

To the Graduate Council:

I am submitting herewith a thesis written by Robert T. McMillen, Jr. entitled "An Investigation of Turbulent Fluxes above a Forest Canopy in Irregular Terrain." I have examined the final copy of this thesis for form and content and recommend that it be accepted in partial fulfillment of the requirements for the degree of Master of Science, with a major in Physics.

W. E. Deeds

William E. Deeds, Major Professor

We have read this thesis
and recommend its acceptance:

George T. Condo

Edward R. Harris

Accepted for the Council:

L. Evans Beth

Vice Chancellor
Graduate Studies and Research

AN INVESTIGATION OF TURBULENT FLUXES ABOVE A FOREST CANOPY
IN IRREGULAR TERRAIN

A Thesis
Presented for the
Master of Science
Degree
The University of Tennessee, Knoxville

Robert T. McMillen, Jr.

June 1983

1 100000

ACKNOWLEDGEMENT

The author wishes to express his appreciation to Dr. W. E. Deeds for his encouragement and guidance throughout the course of this work. He is also grateful to Mr. Bruce Hicks and Dr. Boyd Hutchison of the Atmospheric Turbulence and Diffusion Laboratory for their continued support during this research project.

This work was performed under the auspices of the U. S. Department of Energy and the National Oceanic and Atmospheric Administration.

ABSTRACT

Turbulent wind data were collected above an oak-hickory forest canopy located in irregular terrain in East Tennessee. The data were processed in two distinct ways and the results compared. The first technique involved a three-dimensional coordinate rotation, followed by the accepted method of calculating turbulent fluxes at more ideal sites. The second technique involves real-time mean removal with digital recursive filters, and real-time calculation of the turbulent fluxes. Stability of the solution is examined as a function of the mean removal filter time constant and the accumulation period of the solution.

TABLE OF CONTENTS

CHAPTER		PAGE
I.	INTRODUCTION	1
II.	BACKGROUND AND HISTORY	3
	Turbulent Fluxes.....	3
	Previous Methods.....	10
III.	METHODS AND TECHNIQUES.....	16
	Site.....	16
	Instrumentation--Phase 1	18
	Instrumentation--Phase 2	21
	Data Analysis--Phase 1	22
	Data Analysis--Phase 2	24
IV.	DISCUSSION AND RESULTS	26
	Coordinate Rotation.....	26
	Real-time Analysis.....	31
V.	CONCLUSIONS.....	45
	REFERENCES	46
	VITA	50

LIST OF FIGURES

FIGURE		PAGE
1.	Map of Walker Branch Watershed and Surroundings.....	17
2.	Atmospheric Turbulence and Diffusion Laboratory's 44 Meter Tower.....	19
3.	Gill-type UVW Propellor Anemometer.....	20
4.	Illustration of Coordinate Rotation.....	23
5.	Normalized Standard Deviations of u , v , and w , as a Function of Wind Direction.....	27
6.	Reynolds Stress as a Function of Wind Direction.....	28
7.	U^* as a Function of Wind Direction.....	29
8.	Coefficient of Drag as a Function of Wind Direction.....	30
9.	$\bar{U}$ versus Accumulation Time.....	35
10.	$\bar{W}$ versus Accumulation Time.....	36
11.	Variance of $u'u'$ versus Accumulation Time....	38
12.	Covariance of $u'v'$ versus Accumulation Time..	39
13.	Covariance of $u'w'$ versus Accumulation Time..	40
14.	Variance of $v'v'$ versus Accumulation Time....	41
15.	Covariance of $v'w'$ versus Accumulation Time..	42
16.	Variance of $w'w'$ versus Accumulation Time....	43

CHAPTER I

INTRODUCTION

One of the primary areas of research in micrometeorology is the investigation of turbulent fluxes as a transport mechanism of basic atmospheric quantities. Increasing awareness of environmental problems has emphasized the need of measuring turbulent fluxes of heat, momentum, and atmospheric pollutants at a variety of sites and in varying conditions. In the past, these quantities have usually been measured at carefully selected "ideal" sites, i.e., level terrain with a smooth homogeneous surface, using a variety of techniques such as energy balance, aerodynamic or profile methods, or the eddy correlation method. The need now exists to make such flux measurements at sites far from ideal, and, in particular, at sites above and within forest canopies in irregular or complex terrain.

The Atmospheric Turbulence and Diffusion Laboratory (ATDL) forest meteorology group has begun a program to measure turbulent fluxes of heat, momentum, and pollutants above and within the forest canopy at the Walker Branch Watershed (WBW) in East Tennessee. The surrounding terrain is complex and surface of the oak-hickory forest canopy is inhomogeneous, providing a poor site for

application of existing flux calculation methods but an excellent location for the development of new techniques.

The objectives of this study were two-fold. The first objective was to assess the feasibility of collecting self-consistent turbulent flux data at that site. It was necessary to determine if turbulent parameters could be calculated at that site which would at least have the correct order of magnitude and sign.

The second objective of the study was to develop a computer algorithm which would run on a microcomputer and calculate turbulent fluxes in real-time at the WBW site and similar non-ideal sites. Real-time calculations of the variances and covariances of the fluctuating components of the wind allow the turbulent flux data to be continually monitored for quality assurance. A technique was developed involving mean removal by digital recursive filtering which appears to be successful and has been implemented on several microcomputers.

CHAPTER II

BACKGROUND AND HISTORY

1. Turbulent Fluxes

"Turbulence is the most common, the most important and the most complicated kind of fluid motion. It is particularly resistant to mathematical treatment: indeed, turbulence studies may be defined as the art of understanding the Navier-Stokes equations without actually solving them." (Bradshaw, 1971)

As a basis for any discussion of atmospheric turbulence, we must first introduce some basic aspects of the meteorology of the planetary boundary layer. We assume that air follows the perfect gas law

$$p = \rho R T, \quad (1)$$

where R is R_1 / M , R_1 being the universal gas constant, M is the molecular weight of the mix of atmospheric gases, and T is the air temperature. M is nearly constant, since water vapor is the only variable of consequence in the mixture. We can also write the hydrostatic pressure equation

$$\frac{dp}{dz} = - \rho g, \quad (2)$$

where z is height and g is the gravitational acceleration.

Applying basic thermodynamics, we can write for a parcel of air that exchanges no heat with its surroundings

$$0 = C_p dT - \frac{1}{\rho} dp, \quad (3)$$

where C_p is the specific heat at constant pressure. Using

the hydrostatic equation we can define the adiabatic temperature gradient

$$\frac{dT}{dz} = - \frac{g}{C_p}, \quad (4)$$

which for dry air is about 1 degree C per 100 meters.

Atmospheric stability is a consequence of the temperature profile, which determines the vertical density gradient, as approximated by Boussinesq (1901). If the air temperature falls more quickly with increasing elevation than the adiabatic lapse rate, then the atmosphere in that layer is unstable. A parcel of air undergoing an infinitesimal vertical displacement experiences a negative restoring force. If the temperature falls less slowly than the adiabatic lapse rate, the layer is stable and the displaced parcel experiences a positive restoring force, sending it back to its original position. Stable conditions prevail most dramatically on clear nights, when radiative cooling causes the surface to be much cooler than the adjacent layers of the atmosphere. Unstable conditions prevail during sunny weather, when the surface is warmer than the air. The unstable layer is generally limited to the first one or two hundred meters of the atmosphere. Unstable conditions must be continually driven by an energy source, otherwise the unstable layer quickly becomes thoroughly mixed, and the temperature profile assumes that of the

adiabatic lapse rate, implying neutral stability.

Air flow near the earth's surface is, in almost all cases, highly turbulent and, as such, is not well understood. Turbulent flow was first described by Reynolds in 1883 following experiments involving flow through long tubes. He thought of the flow as consisting of two modes, a mean flow and a superimposed fluctuating flow. The mean flow may be thought of as a "laminar" flow, where each small air parcel follows the exact path of the parcel preceding it and of the parcel following it, i.e., "streamline" flow. The turbulent component of the flow, on the other hand, is almost completely random. These random motions are in some manner related to the bulk or mean flow of the air.

Turbulence near the earth's surface is generated by either of two mechanisms, mechanical mixing caused by surface roughness or buoyancy forces resulting from atmospheric instability. An extreme example of mechanical turbulence generated by surface roughness is the swirling flow in the wake of a building during strong winds. Atmospheric instability has already been discussed and is the primary factor in vertical mixing of the planetary boundary layer.

If we wish to investigate some of the consequences of atmospheric turbulence, the following relationships will be helpful. First, consider two arbitrary atmospheric

quantities a and b , which might be components of wind velocity, temperatures, or concentrations of airborne pollutants. We have already mentioned that the quantity can be thought to consist of a mean and a fluctuating component, which we write as

$$a = \bar{a} + a' \quad , \quad (5)$$

where a is the instantaneously measured value of the quantity, $\bar{a}$ is the mean value, and a' is the fluctuating component or deviation from the mean. Similarly,

$$b = \bar{b} + b' \quad . \quad (6)$$

We define the mean value, $\bar{a}$, using Reynolds averaging, as

$$\bar{a} = \frac{1}{T} \int_{t-T_0/2}^{t+T_0/2} a \, d\tau \quad , \quad (7)$$

where T is an arbitrary averaging time, and we might also note that, since

$$a' = a - \bar{a} \quad , \quad (8)$$

we can write

$$\bar{a'} = \frac{1}{T} \int_{t-T_0/2}^{t+T_0/2} (a - \bar{a}) \, d\tau \quad . \quad (9)$$

We also wish to carry out this averaging on products of arbitrary atmospheric quantities, and following Hinze (1959), we write

$$\overline{a b} = \bar{a} \bar{b} \quad , \quad (10)$$

$$\overline{a' b'} = \overline{a' b'} = 0, \quad (11)$$

and

$$\overline{a b} = \overline{(a + a')(b + b')} = \overline{a b} + \overline{a' b'} \quad (12)$$

We should also note that since a' is, by definition, a deviation from a mean, a'^2 is the variance, and $\sqrt{a'^2}$ is the standard deviation, usually referred to as sigma-a.

Similarly $\overline{a' b'}$ is the covariance of a and b .

The components of wind velocity are usually written as u , the horizontal component in the direction of the mean wind vector, v , the horizontal component perpendicular to u , and w , the vertical wind component. We can write the continuity equation for an arbitrary atmospheric quantity as

$$\frac{da}{dt} = \frac{\partial a}{\partial t} + u \frac{\partial a}{\partial x} + v \frac{\partial a}{\partial y} + w \frac{\partial a}{\partial z} = R + S \quad (13)$$

where R is the sum of the external effects, S is the sum of the internal sources, and x , y , and z are defined to be a Cartesian axes corresponding to the u , v , and w wind components. Averaging this equation yields

$$\frac{d\overline{a}}{dt} = \frac{\partial \overline{a}}{\partial t} + \overline{u \frac{\partial a}{\partial x}} + \overline{v \frac{\partial a}{\partial y}} + \overline{w \frac{\partial a}{\partial z}} = \overline{R} + \overline{S} - \frac{\partial}{\partial x}(\overline{u' a'}) - \frac{\partial}{\partial y}(\overline{v' a'}) - \frac{\partial}{\partial z}(\overline{w' a'}), \quad (14)$$

where the quantities $\overline{u' a'}$, $\overline{v' a'}$, and $\overline{w' a'}$ are the turbulent fluxes of the quantity a and are of particular interest.

If the quantity a is a wind velocity component, then the terms $\overline{u' a'}$, $\overline{v' a'}$, and $\overline{w' a'}$, when multiplied by the density, ρ , are called the Reynolds stresses. The normal

stress, $\overline{\rho u'u'}$, and the shear stresses, $\overline{\rho u'v'}$ and $\overline{\rho u'w'}$, are all technically Reynolds stresses, but we usually refer to only the vertical shear stress, $\overline{\rho u'w'}$, as the Reynolds stress. This quantity, usually called τ , is the vertical flux per unit area of horizontal momentum. The shear stress, a consequence of viscosity, is proportional to the vertical velocity gradient, and we can write

$$\tau = \mu \frac{d\bar{u}}{dz} = -\overline{\rho u'w'} \quad , \quad (15)$$

where μ is the viscosity.

A convenient scaling velocity, u^* , can be defined as

$$u^{*2} = \left| \frac{\tau}{\rho} \right| = \left| \overline{u'w'} \right| \quad (16)$$

and is called the friction velocity. Sutton (1953) points out that u^* has the same order of magnitude as the eddy velocities and a useful rule of thumb is that u^* is about 1/10 of the mean horizontal wind.

Assume that, at a point near the surface, an arbitrary atmospheric quantity can be measured and has an instantaneous value a , and that the air, also at the same point, has an instantaneous density ρ and upward vertical velocity w . The amount of the quantity a being transported upwards per unit area, i.e., the vertical flux of a , is, instantaneously, ρwa . The density may or may not be needed, depending on the nature of the quantity a but will be

included in the discussion for generality.

The complete vertical flux of the quantity a can be written in the form

$$F_a = \rho w a \quad , \quad (17)$$

which we can also write as

$$F_a = \rho (w + \bar{w}') (a + \bar{a}') \quad . \quad (18)$$

Now if we average the vertical flux over some averaging time T , we can write

$$\overline{F_a} = \rho \bar{w} \bar{a} + \rho \overline{w' a'} \quad , \quad (19)$$

since the terms $\overline{w' a'}$ and $\overline{w' a}$ are zero by definition. The two terms on the right are the advective vertical flux and the turbulent vertical flux. In ideal conditions we expect the mean vertical wind, $\bar{w}$, to be zero, so that the entire vertical flux is just the turbulent flux. This term represents the flux to or from the surface.

Problems arise in non-ideal conditions when the first term is non-zero, since $\bar{w}$ is non-zero. This occurs whenever the flow streamline is not horizontal. A non-horizontal streamline can be due to a non-horizontal surface or to changing surface conditions when the surface itself is horizontal. If the surface boundary layer changes due to changing surface conditions, for instance from plowed fields to grassland, then the streamlines near the surface must necessarily be non-horizontal. For this reason an ideal site has a large "fetch" or distance in which the boundary

layer has sufficient time to readjust to the new surface conditions and the streamlines are once again horizontal. To calculate the flux to the surface, or perpendicular to the streamline, we can force $\bar{w}$ to zero by rotating the coordinate system of the wind components, so that $\bar{u}$ is now along the mean wind vector rather than along the horizontal component of the mean wind vector.

Several researchers (Dyer and Hicks, 1972, and Kaimal and Haugen, 1969), have reported large errors in the calculation of Reynolds stresses due to improper leveling of the vertical wind sensor. These errors seem to be about 10 percent per degree of misalignment and since anemometers used from towers tend to be mounted on rather long booms, this requirement presents a serious problem to the field researcher. The same condition occurs if the sensor is vertical but the flow streamline is not horizontal, a condition which occurs frequently in irregular terrain. For this reason direct measurements of vertical fluxes at these sites have not been thought to be effective.

2. Previous Methods

Turbulent exchange of atmospheric quantities is driven by eddies ranging in size from several hundred meters to molecular scales. Examinations of power spectra of turbulence imply that frequencies greater than 2 Hertz (Hz)

contribute little , if any, to the total flux (Hicks, 1972). At the other end of the scale the problem becomes one of definition. At what point do the eddies cease to be eddies and become slowly changing background or mean conditions? Different averaging times are probably necessary at different sites, but most researchers seem to average over periods ranging from 15 minutes to one hour.

A useful approximation regarding the size of eddies contributing to vertical flux is that the maximum size of the eddy should have the same order of magnitude as the distance which the anemometer is mounted above the surface. This distance, divided by the mean velocity gives a good lower bound for frequencies of turbulence which contribute to vertical fluxes.

Turbulent fluxes can be calculated directly only with fast response instrumentation and analogue or fast digital computers (fast in micrometeorology being 1 Hz or faster). Before these devices became available, turbulent fluxes could be inferred only indirectly by means of energy balance methods or aerodynamic or profile methods.

In the energy balance method, one tries to calculate all but one of the terms in the energy balance equation

$$R - G = H + LE \quad (20)$$

where R is net radiation, G is heat flux to the ground, H is

the sensible heat flux, and LE is the latent heat flux. The first term can be measured with a net radiometer to at least within a few percent. The second term is measured less accurately, depending on the method of measurement of heat flux to the soil, but is usually quite small so lack of measurement accuracy introduces no large errors. To determine H and LE it is then necessary to assume that the transfer coefficients of the two fluxes are equal. Several researchers (Swinbank and Dyer, 1967 and Dyer, 1967) have found this to be true during lapse or neutral conditions. With this assumption we can apply the Bowen ratio technique

$$\beta = \frac{H}{LE} = \frac{C_p}{L_p} \left(\frac{T_2 - T_1}{q_2 - q_1} \right) \quad , \quad (21)$$

and then writing

$$LE = \frac{R - G}{1 + \beta} \quad , \quad (22)$$

solve for the latent and sensible heat fluxes.

Another indirect technique is the aerodynamic or profile method where, following Sutton (1953), the mean flux of a quantity a is approximated by

$$F_a = K_a \frac{d\bar{a}}{dz} \quad . \quad (23)$$

The term K is called the turbulent exchange coefficient or the diffusivity coefficient. A necessary assumption of this technique is that the exchange coefficients for sensible and latent heat, horizontal momentum, and pollutants are equal,

at least under conditions when the method is valid.

After making this assumption we can write

$$F_a = F_m \frac{\bar{a}_2 - \bar{a}_1}{\bar{u}_2 - \bar{u}_1} \quad . \quad (24)$$

This requires accurate knowledge of F (or any other vertical flux), as well as precise measurements of $\bar{u}$ and $\bar{a}$ at several heights above the ground. A major problem with the aerodynamic technique is that it requires ideal (i.e., homogeneous) conditions and terrain.

It appears that at non-ideal sites, turbulent vertical fluxes should be measured directly via the eddy correlation or covariance method. As noted earlier, this involves the direct measurement of the fluctuating components, w' and a' , and calculation of the covariance of these two terms. The areal representativeness of the flux may be questionable at non-ideal sites, but the primary objective has been met. If areal representativeness of the measurements is indeed a problem, it might be resolved by spatial replication of the flux measurements. This problem has not been investigated at our site.

Covariance methods were first applied to the calculation of vertical momentum fluxes by analysis of strip chart data of the horizontal and vertical components of fast response anemometers by British workers in the early 50's. The means of the instantaneous signals were subjectively estimated from strip chart traces and the fluctuating

components calculated and the covariances formed. Later, analog computer techniques were successfully employed to obtain the fluxes in real time. The analog methods were successful, but suffered from electronic drift, were somewhat balky in use, and needed careful alignment and calibration in the laboratory before being taken into the field. The Evapotron, reported by Dyer and Maher (1965), used measurements from heated wire anemometers and fine wire wet and dry bulb thermometers to calculate sensible and latent heat fluxes. The most successful analog approach, the Fluxatron, reported by Dyer and Hicks (1967), was used to measure fluxes of sensible and latent heat. The sensors used were a vertical propeller anemometer, a vane-oriented horizontal propeller anemometer, and fast-response thermocouples for wet and dry bulb temperatures.

More recently, emphasis has shifted to digital techniques throughout the process of calculating atmospheric fluxes. Analog measurements can be duplicated with digital techniques if the experimenter has access to a reasonably fast digital computer, coupled with a reasonably fast analog to digital (A/D) converter. If this combination is fast enough to obtain the calculated fluxes in real time, the experimenter can effectively sidestep the major difficulties of the analog process, namely calibration drift due to instabilities in the analog amplifiers and

inaccuracies in the filters. Since portable digital machines of this power are now readily available, the digital technique has become the method of choice.

CHAPTER III

METHODS AND TECHNIQUES

1. SITE

The ATDL Forest Meteorology Research Facility is located on the U.S. Department of Energy's Oak Ridge Reservation and is situated on the west boundary of Oak Ridge National Laboratory's (ORNL) Walker Branch Watershed (Figure 1). The ATDL instrument tower is situated near the top of a ridge rising to the northeast with narrow valleys falling off to the southwest and northwest. The surrounding terrain is quite convoluted. The forest is an uneven-aged oak-hickory stand, typical of forests throughout the Appalachian region, with an average tree height of about 22 meters. The leaf area index, i.e., the total leaf area (one side of the leaf only) per unit area of earth surface, is about 4.9 at full leaf and the woody plant area index, defined similarly as the total projected plant stem area per unit area of earth surface, is about 0.6, giving a total plant area index of about 5.5.

Surface flows in the East Tennessee area tend to be channeled by the overall ridge/valley topography into a southwesterly or northeasterly direction, and they follow this pattern at our site. Winds from the northeast exhibit

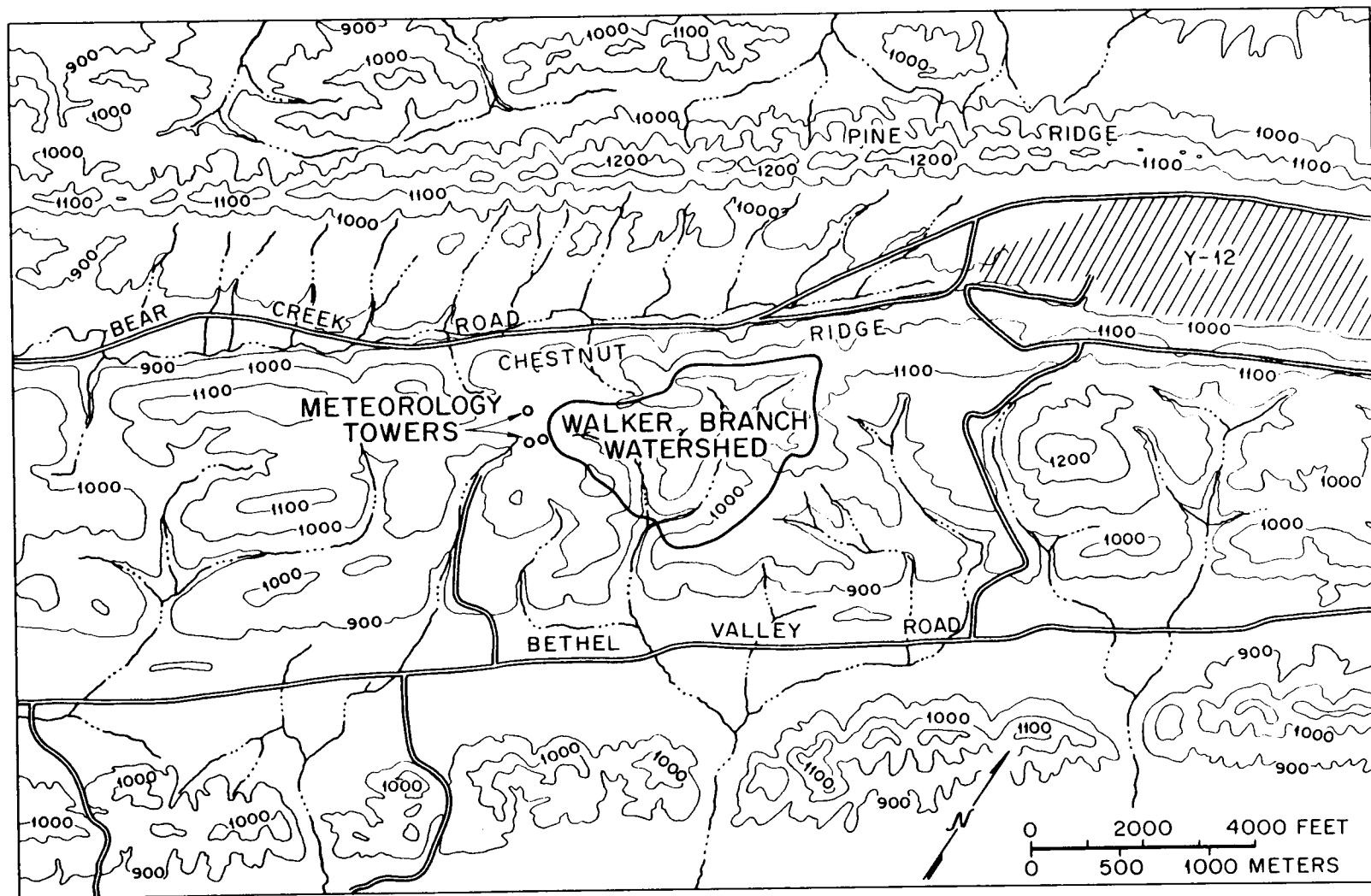


Figure 1. Map of Walker Branch Watershed and Surroundings

a downslope angle of about 0.4 degrees with respect to the horizontal, while winds from the southwest exhibit an upslope angle of about 5 degrees.

A 44 meter walkup tower is located at the site (Figure 2), as well as two 33 meter triangular towers. A small utility building and trailer house the data logging equipment and maintenance supplies.

2. Instrumentation--Phase 1

Measurements began in the fall of 1981, and were continued at irregular intervals through the fall of 1982. A R. M. Young Gill-type UVW propeller anemometer, illustrated in Figure 3, was used to measure the three components of the vector wind. The Gill propeller anemometer is used extensively in micrometeorology. Developed by Gill (1967), it is an orthogonal set of propeller anemometers. Each individual anemometer consists of a generator driven by a helicoid propeller of either styrene or, in our case, styrofoam. The threshold for the propeller is about .25 meters per second, and response to axial flow is linear above this threshold. The response length of the styrofoam propellor is about 1 meter, which, under the conditions in which we operate, corresponds to a response time of 0.5 to 1.5 seconds. The off-axis cosine response of these anemometers is not ideal, but is well

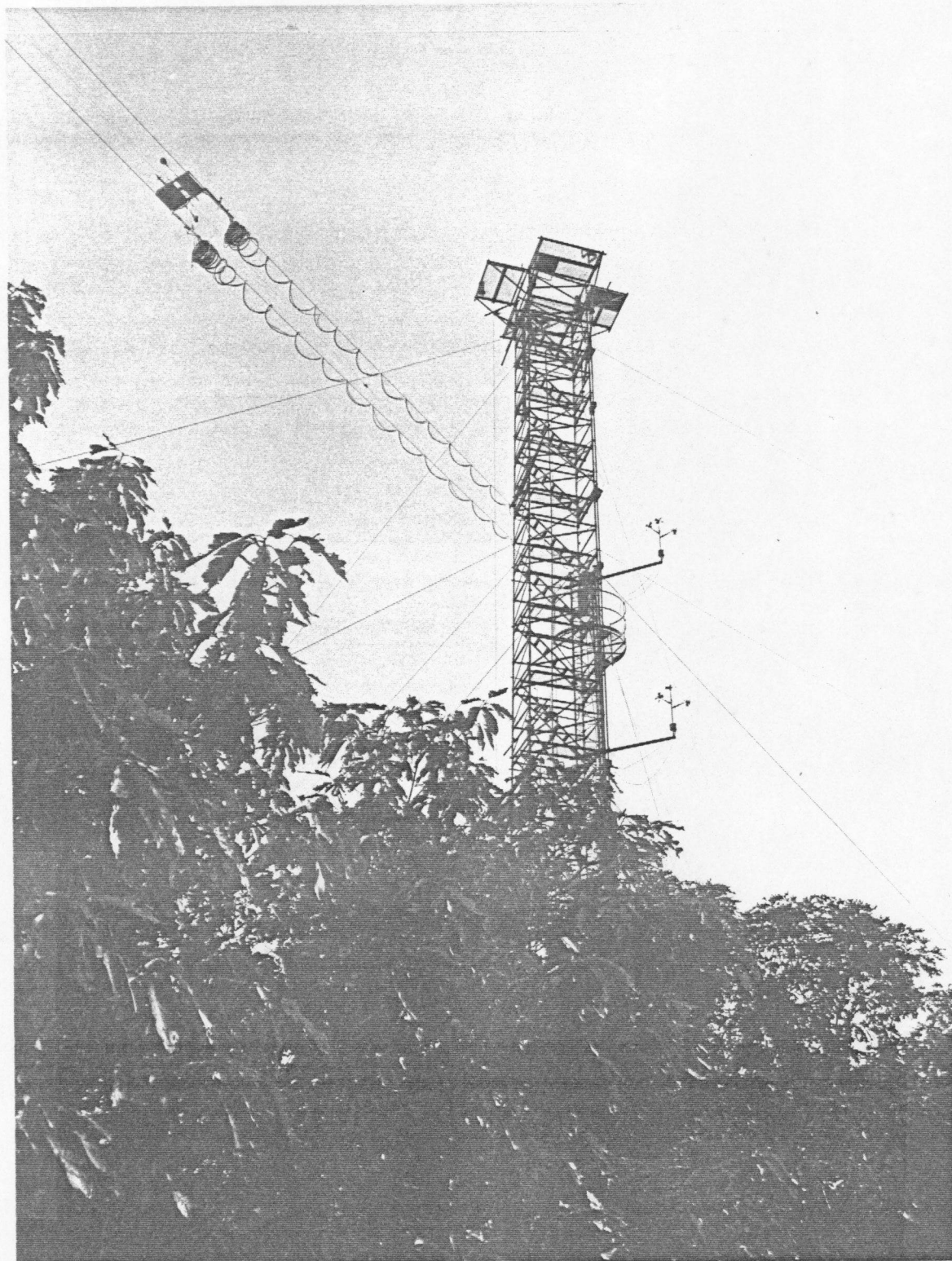


Figure 2. Atmospheric Turbulence and Diffusion
Laboratory's 44 Meter Tower

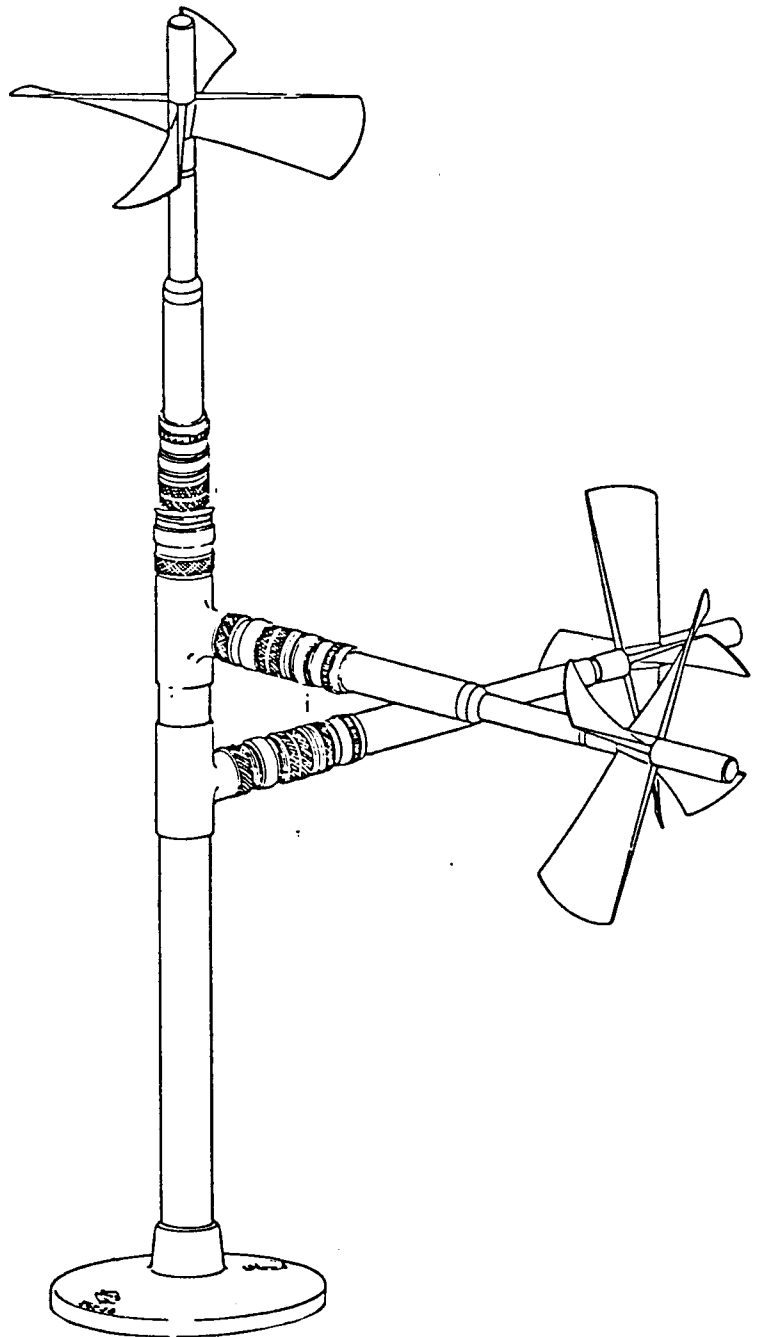


Figure 3. Gill-type UVW Propellor Anemometer

documented by Gill (1975) and Horst (1973), and the data can be corrected for non-cosine response. The anemometer was mounted on a 2 meter long horizontal boom affixed to the 44 meter walkup tower at a height of 29 meters, about 7 meters above the canopy top.

The data acquisition system was controlled by a PDP-8E minicomputer, and had a maximum scanning rate of 12 channels per second. The three axes of the anemometer were scanned once per second for a total of 102 hours, recorded on 7-track magnetic tape, and subsequently processed on an IBM 3033 mainframe computer at ORNL. All propellers and generators of the anemometer were calibrated in the ATDL wind tunnel for along-axis flow and for non-cosine response.

3. Instrumentation--Phase 2

The second phase of the study took place in October of 1982. At that time, data were taken using a Hewlett-Packard 1000L minicomputer-based data acquisition system. The UVW anemometer was positioned at the same height, but in this case, the output signal of each axis of the anemometer was scanned four times per second. The data were recorded on 9-track magnetic tape, and were processed on the IBM 3033 computer as well as on a Hewlett-Packard HP-85A microcomputer. Approximately 50 hours of 4 Hz data are available, but only a few hours have been examined.

4. Data Analyses--Phase 1

The data were first adjusted for non-cosine response using an iterative technique described by Horst (1971). This technique determines the direction cosines from the uncorrected wind components, applies the corrections for non-cosine response to the components, and then recalculates the direction cosines using the corrected wind components. This procedure is repeated until the difference between successive calculations of the direction cosines is less than 1 percent.

After adjusting the data for non-cosine response, hourly averages of the components of wind velocity and the mean wind vector are calculated. Then a coordinate rotation is performed, first rotating the anemometer data about the w axis to bring the uw plane coincident with the mean wind vector. This rotation is angle α in Figure 4. Next the data are rotated about the current v axis, aligning the u axis with the mean wind vector (rotation angle β in the illustration). A final small rotation, γ , is made about the u axis which minimizes the mean covariance of $\overline{u'v'}$. The last correction is usually quite small, less than 1 degree.

At this point the coordinates of the data have been adjusted so that $\bar{u}$ is into the mean wind vector. The

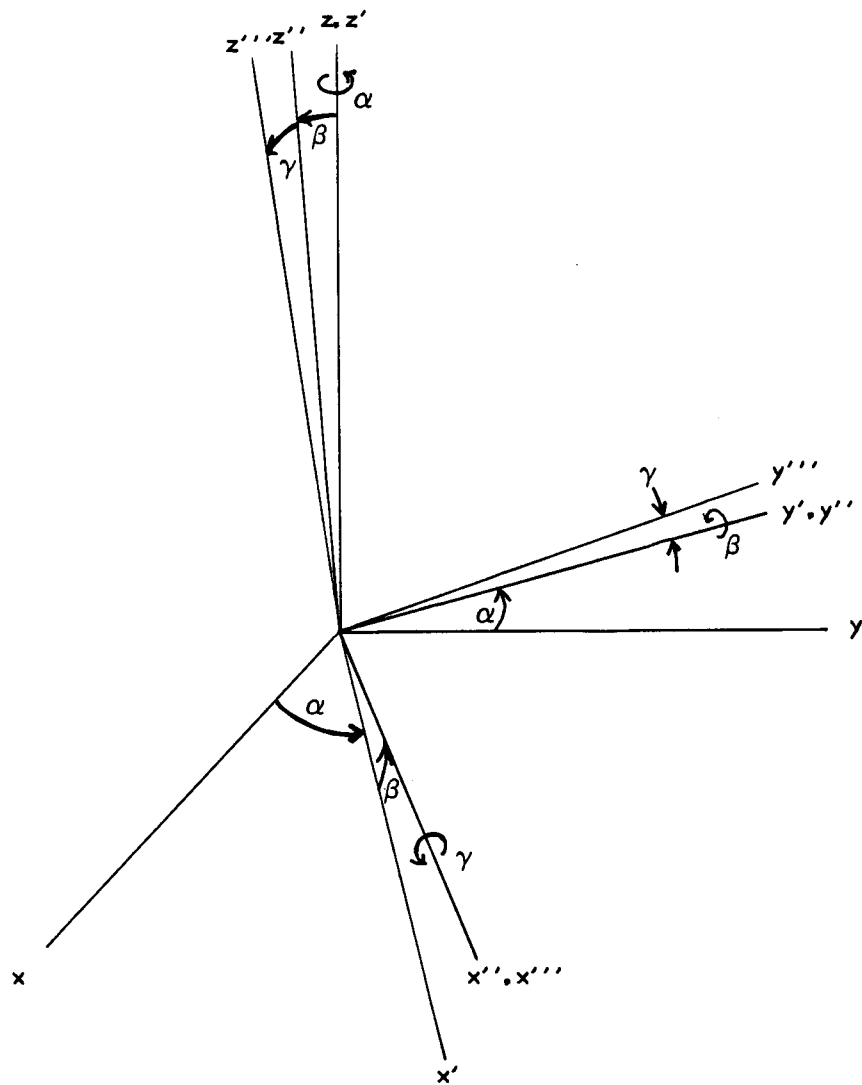


Figure 4. Illustration of Coordinate Rotation

only problem is that in most cases at our site the vertical axis determined by this procedure is no longer aligned with the geopotential vertical. The misalignment is usually in the range of .5 to 4 degrees, and has been ignored.

New hourly averages of the rotated wind components were calculated ($\bar{v}$ and $\bar{w}$ were zero by definition), and turbulent fluxes, as well as other turbulence parameters, are calculated. The calculation of vertical flux is just the average cross-product of the fluctuating components. The other turbulence parameters calculated were the friction velocity, σ_u , σ_v , σ_w , and the drag coefficient, where the drag coefficient is defined as

$$C_d = \frac{\tau}{\rho \bar{u}^2} = \left(\frac{u^*}{\bar{u}} \right)^2 \quad . \quad (25)$$

5. Data Analyses--Phase 2

Data analyses of Phase 2 data consisted of two parts. In the first part, the data were processed exactly as in Phase 1. Turbulence parameters were calculated for 10 hours of data. For the second part, a different technique was employed; the data were analyzed in a pseudo-real-time, that is, the computer only had access to data being currently read or to data previously read. The purpose was to develop and verify a real-time data processing technique which would provide the same results as the technique

employed in Phase 1.

This technique employs recursive digital filters to calculate running means of the instantaneous data and eliminates the need for coordinate rotation, which can only be done correctly during post-processing, precluding real-time analysis.

CHAPTER IV

DISCUSSION AND RESULTS

1. Coordinate Rotation

Our primary concern during the first stage of this study was whether we could calculate "reasonable" turbulence parameters from measurements made over the irregular terrain at our site. As previously mentioned, Dyer and Hicks (1972) have reported errors in the calculation of the Reynold's stress of between 5 and 11 percent per degree of misalignment of the w-axis sensor with the geopotential vertical. The errors are the same whether the instrument is misaligned with the vertical or the mean flow is misaligned with the horizontal. The most straightforward solution to the misalignment problem would be to orient the sensors along the mean streamline; this technique has been used in very steady non-horizontal flows, but is not practical in the generally light and variable winds that prevail at our site.

Hourly averages of u , v , and w were used to define the new orientation of the coordinate system used for data analyses in phase 1 of the study. The results presented in Figures 5 through 8 are various turbulence parameters calculated with the rotated wind data. Shown in Figure 5

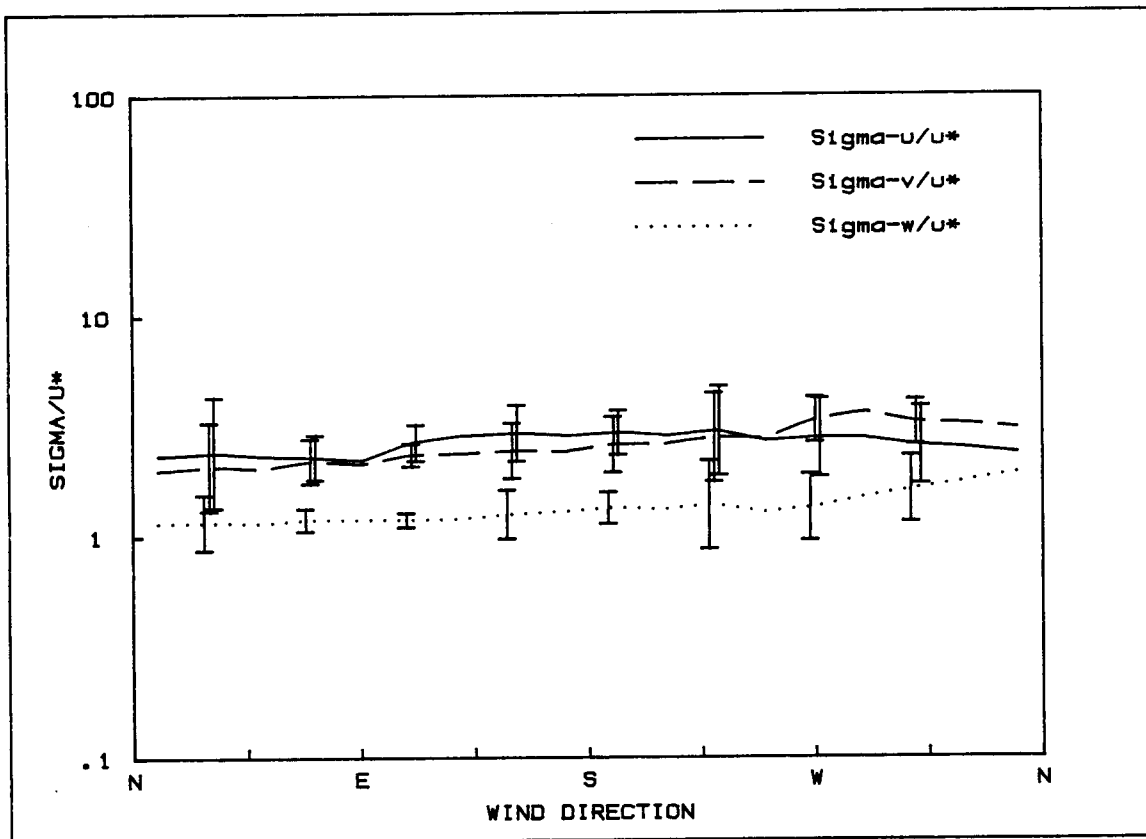


Figure 5. Normalized Standard Deviations of u , v , and w , as a Function of Wind Direction

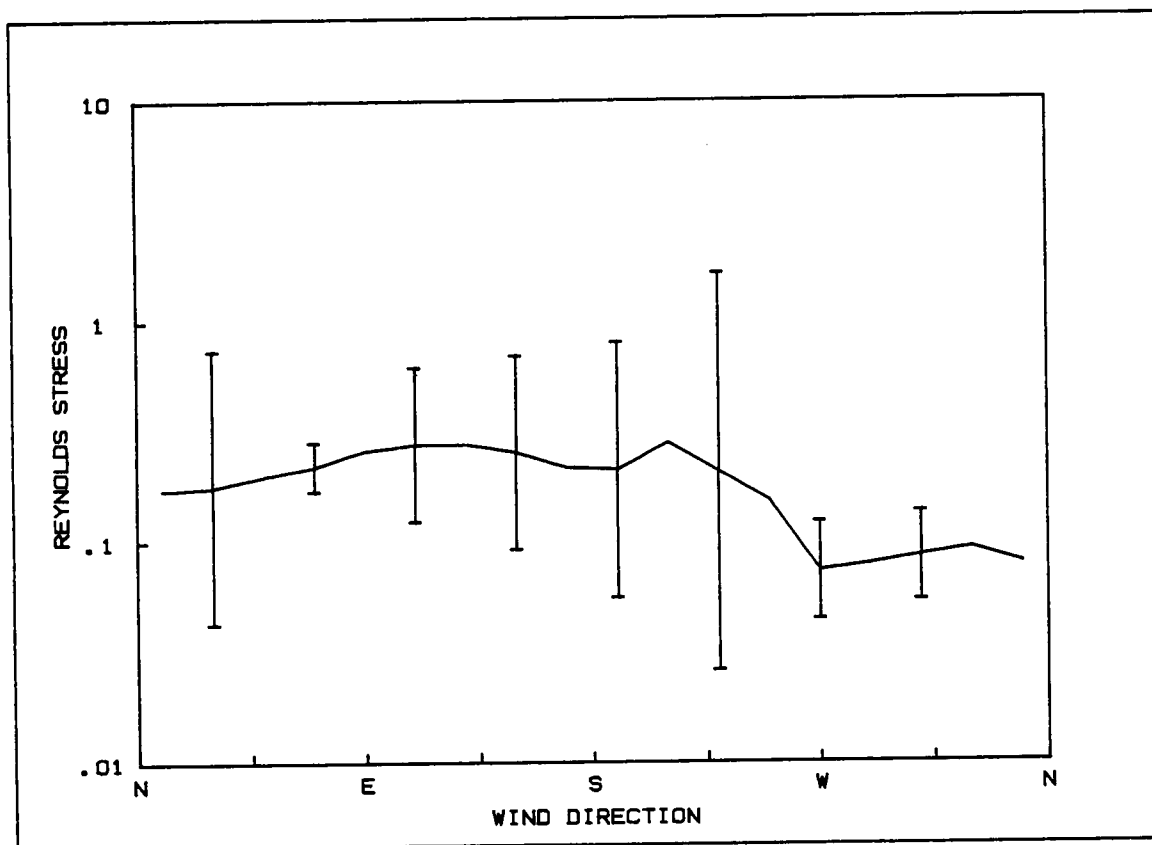


Figure 6. Reynolds Stress as a Function of Wind Direction

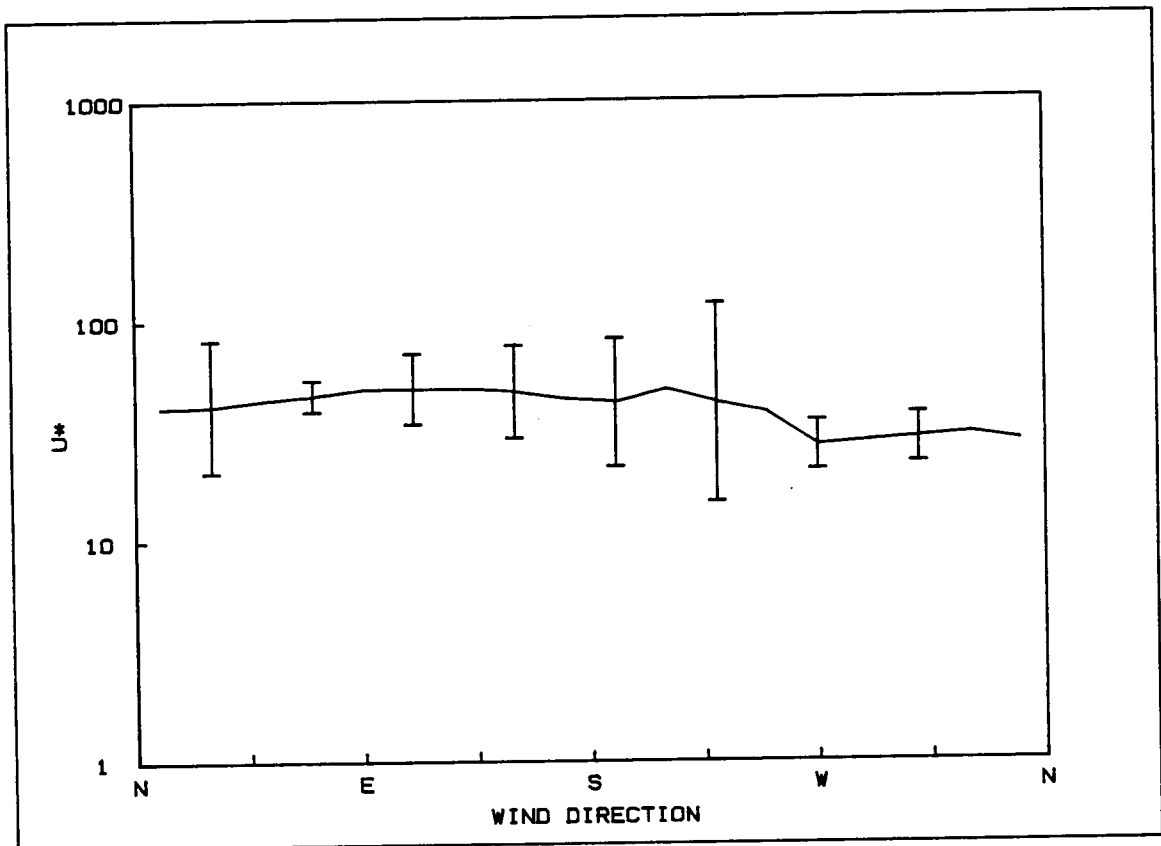


Figure 7. U^* as a Function of Wind Direction

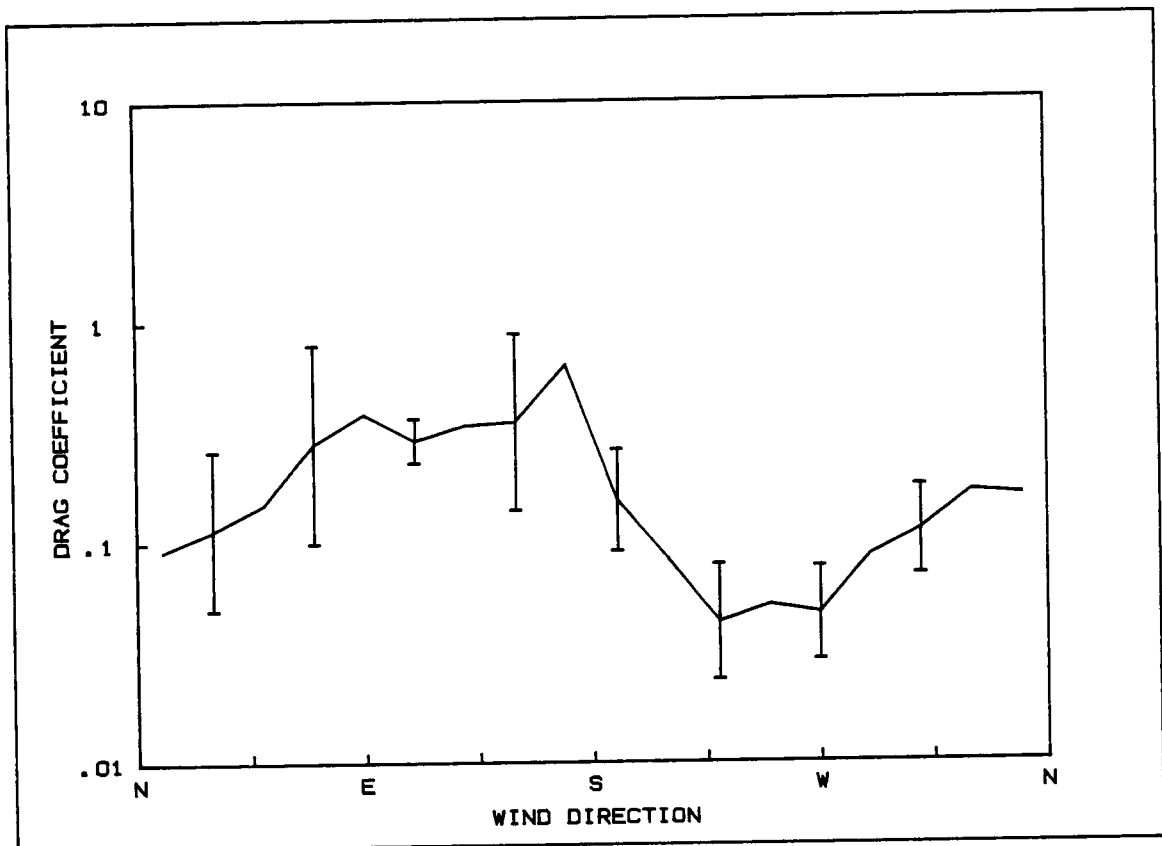


Figure 8. Coefficient of Drag as a Function of Wind Direction

are $\sigma-u/u^*$, $\sigma-v/u^*$, and $\sigma-w/u^*$, all as a function of the wind direction, θ . Figures 6, 7, and 8 illustrate the Reynolds stress, u^* , and the drag coefficient, again all as functions of the wind direction, θ . The data are well-behaved, with little scatter, and the sigmas have ratios comparable to those reported by Hicks (1981). Error bars are shown on these plots and are plus and minus one standard deviation. The means for $\sigma-u/u^*$, $\sigma-v/u^*$, and $\sigma-w/u^*$ were found to be 2.57, 2.25, and 1.29 respectively, when all stability conditions were included. These results indicate that useful eddy correlation measurements can be made at this site, providing that we first transform the coordinate system, since they are comparable to values measured at uniform sites.

2. Real-Time Analysis

The technique used in the preceding section, while affording plausible results, also leads to difficulties. The one hour averaging time is quite arbitrary and, indeed, selection of a proper averaging time presents a problem. The averaging period should be long enough for statistically stable solutions and short enough to assure stationarity of the variables we wish to measure. In complex terrain these two criteria are often mutually exclusive. We know that the

driving force for most turbulent fluxes is solar insolation which, on a clear day, changes about 15 percent per hour. In addition, we have the variability of the wind direction with which to contend. Since the variables of interest are covariances, the selection of a proper averaging time is critical.

It seems that the best solution would be to use two different averaging times in the calculation process, one for the removal of the means and a different averaging time for calculation of a stable covariance. A running mean, used for mean removal to calculate the fluctuating components, together with a separate averaging time for the cross-products of the fluctuating components should resolve the problem of selection of a proper averaging time.

Having shown that the site was adequate for eddy correlation calculations, a real-time calculation algorithm was developed. It is very convenient, and perhaps essential, to have eddy correlation data available in real-time in order to monitor the progress of the experiment and check the validity of the data being collected. Real-time calculation of fluxes cannot be done using the complete three-axis coordinate rotation scheme outlined above since the anemometer array is "rotated" using after-the-fact calculated ensemble averages of the appropriate measured quantities. A running mean seemed the best approximation to

the time or ensemble average, but due to limited real-time computing power in a portable system, the running mean was approximated with a recursive digital filter, as described by Enochson and Otnes (1968).

A recursive digital filter is an exact analog of an ideal electronic R-C filter, with an easily specified time constant, and, with the given time constant, closely approximates a running mean over the same time. The most general expression for a compound digital recursive filter can be written

$$y_i = \sum_{k=M_1}^{M_2} h_k x_{i+k} + \sum_{l=1}^{M_3} g_l y_{i-l} \quad , \quad (26)$$

where x_i is the original record of time series data, y_i is the filtered data, h_k and g_l are arbitrary constants determined by the desired transfer function of the filter. Perhaps the most simple digital recursive filter, and the one used in our analysis is written

$$y_i = \alpha x_i - (1 - \alpha) y_{i-1} \quad , \quad (27)$$

where $\alpha = e^{-(\Delta t/RC)}$, which is completely analagous to an electronic R-C low-pass filter. The filter serves two purposes: first, it removes trends in the data, which is especially necessary in complex terrain; and second, it approximates the ensemble averages in real-time, allowing real-time calculation of the fluctuating components of the

wind and their variances and covariances. While the raw data cannot be rotated in real time, it is possible to rotate the accumulated cross-products through angles determined by the running means. It was found that only a single-axis rotation was necessary to provide good results; the accumulated covariances are rotated about the vertical to force the mean crosswind velocity component to zero. This rotation seemed to marginally improve the calculation of the covariances involving only u' or v' and was nearly negligible for cross-products containing w' . An additional rotation bringing the mean vertical velocity component to zero did not improve the results. Since the covariance of primary interest, $\overline{u'w'}$, is largely unaffected by this rotation, it was neglected in the final formulation of the technique.

Figures 9 and 10 are the filtered velocity components calculated by applying digital filters with time constants of 300, 500, 750, 1000, and 1500 seconds to the raw data. The horizontal axis is the time over which the filter has been applied to the data. These are the same quantities that we would see if we applied R-C filters with the same time constants to the analog input signals. The heavy horizontal lines in Figures 9 and 10 are the hourly averages of the respective velocity components calculated from the raw data after the fact. These figures are

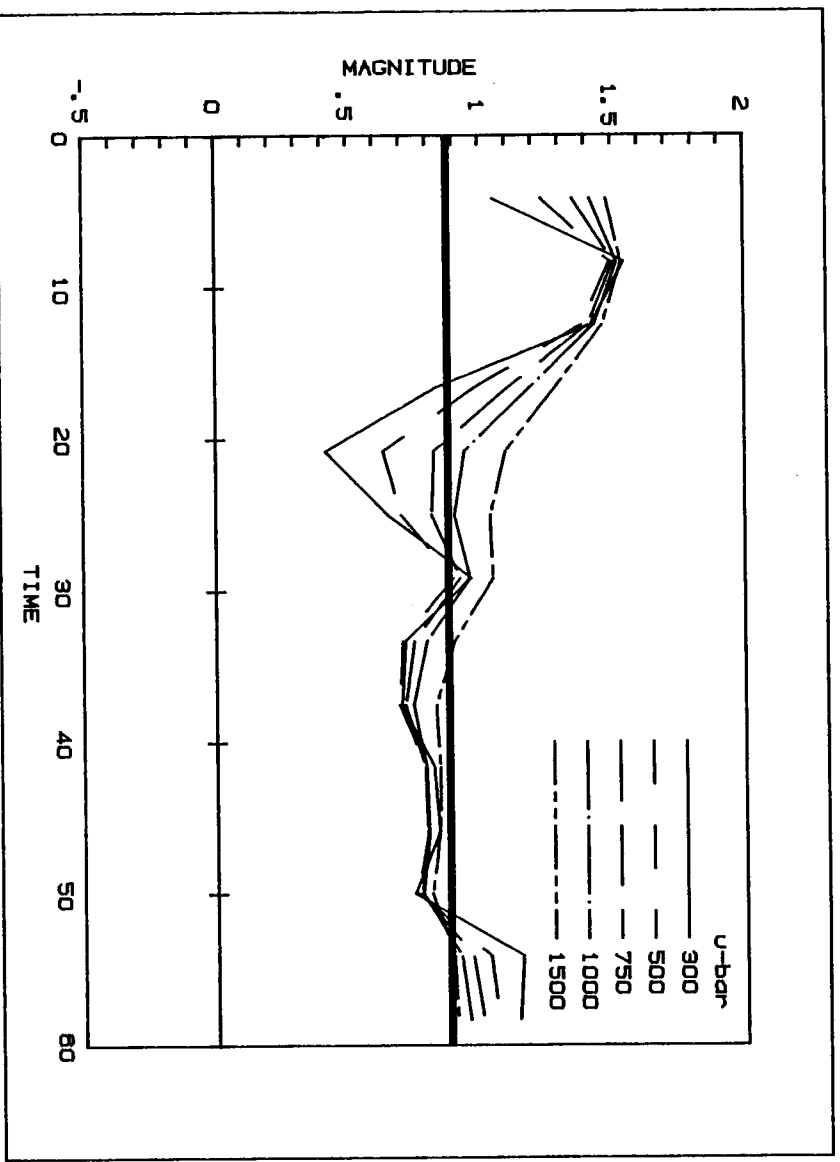


Figure 9. $\bar{U}$ -bar versus Accumulation Time

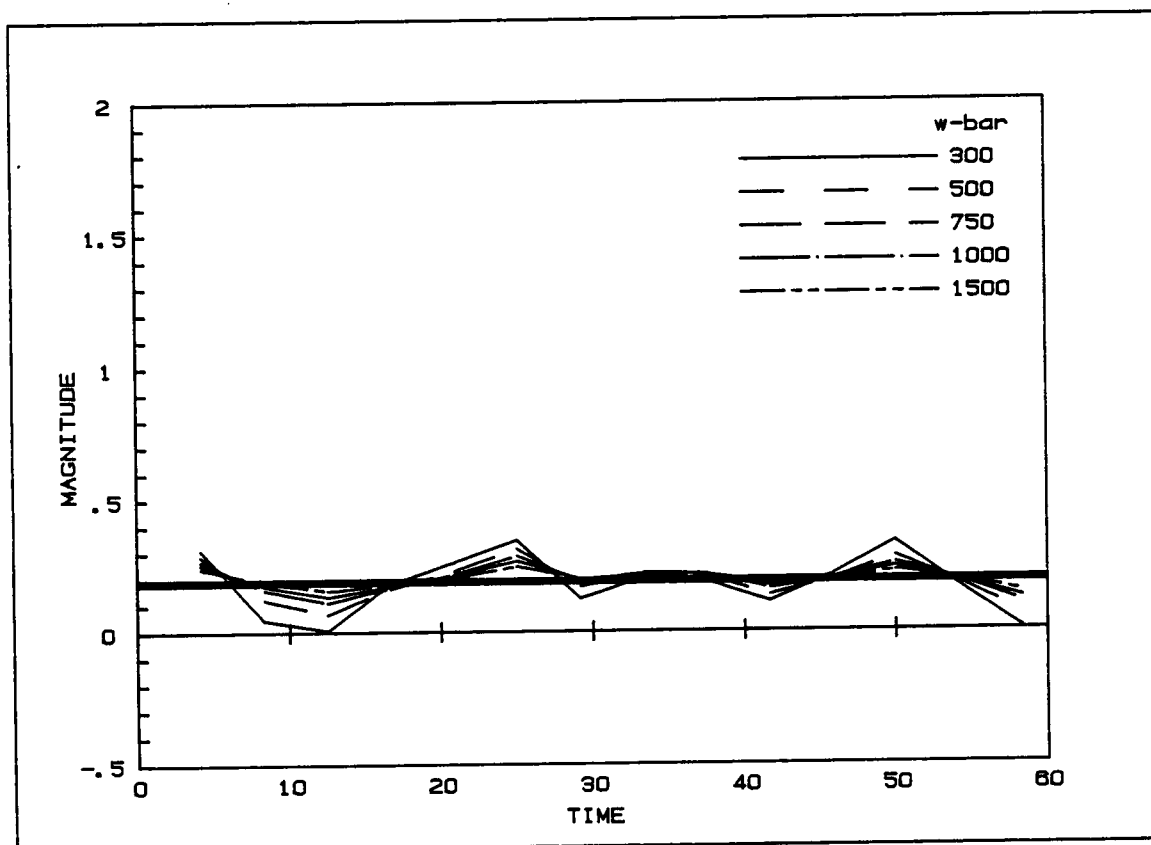


Figure 10. $\bar{W}$ -bar versus Accumulation Time

included simply as an illustration of the running means which are subtracted from the raw data to derive the fluctuating components of the turbulence.

Figures 11 through 16 represent the covariances calculated by subtracting the filtered velocity components described above from the raw data and then forming the variances and cross-products. The horizontal axis is the time that the cross-product has been accumulating, or the time over which the covariance is averaged. The individual lines correspond to different mean removal time constants, and can be specified independently from the accumulating time. In these figures, the heavy horizontal line is the hourly average of the covariance, evaluated in the normal manner described in the previous section. The hourly averages were calculated on a mainframe computer by first applying a three-rotation coordinate transformation, calculating the fluctuating components, and then forming and averaging the cross-products.

The real-time cross-products containing a vertical component, w' , show little sensitivity to the time constant of the filter or to the accumulation time. The cross-products which do not contain w' are more sensitive to both the time constant of the filter and the accumulation time, but are stable after about 40 minutes of accumulation. Fortunately, the $\overline{u'w'}$ term, or Reynolds stress, shown in

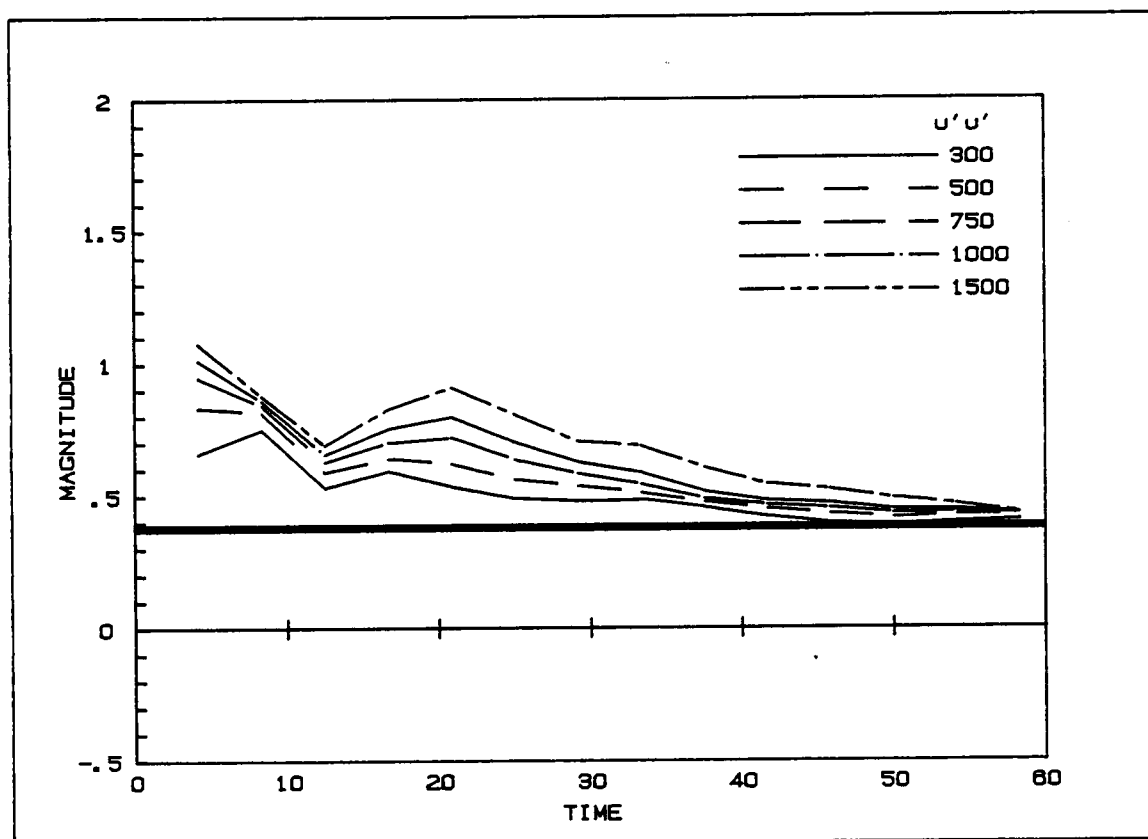


Figure 11. Variance of $u'u'$ versus Accumulation Time

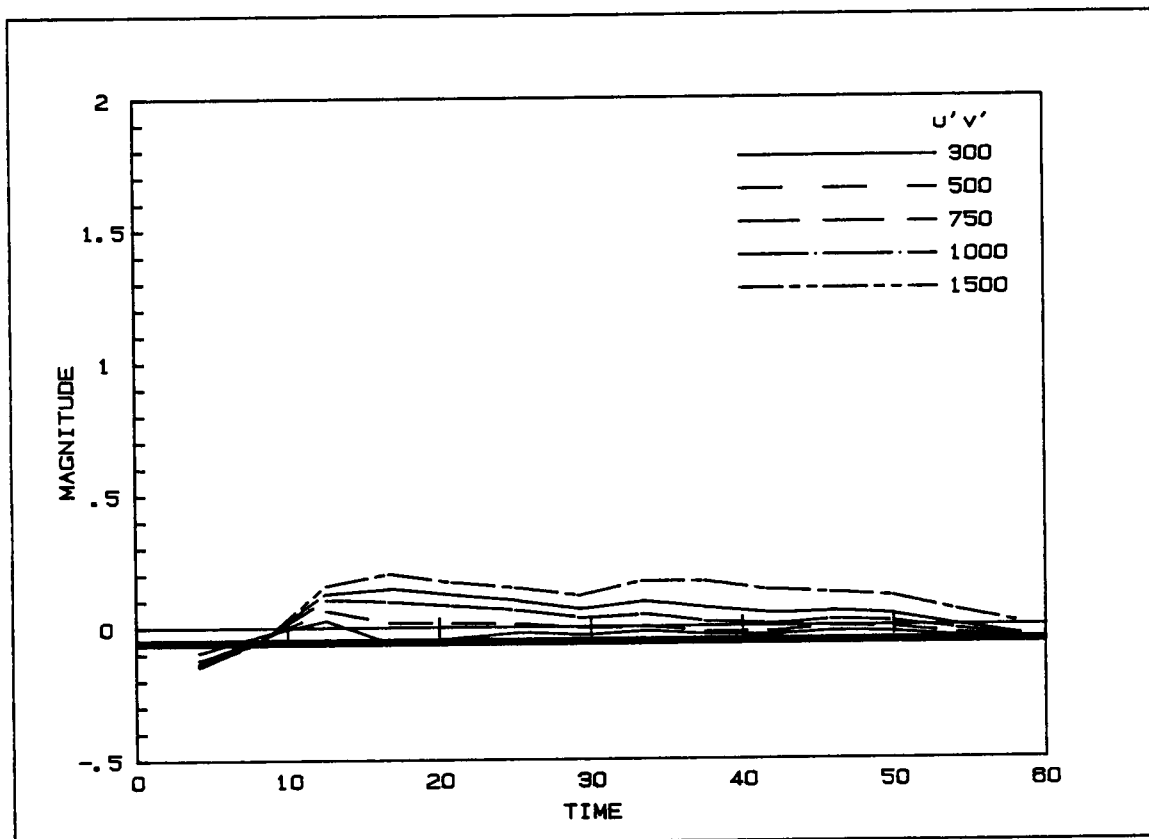


Figure 12. Covariance of $u'v'$ versus Accumulation Time

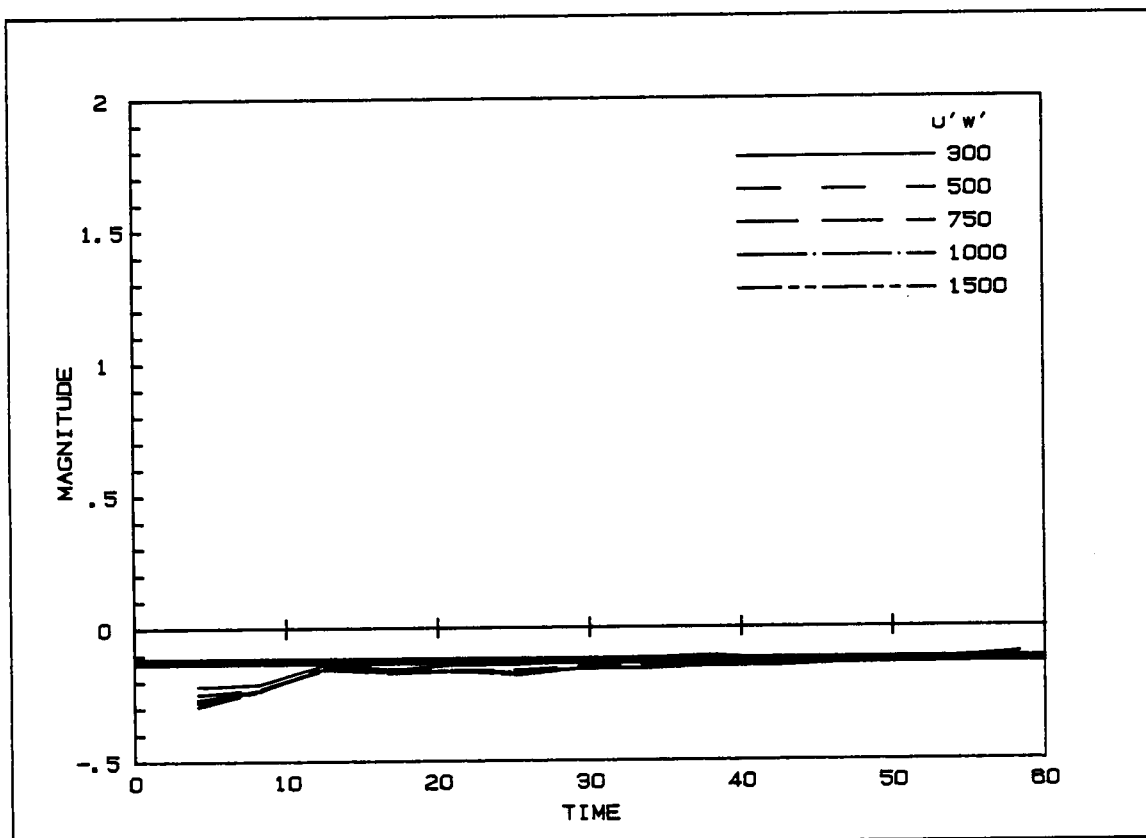


Figure 13. Covariance of $u'w'$ versus Accumulation Time

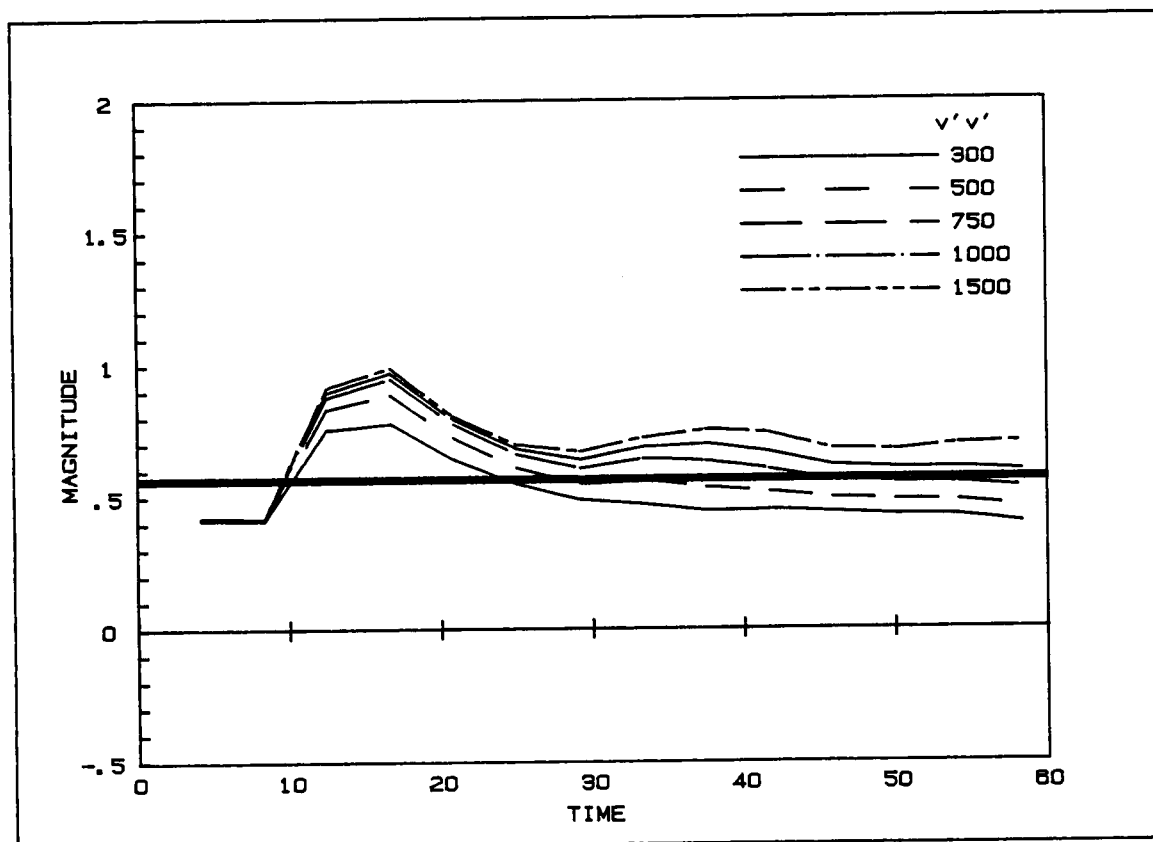


Figure 14. Variance of $v'v'$ versus Accumulation Time

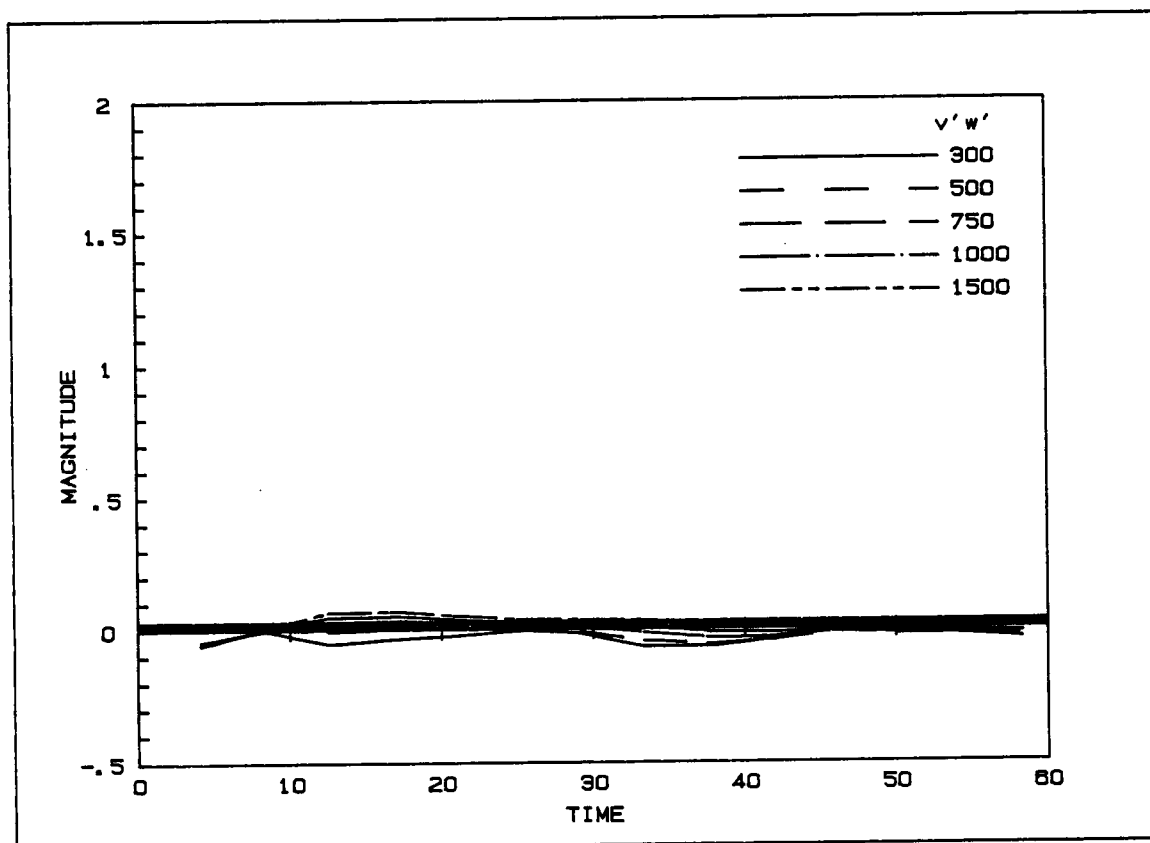


Figure 15. Covariance of $v'w'$ versus Accumulation Time

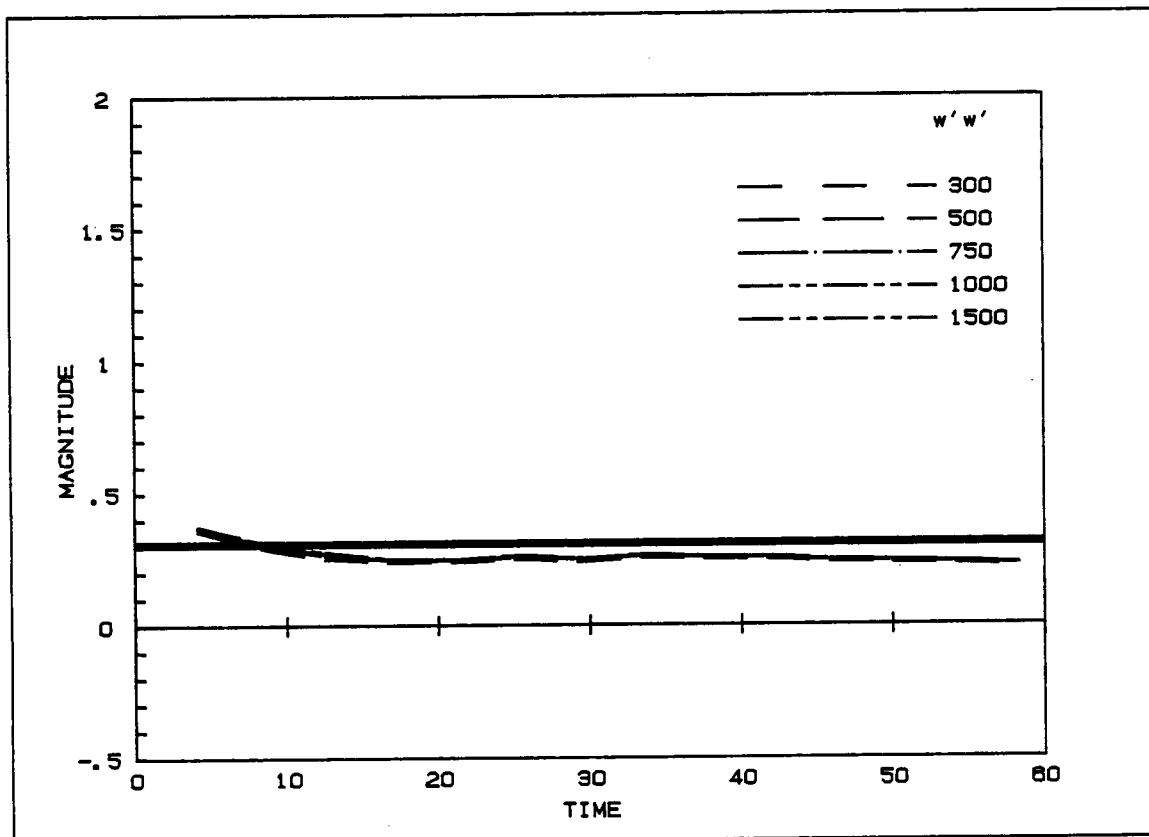


Figure 16. Variance of $w'w'$ versus Accumulation Time

Figure 13, which is of greatest interest, is the most stable. Excellent results are obtained for all the filters for accumulation times greater than about 12 minutes.

CHAPTER V

CONCLUSIONS

The absence of scatter in the data and the close agreement between the values of turbulence parameters shown in Figure 5 with those of other researchers indicate that one can calculate stable turbulence parameters from anemometer data collected at sites in irregular terrain, provided that a coordinate rotation is performed on the data. While this technique is available, it suffers from several disadvantages. It is impossible to use separate averaging times for mean removal, coordinate rotation, and averaging time for the covariances. It is also impossible to obtain the results in real-time without a mainframe or very fast minicomputer. These problems limit the usefulness of the technique.

A viable means of computing turbulent variances and covariances in real-time was demonstrated with good results. As Figure 13 shows, the quantity of primary interest, $\overline{u'w'}$, is very stable and is an excellent approximation of the value obtained by the more conventional means. At present, this technique has been used in three field studies of dry pollutant deposition with good success.

REFERENCES

REFERENCES

- Boussinesq, J. 1901. Physique Mathematique - Sur la pouvoir refroidissant d'un courant liquide ou gazeux. C. R. Acad. Sci. 133, 257.
- Businger, J. A., J. C. Wyngaard, Y. Izumi and E. F. Bradley. 1971. Flux profile relationships in the atmospheric surface layer. J. Atmos. Sci. 28, 181-189.
- Enochson, L. D., and R. K. Otnes. 1968. Programming and Analysis of Digital Time Series Data. (The Shock and Vibration Information Center, U.S. Department of Defense). 277 pp.
- Dyer, A. J., and B. B. Hicks. 1972. The spatial variability of eddy fluxes in the constant flux layer. Quart. J. Roy. Meteorol. Soc. 98, 206-212.
- Dyer, A. H., B. B. Hicks, and K. M. King. 1967. The Fluxatron - a revised approach to the measurement of eddy fluxes in the lower atmosphere. J. Appl. Meteorol. 6, 408-413.
- Dyer, J. J., B. B. Hicks and V. Sitaraman. 1970. Minimizing the levelling error in Reynolds stress measurement by filtering. J. Appl. Meteor. 9, 532-534.
- Dyer, A. H., and F. J. Maher. 1965. The "Evapotron": an instrument for the measurement of eddy fluxes in the lower atmosphere. CSIRO Div. Met. Phys. Tech. Pap. No. 15.
- Gill, G. C. 1975. Development and use of the Gill UVW anemometer. From Preprint Volume, 3rd Symposium on Meteorological Observations and Instrumentation, Amer. Meteor. Soc., Boston, MA.
- Hanna, S. R., G. A. Briggs and R. P. Hosker. 1982. Handbook on Atmospheric Diffusion. (Technical Information Center, U. S. Dept. of Energy). 102 pp.
- Haugen, D. A. (ed). 1973. Workshop on Micrometeorology. Amer. Meteor. Soc., Boston, MA. 392 pp.
- Hicks, B. B. and C. M. Sheih. 1977. Some observations of eddy momentum fluxes within a maize canopy. Boundary Layer Meteorol. 11, 515-519.

- Hicks, B. B. 1973. Eddy fluxes over a vineyard. Agric. Meteorol. 12, 203-215.
- Hicks, B. B. 1981. An examination of turbulence statistics in the surface boundary layer. Boundary Layer Meteorol. 21, 389-402.
- Hicks, B. B. 1972. Propeller anemometers as sensors of atmospheric turbulence. Boundary Layer Meteorol. 3, 214-228.
- Hicks, B. B. and M. L. Wesely. 1981. Heat and momentum transfer characteristics of adjacent fields of soybeans and maize. Boundary Layer Meteorol. 20, 175-185.
- Horst, T. W. 1973. Corrections for response errors in a three-component anemometer. J. Appl. Meteorol. 12, 716-725.
- Hyson, P., J. R. Garrat and R. J. Francey. 1977. Algebraic and electronic corrections of measured uw covariance in the lower atmosphere. J. Appl. Meteorol. 16, 43-47.
- Jones, E. P. and S. D. Smith. 1978. The air density correction to eddy flux measurements. Boundary Layer Meteorol. 15, 357-360.
- Kaimal, J. C., and D. A. Haugen. 1971. Comments on "Minimizing the leveling error in Reynolds stress measurement by filtering". J. Appl. Meteorol. 10, 337-339.
- Kaimal, J. C., and D. A. Haugen. 1969. Some errors in the measurements of Reynolds stress. J. Appl. Meteorol. 8, 460-462.
- Leuning, R. and P. M. Attiwill. 1978. Mass, heat and momentum exchange between a mature Eucalyptus forest and the atmosphere. Agric. Meteorol. 19, 215-241.
- Shaw, R. H., R. H. Silversides and G. W. Thurtell. 1974. Some observations of turbulence and turbulent transport within and above plant canopies. Boundary Layer Meteorol. 5, 429-449.
- Sutton, O. G. 1953. Micrometeorology. McGraw-Hill Book Company, Inc. New York, NY. 333 pp.

- Swinbank, W. C. 1955. An experimental study of eddy transports in the lower atmosphere. CSIRO Aust. Div. Met. Phys., Tech. Pap. No. 2.
- Swinbank, W. C. 1965. The measurement of exchanges between the atmosphere and the underlying surface. Methodology of Plant Eco-physiology: Proceedings of the Montpellier Symposium. 83-86.
- Wesely, M. L. 1974. Magnitudes of the turbulent fluctuations in the atmospheric surface layer. From Preprint Volume, Symposium on Atmospheric Diffusion and Air Pollution. Amer. Meteor. Soc., Boston, MA.

VITA

Robert Tilman McMillen, Jr. was born in Knoxville, Tennessee on August 8, 1941. He attended elementary schools in that city and was graduated from Knoxville Catholic High School as Valedictorian in June, 1959. The following September he entered Princeton University. In 1963 he transferred to the The University of Tennessee, Knoxville, and in December 1965 he received the Bachelor of Science Degree in Engineering Physics.

He was employed by Computer Sciences Corporation in Huntsville, Alabama for two years. In September, 1968, he entered the Graduate School of the The University of Tennessee, Knoxville.

He was employed by Smoky Mountain Outfitters from 1972 until 1976. Since 1976 he has been employed by the National Oceanic and Atmospheric Administration's Atmospheric Turbulence and Diffusion Laboratory, in Oak Ridge, Tennessee, as a research scientist.

He received the Master of Science degree with a major in Physics in June, 1983. He is a member of the American Association for the Advancement of Science and the American Meteorological Society.

He is married to the former Ingrid Lydia May of Heidleberg, Germany, and they have two daughters, Catherine and Elizabeth.