xi_0 > 0^\circ$, the spectral absorption in cumulus is less than that in stratus (Figure 1). The situation is reverse for $\lambda \geq 1.2 \mu m$, in which case the water droplets strongly absorb ($0.485 \leq w_\lambda \leq 0.999$).

The differences in spectral absorption between stratus and cumulus clouds (the other parameters coinciding) have been discussed above. Now let us discuss the dependence of spectral absorption on solar zenith angle, for each of the cloud types. To this end we modify Figure 1 somewhat to obtain Figure 2. As zenith solar angle grows, the incident solar radiation upon the cloud top is increasingly depleted by the gaseous absorption and aerosol extinction; in addition, the cloud albedo increases, and the mean scattering order and mean path length of reflected photons both decrease. For these reasons, the absorption by stratus clouds decreases as ξ_0 increases. A more complicated pattern is observed in the cumulus clouds case. For a moderate gaseous absorption, the increasing fraction of diffuse radiation dominates over the other effects mentioned above, so that the cumulus cloud absorption increases. For a strongly absorptive case, the dependence is reversed.

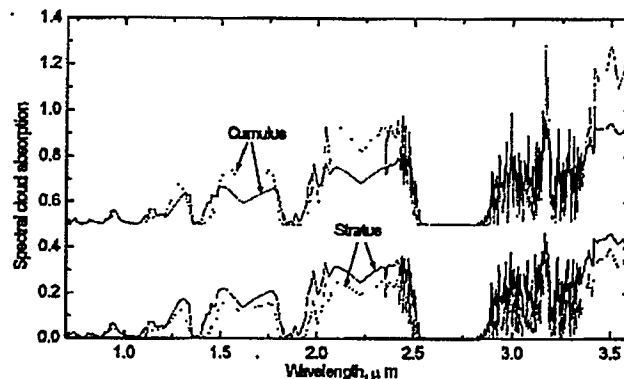


Figure 2. Same as in Figure 1 but for spectral cloud absorption with $\xi_0 = 0^\circ$ (solid) and $\xi_0 = 75^\circ$ (dots). For convenience, values of cumulus cloud absorption are exaggerated by 0.5.

Let us suppose now that the underlying surface reflects according to Lambert's law and has the albedo $A_s > 0$. This surface can be viewed as a diffuse source whose power is proportional to $A_s Q(\lambda)$, with $Q(\lambda)$ the spectral transmittance at the surface level. Obviously, only in the case of a weak to moderate droplet and gaseous absorption will the source be powerful enough to affect noticeably the spectral absorption by clouds. Radiation reflected from the surface can interact with the sides of numerous cumulus clouds (i.e., may be scattered and absorbed). Quantitatively, this means that the fraction of radiation passing in gaps (holes) between clouds will be much less in cumulus than in stratus clouds. As a consequence, cumulus clouds absorb more radiation reflected from the surface than the stratus cloud do.

Integrated Absorption

By integrated absorption we mean the quantity

$$P = \frac{\int_{0.7 \mu m}^{3.6 \mu m} \pi S_\lambda P(\lambda) \cos \xi_0 d\lambda}{\pi S \cos \xi_0} \times 100\%, \quad (2)$$

where $\pi S = 1353 \text{ W/m}^2$ is the integrated solar constant. Major contributors to the integrated absorption are those spectral intervals in which $P(\lambda)$ and πS_λ are sufficiently large. Clearly, the behavior of integrated absorption is

dominated by the dependence of spectral absorption on problem parameters, so some obvious cases will be given without comments. Recall that the wavelength interval $0.7-1.2 \mu\text{m}$ comprises $\approx 30\%$ of the incident solar flux on the top of the atmosphere, while for the $1.2-2 \mu\text{m}$ and $2-4 \mu\text{m}$ intervals the percentages are $\approx 15\%$ and 5% , respectively.

As the solar zenith angle increases, the integrated cloud absorption P_{Cloud} decreases for stratus and slightly increases for cumulus (Figure 3). The latter is because the spectral absorption in cumulus, $P_{\text{Cu}}(\lambda)$, may both increase and decrease with growing ξ_0 (Figure 2); thus P_{Cloud} consists of two contributions varying in opposite directions with ξ_0 . For $\lambda \leq 2 \mu\text{m}$, and as ξ_0 grows, there are spectral intervals in which $P_{\text{Cu}}(\lambda)$ decreases, and πS_λ is sufficiently large there, so these intervals are important contributors, and P_{Cloud} increases little. At the same time, the lower the solar elevation, the higher the integrated absorption due to the above-cloud atmosphere, so the larger the total atmospheric integrated absorption in both cumulus and stratus cases.

The effects associated with finite horizontal extents of cumulus clouds are responsible for the nonlinear (in contrast to stratus case) dependence of integrated absorption on

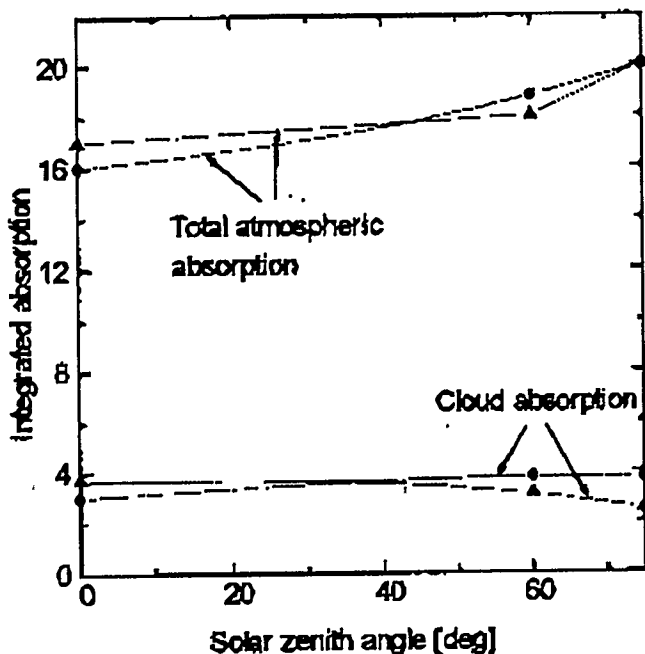


Figure 3. The influence of zenith solar angle on integrated absorption by cumulus (circles) and stratus (triangles) with $\sigma_{0.71 \mu\text{m}} = 30 \text{ km}^{-1}$, $N = 0.5$, $D = 0.25 \text{ km}$, $A_s = 0.0$.

cloud fraction N (Figure 4). Maximum differences in integrated absorption between cumulus and stratus clouds occur at intermediate cloud fractions and grow with increasing surface albedo.

In summary, the results above clearly demonstrate the substantial influence of effects, caused by the stochastic geometry of cloud fields, upon the spectral and integrated absorption of solar radiation in cumulus clouds. Currently we possess an extensive body of results for upward and downward fluxes of solar radiation, which can be viewed as a numerical radiation model of broken clouds. The next step of studies must be directed toward the improvement of now existing GCM radiation codes. This can be done through development of simple techniques for calculating the radiant fluxes in broken clouds, which would be sufficiently efficient and could adequately treat the effects caused by the stochastic geometry of cloud fields. Their accuracy and applicability range can be assessed by means of the developed numerical radiation model of broken clouds.

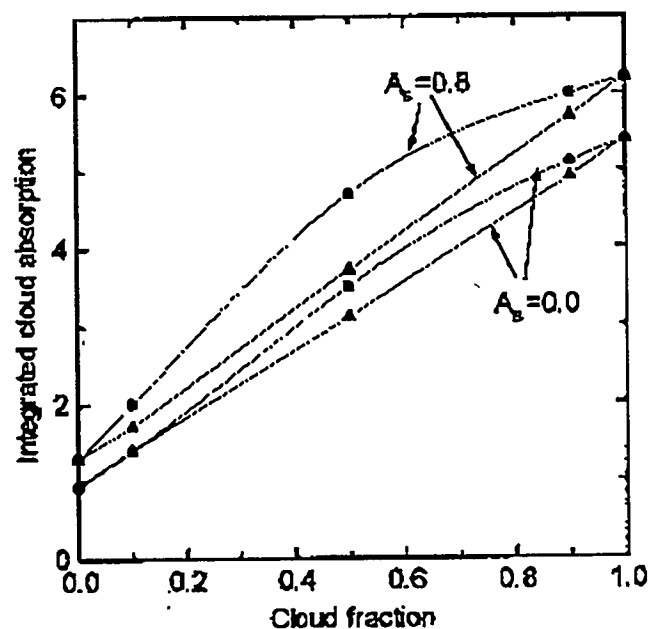


Figure 4. The dependence of the integrated absorption in cumulus (circles) and stratus (triangles) on cloud fraction with $\sigma_{0.71 \mu\text{m}} = 30 \text{ km}^{-1}$, $D = 1.0 \text{ km}$, $\xi_0 = 60^\circ$ and for different values of surface albedo A_s .

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Author Index

Author Index

- Ackerman, T.P. 1, 173, 183, 185, 187, 273, 283
Albert, B. 1, 155, 159
Albrecht, B. 1
Alexander, G.D. 5
Aliss, R.J. 9
Alzheimer, J.M. 287
Anderson, G.A. 287
Anderson, G.P. 293
Arsky, N. 13
Baer, F. 13
Barnett, T.P. 263
Barr-Kumarakulasinghe, S.A. 19
Belan, B.D. 241
Berg, L. 327
Berkheiser, W. 205
Bernstein, R.L. 291
Best, F.A. 253
Bian, X. 23, 359, 377
Bisson, S.E. 121, 311
Blair, F.H. 121
Bowling, S.-A. 373
Bradley, M.M. 43
Brosamlén, G. 171
Brown, P.D. 29
Buch, K.A. 33
Byrne, R.N. 39
Cahalan, R. 363
Carlson, B.E. 151
Carter, E. 215
Chetwynd, J.H. 293
Chin, H.-N.S. 43
Chuang, C.C. 49
Clayson, C.A. 349
Clothiaux, E.E. 283
Clough, S.A. 29, 219
Collatz, G.J. 73
Cook, D.R. 53, 163
Cotton, W.R. 5, 93, 235, 321
Coulter, R.L. 53, 59
Cox, S. 343
Currey, J.A. 63, 349
Davis, A. 363
Dedecker, R.G. 253
Del Genio, A.D. 69
Demos, B. 311
Deng, A. 277
Ding, M. 343
Dirkx, T.P. 253
Dong, X. 1
Doran, J.C. 73
Dudek, M.P. 343
Dudhia, J. 77, 83
Ellingson, R.G. 193, 335
Evans, K.D. 193, 311
Evans, J.L. 89
Fairall, C.W. 101, 355
Fast, J.D. 377
Feingold, G. 93, 321
Ferrare, R.A. 193, 311
Firda, J.M. 279, 283
Flatley, J.M. 97
Frank, W.M. 89
Frisch, A.S. 93, 101
Galbraith, A. 297
Gao, W. 163
Garcia, R. 253
Gautier, C. 105
George, A.T. 173, 185
Ghan, S.J. 111, 113, 117
Goldsmith, J.E.M. 121, 311
Gray, W.M. 125
Grund, C.J. 129
Guggenheim, S.F. 155, 159
Guo, Y.-R. 77, 133
Guo, Z. 273
Hardesty, R.M. 129
Harrison, L. 199, 205
Heymsfield, A.J. 189
Holdridge, D.J. 53, 59
Hong, X. 139
Howell, H.B. 253
Hu, Q. 73
Hubbe, J.M. 73, 287
Jin, Z. 143
Joseph, E. 343
Kain, J.S. 277
Kasayanov, E.I. 389
Knuteson, R.O. 29, 253
Knuth, R.H. 155, 159
Kogan, Y.L. 147
Kogan, Z.N. 147
Koracin, D. 215
Kosters, J. 225
Kreidenweis, S. 235
Kromidas, L. 155
Kropfli, R.A. 189
Krueger, S.K. 167
Kuo, Y.-H. 133
Lacis, A.A. 151, 211
Lare, A.R. 311
Lavalec, D. 105
Lee, H.N. 155
Leifer, R. 155, 159
Lenshow, D.H. 101
Leontieva, E. 307
Lesht, B.M. 53
Leung, L.R. 113, 117
Liang, X.-Z. 343
Liaw, Y.P. 163
Liljegren, J.C. 73
Lilly, D.K. 147
Liu, S. 167
Lohmann, U. 263
Long, C.N. 173
Lovejoy, S. 177, 277
Mace, G.G. 1, 97, 173, 183, 185, 187, 283, 311
Macke, A. 211
Manson, P. 245
Marsden, S. 245
Marshak, A. 363
Martin, T.J. 53, 59
Matrosov, S.Y. 189
Mayer, S.D. 101
McCaa, J. 117
McInnes, K.L. 63
McIntosh, R.E. 259, 279, 283
Melfi, S.H. 193, 311
Meyers, M.P. 235
Michalsky, J. 199, 205
Miller, S. 293
Miller, N.E. 29
Min, Q. 205
Minnett, P.J. 19
-

- Mischenko, M.I. 151, 211
Mitchell, D.L. 215
Mlawer, E.J. 219
Modzelewski, H. 327
Moncet, J.L. 293
Murcay, F. 225
Naud, C. 227
Nemesure, S. 231
O'Hirok, W. 105
Olsson, P.Q. 235
Panchenko, M.V. 241
Parsons, D.B. 83
Patterson, G.R. 245
Pecknold, S. 177
Penner, J.E. 49
Pfeifer, S. 317
Pinto, J.O. 63
Platt, C.M.R. 245
Pol'kin, V.V. 241
Prigarin, S.M. 389
Raman, S. 9, 367
Ramanathan, V. 263
Ramen, S. 139
Reagan, J.A. 297
Revercomb, H.E. 29, 253
Reynolds, R.M. 19
Ricchiazzi, P. 105
Rocque, K. 13
Roeckner, E. 263
Rossow, W.B. 211
Rye, B.J. 129
Sassen, K. 259, 311
Schertzer, D. 177, 277
Schlemmer, J. 205
Schneider, N. 263
Schrieber, K. 327
Schwartz, S.E. 231, 341
Scott, K.A. 287
Scott, S. 187
Scott, V.S. 297
Seaman, N.L. 273, 277
Sekelsky, S.M. 259, 279, 283
Shaw, W.J. 73, 287
Sheaffer, J.D. 125
Shi, L. 291
Shippert, T.R. 29
Silas, P. 177
Sistersen, D.L. 163
Smith, W.L. 29, 253
Snell, H.E. 293
Snider, J.B. 101, 189
Somerville, R.C.J. 39
Spinhirne, J.D. 187, 297
Splitt, M.E. 301
Stage, S.A. 23, 377
Stamnes, K. 143, 307, 373
Starr, D.O'C. 193, 311
Stephen, T. 225
Stevens, B. 93, 321
Strensud, D.J. 317
Stull, R. 327
Subasilar, B. 39
Sun, C.-H. 33, 331
Syrett, W. 1
Takara, E.E. 335
Taubman, S.J. 219
Terpugova, S.A. 241
Thorne, L.R. 33, 331
Thorson, W.B. 125
Titov, G.A. 389, 393, 397
Tsay, S.-C. 143
Tumakov, A.G. 241
Turner, D.D. 29
Wagener, R. 231, 341
Wang, W.-C. 343
Wang, J. 293
Watson, B. 177
Webster, P.J. 349
Weeks, W.F. 143
Wesely, M.L. 53, 301
White, A.B. 355
Whiteman, C.D. 23, 287, 359, 377
Whiteman, D. 193
Whiteman, D.N. 311
Whitney, B.A. 253
Wiscombe, W. 363
Wolf, A.B. 69
Wu, Y. 367
Yang, S.R. 105
Young, S.A. 245
Young, G.S. 89
Zhang, T. 373
Zhang, G.J. 263
Zhong, S. 23, 73, 377
Zhou, C. 205
Zhu, L. 343
Zhuravleva, T.B. 397
Zou, X. 117, 133
Zuev, V.E. 241, 389, 393, 397
Zuev, V.V. 381