NOAA Technical Memorandum ERL BOMAP-9

U.S. DEPARTMENT OF COMMERCE
National Oceanic and Atmospheric Administration
Environmental Research Laboratories

The BOMEX Sea-Air Interaction Program: Background and Results to Date

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CONTENTS

ABSTRACT

The Barbados Oceanographic and Meteorological Experiment (BOMEX) Sea-Air Interaction Program has been the outgrowth of two lines of planning, begun in the early 1960's, in which the American Meteorological Society, the National Academy of Sciences, and Federal agencies have played important parts. One line was directed towards a national effort on the scientific problems of sea-air interaction. The other was directed towards an international effort to extend the range of meteorological forecasts. The Sea-Air Interaction Program, or "Core Experiment," of BOMEX ruled the layout of the ship array, scheduling of observations, and deployment of aircraft during the first three BOMEX Observation Periods: May 3-15, May 24-June 10, and June 19-July 2, 1969. It was designed to provide data on the sea-air flux of energy by three methods: (1) measurement of atmospheric budget terms over a 500-km square; (2) direct measurement of surface-layer vertical eddy fluxes at various times and places within the square; and (3) measurement of the major terms of the heat budget of the upper ocean at each of the ship stations. The momentum flux was also to be evaluated by the first two methods. After a long period of data reduction, some estimates of the evaporation rate, stress, sensible heat flux, and kinetic energy flux have now been obtained by each of the methods. Preliminary estimates of the evaporation rate are typically about 5 mm/day during early May, based on data from several investigators on the Navy's Floating Laboratory Instrument Platform (FLIP), and from turbulence measurements made on NOAA's Research Flight Facility (RFF) DC-6 aircraft. By late June, when FLIP was no longer in the BOMEX array, evaporation rates of at least 6 mm/day are indicated by the RFF aircraft data and preliminary volume budget analyses. Previous estimates were 4 to 5 mm/day based on Budyko's and Jacobs' climatological analyses for this season. Stresses measured by various methods range from a few tenths to more than 1 dyne/cm^2.

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1. INTRODUCTION

The Barbados Oceanographic and Meteorological Experiment (BOMEX) was a multiagency national research project, managed by the Environmental Science Services Administration (now incorporated in the National Oceanic and Atmospheric Administration) as lead agency, and with the cooperation of the Government of Barbados (Davidson, 1968; Kuettner and Holland, 1969). Data were collected during May, June, and July 1969 by instruments mounted on ships, aircraft, buoys, balloons, and satellites and on the island of Barbados (BOMAP Office, 1971b). The observational period of the BOMEX Sea-Air Interaction Program was May 1 through July 2. During that period the observations were concentrated on a 500-km square east of Barbados (fig. 1). During the last 3 weeks of July the observational array was changed to support the BOMEX Tropical Convection Program.

The concept of BOMEX responded very closely to requirements that arose through two different streams of planning.

The first stream was concerned with advancing our understanding of the interaction of the atmosphere and ocean. The requirement for "area studies" was articulated in the recommendation of the National Academy of Sciences Joint Panel on Air-Sea Interaction, reporting to both the Committee on Atmospheric Sciences and the Committee on Oceanography (Benton, 1962):

> "Recommendation III proposes the establishment of supporting facilities in special areas, at which field tests can be undertaken as new theories are formulated. New measurement techniques should also be tested and compared with other available methods. In these programs, improved geographic knowledge of the specific areas will be of minor importance compared with increased understanding of geophysical processes.
>
> "The areas should be chosen for their proximity to the continental United States and the ease with which personnel and equipment can be moved to the site in question; for the geographic isolation of the region from the surrounding areas, so that appropriate area balance techniques can be employed in measuring exchange phenomena; and for the availability of a good supporting network of meteorological and oceanographic stations. At least one oceanic region should be selected with these criteria in mind. One of the Great Lakes is also recommended."

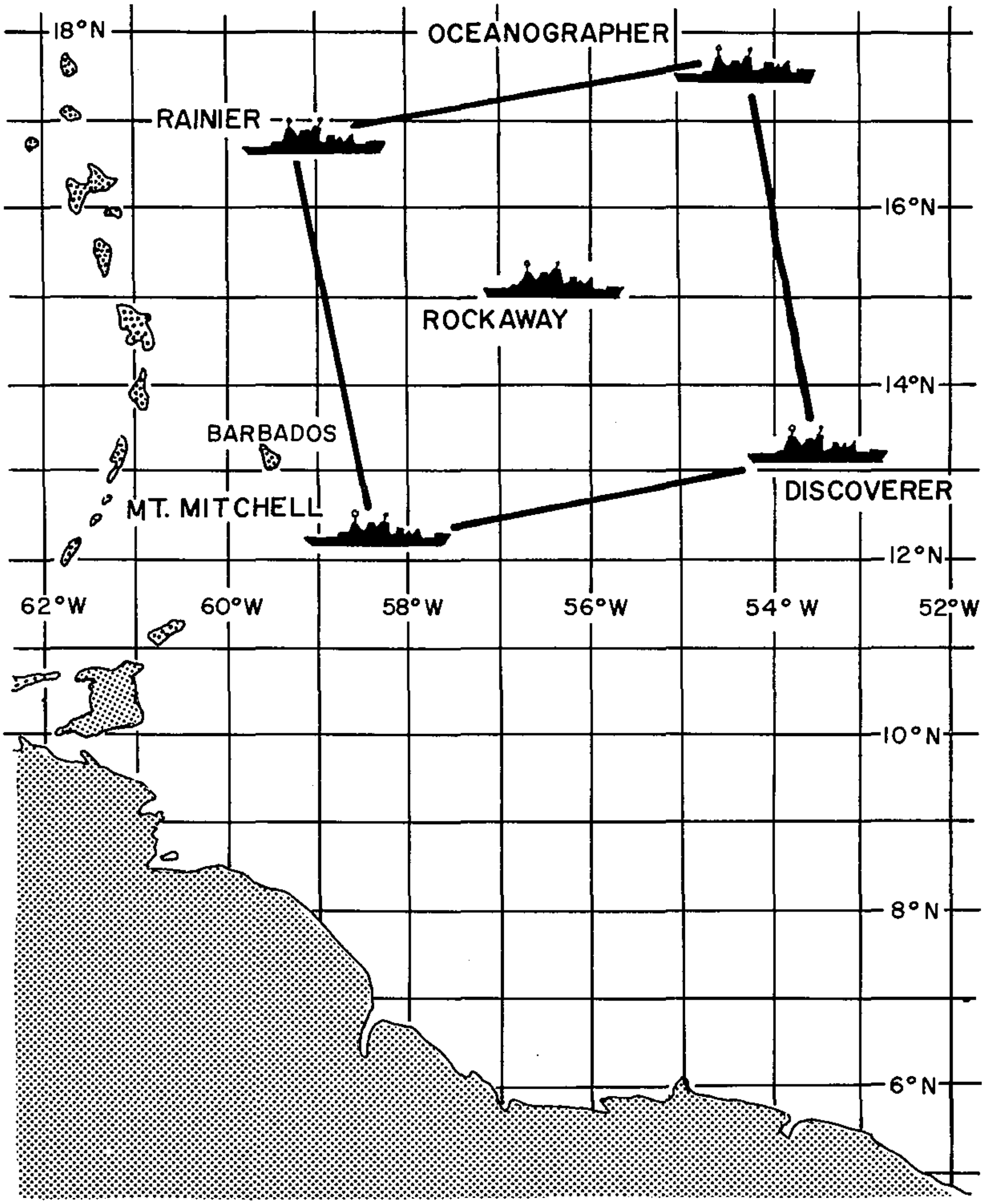

Figure 1. BOMEX ship array during the first three Observation Periods, May 3-15, May 24-June 10, and June 19, July 2, 1969.

The second stream was concerned with developing future prediction techniques and observation networks. The requirement for regional experiments is specified in the following excerpts from the report of the Panel on International Meteorological Cooperation of the National Academy of Sciences Committee on Atmospheric Sciences (Charney, 1966):

"In the development of techniques of middle- and long-range forecasting, and the formulation of mathematical models of the atmosphere suitable for the successful analysis of problems of large-scale weather and climatic modification, smaller-scale processes must be incorporated into our theory of atmospheric motions. These processes include not only the turbulent exchange between the atmosphere and the underlying surface of such quantities as momentum, heat, and water vapor, but also the redistribution of these quantities within the atmosphere itself due to turbulent motions on the meso- and micro-scale. At present, only crude quantitative estimates of turbulent exchange resulting from smaller-scale motions are available. This represents an important deficiency in our knowledge of large-scale weather prediction and modification."

* * * * * * *

"The macro-scale variables with which internal exchange and energy-transformation processes are to be related have yet to be formulated. Quantities to be examined will obviously include the field of mean vertical motion, and macro-scale variations of temperature, water vapor, and wind with height. The use of satellite information on cloud distribution and incoming and outgoing radiation will provide valuable information required to establish the validity of empirical relationships. Radar measurements of condensation processes will also be useful."

* * * * * * *

"Boundary fluxes must be obtained which are compatible with the macro-scale observing system, that is, which are representative of areas of about 250,000 km^2 and of time intervals of 3 to 12 hr. The primary objective of the research investigations must be to relate the boundary fluxes to the macro-scale observations. The most fruitful approach is likely to be both to extend fundamental understanding of the physics of the transfer processes and to develop empirical relations between the global observations and average fluxes. Fundamental understanding and derivation of empirical relations can be sought simultaneously through a carefully planned program in which a variety of independent methods are used to compute the vertical fluxes."

* * * * * * *

"Ultimately it is clear that some way must be found to express the influences of meso- and micro-scale systems in terms of synoptic and general-circulation-scale parameters. Hopefully, it will be possible to do this, but only after the energy fluxes and conversions on all scales and their interactions have been investigated quantitatively with adequate data."

In September 1968, the late Ben Davidson, then Scientific Director of BOMEX, published the following statement of the scientific objectives of BOMEX (Davidson, 1968):

"On a global scale the atmosphere as an isolated system loses heat radiatively at a rate that is roughly independent of latitude (Sellers, 1965). The major energy input (either as latent or sensible heat) into the atmosphere comes at the interface between the atmosphere and ocean. It is this energy input which varies with latitude and is the source for the large scale motions of the atmosphere. And yet, oddly enough, the energy enters the atmosphere through the action of small scale turbulence operating in a local field of mean vertical gradients which are maintained by the boundary conditions at the interface. Moreover, these small scale turbulent motions are the basic mechanism for the dissipation of mechanical energy and for the transfer of momentum to the sea surface. To understand and predict the large scale motions of the atmosphere and ocean on an extended time scale it is necessary to know the effective energy inputs and dissipations associated with the smallest scale motions of the fluid systems and to relate these inputs to feasible synoptic measurements in the weather and oceanographic systems of the future.

"The specific objectives concerning air-sea interactions are:
a) Study of the vertical flux of momentum, sensible heat, latent heat, radioactivity, and other properties at the interface and the horizontal transport of these properties through the lateral boundaries of the observational array.
b) Study of the vertical and horizontal divergence of these fluxes within the interior of each fluid.
c) Study of the feasibility of parameterizing the area-wide integral of at least the surface fluxes from conventional observation at the fixed corners of the array."

In that article Dr. Davidson described the project as of mid-1968. Subsequently, many additional research projects were incorporated in BOMEX. Because of the responsiveness of the basic design of the grid to the widely felt requirement of relating small-scale phenomena to synoptic-scale meteorological and oceanographic conditions, and because of the emphasis Dr. Davidson gave to the development of a coherent air-sea interaction program, BOMEX attracted virtually the entire U.S. population of scientists actively engaged in this area of endeavor. In particular, many of the scientists who participated in the earlier studies, such as those of the NAS-NRC panels quoted above, participated in BOMEX.

In addition, projects that primarily took advantage of BOMEX to pursue objectives solely related to oceanography, to the verification of satellite data, or to other agency missions, also contributed unique and needed inputs to the objectives of BOMEX and thus made BOMEX a more complete and comprehensive experiment. This is particularly true of BOMEX in its aspect as a test of methods. Because nothing approaching the complexity of BOMEX had been attempted before in this field, many questions of feasibility or relative merit of methods had to be answered before a definitive experiment on this scale could be designed.

The methods to be tested include *parameterization methods* and *measurement methods*.

The *parameterization methods* derive the source terms and fluxes for the various forms of energy from observed or predicted synoptic-scale averages and gradients of measured variables. First the validity of the assumptions underlying the simplified mathematical relationships must be tested against measurements. If the forms of the equations prove valid, some of the coefficients appearing in them must be determined empirically, or the correctness of values predicted from theory must be verified. To do this will ultimately require measurement programs under a wide range of climatic conditions.

Measurement methods for developing or testing these parameterization methods must, in turn, be developed, tested, and evaluated. The oceanic area east of Barbados has the following advantages for such a test and evaluation experiment:

(a) It lies within the vast equatorial oceanic belt that contributes a major part to the global atmospheric and oceanic energy budget and thus represents a high-priority climatic regime.

(b) It is relatively accessible to the United States.

(c) Of the many climatic regimes that would, in principle, have to be sampled, this area, during late spring and early summer, provides a relatively narrow range of environmental variations and thus maximizes the significance with which small differences between large quantities can be determined statistically by repeated observations.

Clearly the requirement for wide-range applicability for the ultimate product and narrow range of experimental conditions for the first test of methods are incompatible in an experiment of limited area and duration such as BOMEX. Thus conclusions regarding feasibility and limits of error for the measurement methods should be reliable, and the energy fluxes, source terms, and conversion rates determined in BOMEX will be accurate for the time and place of the experiment. On the other hand, they will not provide a reliable basis for extrapolation to other times and places, although they can provide a limited check on parameterization formulas.

2. THE EXPERIMENT

The design of the experiment, an expanded version of that used in the Indian Ocean Expedition (Fleagle et al., 1967), has been described in an earlier report (Holland, 1970), and will only be summarized briefly here. Three independent methods are being used to obtain estimates of the sea-air flux of energy. Two of these also give the stress.

First, the atmospheric budget method employs surface, rawinsonde, and radiometersonde data from the five fixed ships at the corners and center of the 500-km square, aircraft measurements along the perimeter of the square, dropsonde data in the interior of the square, and radar and satellite coverage. Based on these data, estimates of the horizontal flux divergence, vertical flux due to the mean vertical velocity, rate of change of storage, and sources or sinks of water vapor, heat, and momentum are evaluated for a series of 10-mb layers from sea level to a surface on which the pressure is 500 mb below sea level pressure. These terms are combined in the budget equations (Rasmusson, 1971a) to derive the sea-air flux by vertical integration of the vertical eddy flux divergence, which is obtained as the residual of each budget equation.

Second, the ocean heat budget method uses salinity-temperature-depth (STD) soundings at the five fixed ships together with solar and net radiation measurements on the ship booms to evaluate the net radiative input and rate of change of heat storage in the upper mixed layer of the ocean. While no direct measurements of current velocity are available to permit computation of the horizontal advection, there is sufficient indirect evidence to place upper limits on this term. Vertical flux through the lower integration boundary is minimized by choosing a depth in the thermocline at which the diurnal temperature variation is vanishingly small. Additional information bearing on the range of vertical eddy diffusion is available from the diurnal variation of the vertical temperature and salinity profiles and from the radionuclide experiments (Broecker and Peng, 1971; Young and Silker, 1971; Schink et al., 1970).

The third method is to measure directly the vertical fluxes of water vapor, heat, and momentum in the surface layer ("constant flux layer") of the atmosphere. A number of experiments of this type were included in BOMEX.

On the Navy's Floating Laboratory Instrument Platform (FLIP) turbulent fluctuations of velocity were measured by Gibson and Stegen (Gibson, Stegen, and Williams, 1970) of the University of California at San Diego using hot wires, by Stewart and Miyake of the University of British Columbia using a sonic anemometer (Pond et al., 1971), and by Portman and Davidson of the University of Michigan using hot film and hot-wire anemometers (Portman et al., 1970). Pond of Oregon State University measured humidity fluctuations with a Lyman alpha hygrometer and temperature fluctuations with a resistance thermometer (Pond et al., 1971). Additional series of fast-response temperature measurements, using a thermocouple, were made by Fleagle and Paulson of the University of Washington, who also measured vertical profiles of wind, temperature, and humidity during the first 2 weeks of May 1969 (Paulson et al., 1970). During the last 2 weeks of May, another series of wind, temperature, and humidity profile measurements was carried out by Superior of Thornthwaite Associates (Superior, 1969). Measurements of turbulent wind velocity fluctuations were also made by Elderkin of Battelle Northwest Laboratories on the Florida State University Triton buoy, by Frenzen of Argonne National Laboratory on the Tern buoy, and by Deleonibus of the Naval Oceanographic Office on the USNS *Gilliss*.

Low-level aircraft measurements of turbulent fluxes were made by Bean of NOAA's Wave Propagation Laboratory on a DC-6 aircraft of NOAA's Research Flight Facility using an angle-of-attack vane and microwave refractometer (Bean et al., 1971); by Miyake and Donelan of the University of British Columbia using sonic and hot-wire anemometers, a thermistor, and a Lyman alpha humidiometer mounted on a Queen-Air airplane of the National Center for Atmospheric Research (Miyake, Donelan, and Mitsuta, 1970); and by Bunker of Woods Hole Oceanographic Institution on a DC-4 equipped with angle-of-attack vane and wet and dry resistance thermometers (Bunker, 1970).

The results of preliminary atmospheric and oceanic budget analyses are being reported elsewhere and will only be summarized here. The boundary-layer flux measurements will be discussed more fully.

3. ATMOSPHERIC BUDGET RESULTS

A trial integration of the budget equations has been carried out for the 5-day period June 22-26, 1969 (Holland and Rasmusson, 1971) based on surface and rawinsonde data at the four corner ships of the BOMEX array. A total of 233 soundings were available during the selected period, which was characterized by typical undisturbed tradewind weather (see, for example, BOMAP Office, 1971a). The 5-day precipitation was only about 1 mm as estimated from the rain-gage, radar, and satellite data (Hudlow, 1971; Scherer and Hudlow, 1971).

Table 1 gives the 5-day average vertical fluxes of water vapor, latent heat, sensible heat (enthalpy), kinetic energy, and momentum at the sea surface from the atmospheric budget analyses, the standard deviation of each flux based on 77 values, computed for each 1 1/2-hourly observation time, and an estimate of the standard error of the 5-day mean, based on the assumption that all the variance is due to random, uncorrelated errors. These error estimates must be taken as rough, preliminary values pending further analysis. In fact, it will be seen later that real variations of a factor of two in evaporation rate apparently do occur over periods of a few days, which would cause this standard error estimate to be too high. On the other hand, the individual values were not strictly independent because some time filtering had been done, so that the proper degrees of freedom would be less than 77.

The Bowen ratio (sensible heat flux/latent heat flux), based on this budget analysis is 0.086. In other words, during this undisturbed period 92 percent of the heat given up by the ocean was used to evaporate water, and 8 percent to heat the air directly. The total kinetic energy lost from the tradewind flow, both by boundary layer dissipation via turbulence and by transfer to the sea, was about 1/2 percent of the energy gained from the sea by evaporation and sensible heat transfer.

Table 1. Sea-air fluxes from atmospheric budgets, June 22-26, 1969

	5-day mean	Standard deviation	Standard error
Water vapor, mm day^{-1}	6.0	5.3	$\pm$ 0.6
Latent heat, cal cm^{-2} day^{-1}	349	308	$\pm$ 35
Sensible heat, cal cm^{-2} day^{-1}	30	260	$\pm$ 30
Kinetic energy*, ergs cm^{-2} sec^{-1}	-820	790	$\pm$ 90
Kinetic energy, cal cm^{-2} day^{-1}	-1.8	1.7	$\pm$ 0.2
Total energy, cal cm^{-2} day^{-1}	377	404	$\pm$ 46
Momentum, dyne cm^{-2}	-0.65	0.90	$\pm$ 0.10

*Includes boundary layer dissipation.

4. OCEAN HEAT BUDGET RESULTS

An upper-ocean heat budget analysis based on the fixed-ship STD and radiation data for the BOMEX third period, June 20 to July 2, 1969, has been carried out by Delnore (1971). The total energy transferred from the sea to the atmosphere can be determined by this method for the stations occupied by the *Discoverer*, *Rockaway*, and *Rainier*, which were equipped with pyranometers. The period was divided into two parts separated by a break in the observation series (maintenance and calibration day). The first 7-day period, June 20-26, starting 2 days earlier than that of the atmospheric budget analysis already mentioned, was a relatively undisturbed period, although the radar showed more convective shower activity on June 21 than during the following 5 days. The final 5-day period, June 28 to July 2, was considerably more disturbed, especially the first 2 days, during which a large convective system passed through the BOMEX array. For the whole period, 285 STD soundings were available, 179 in the first 7 days and 106 in the last 5 days.

Six separate determinations of the sea-air heat flux were made, one for each of the three ships for each of the two time periods. These and the averages for each period and for the entire period are shown in table 2. The standard deviation of the six values is 39 cal cm^{-2} day^{-1}. A rough estimate of the standard error of the mean based on the assumption that the variations are due only to random, uncorrelated errors of measurement, would be about $\pm$ 20 cal cm^{-2} day^{-1}.

Table 2. Sea-air sensible + latent heat flux from ocean heat budget (cal cm^{-2} day^{-1})

Ship	June 20-June 26	June 28-July 2	June 20-July 2
Discoverer	439	380	409
Rockaway	345	422	386
Rainier	332	360	346
Average	371	389	380

5. ATMOSPHERIC BOUNDARY LAYER FLUX MEASUREMENTS

The various BOMEX sets of atmospheric boundary layer flux data must be related to the height range and spectral bandwidth over which they were obtained, and to the vertical structure of the tropical atmosphere. Figure 2 shows mean profiles of potential temperature, specific humidity, and wind speed on a logarithmic height scale. There are discontinuities due to the fact that the data from FLIP (2 to 12 m), the Boundary Layer Instrument Package (BLIP) (10 to 300 m), and rawinsondes (10 to 6,000 m) do not represent identical time periods. The main points to be noted are that in the lowest few meters both potential temperature and humidity decreased approximately logarithmically with height, that there was a well-mixed layer about 600 m deep, and that above this the specific humidity decreased more or less strongly, while the potential temperature increased. These changes were greatest in the region of the tradewind inversion, 1,500 to 2,000 m. The tradewind speed maximum occurred at about 1,000 m during this period.

The range of horizontal scales for each data set, from the wavelength corresponding to the Nyquist (folding) frequency to the length corresponding to the mean wind (or aircraft) travel over the total duration of each data set, is shown against the height range of each set in figure 3.

During the first 2 weeks of May 1969, the teams from the University of California at San Diego (UCSD), University of British Columbia (UBC), Oregon State University (OSU) and University of Washington (UW) were aboard FLIP. During that time numerous gust probe runs were made close to FLIP by the

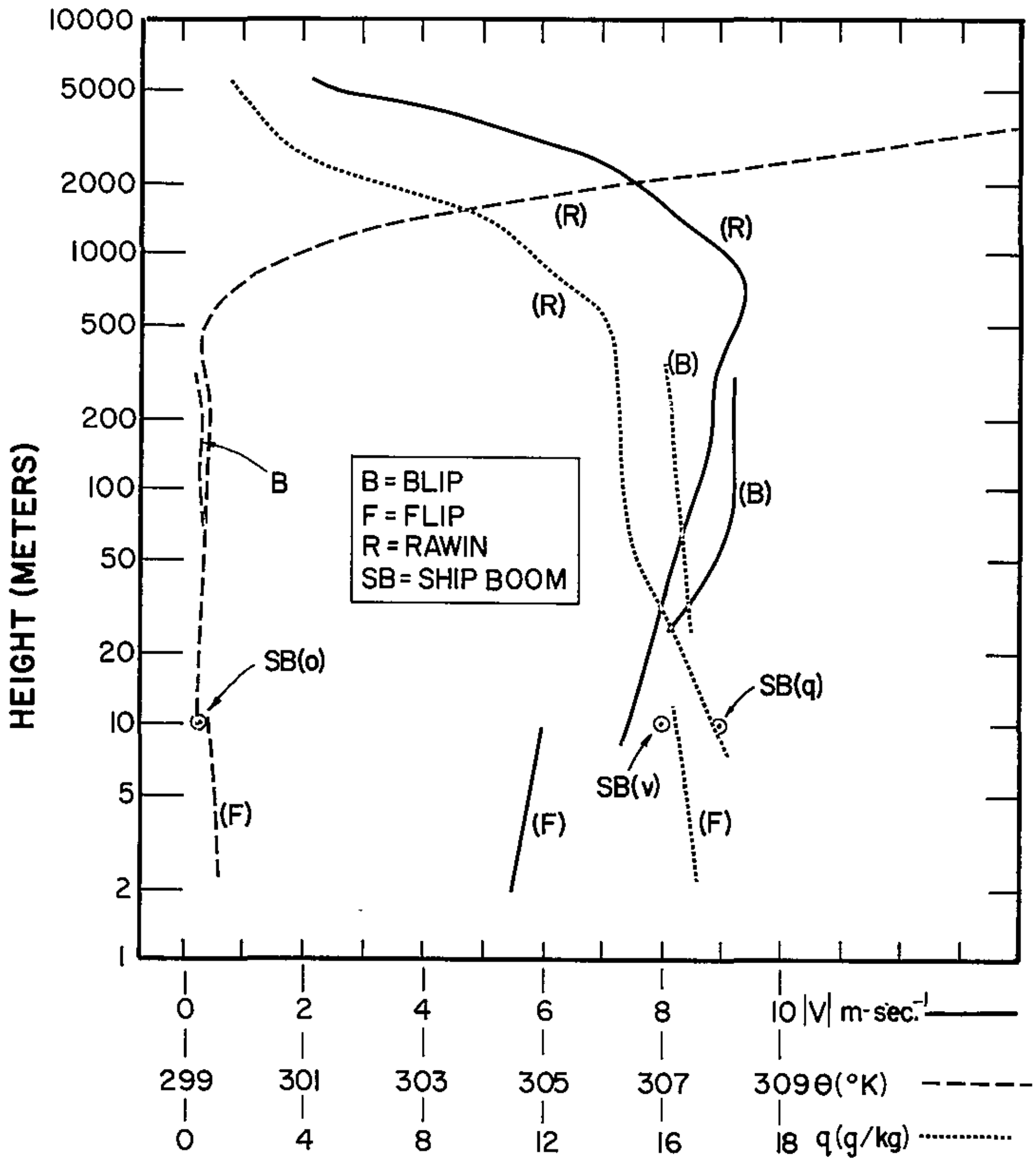

Figure 2. *Profiles of mean wind speed (V), potential temperature (θ), and specific humidity (q) on a logarithmic height scale. Segments for different altitude ranges obtained from FLIP (F), BLIP (B), and rawinsondes (R) are based on different time periods; their slopes are believed to be representative, but no attempt has been made to normalize their absolute values. Points at 10 m are based on ship boom (SB) data.*

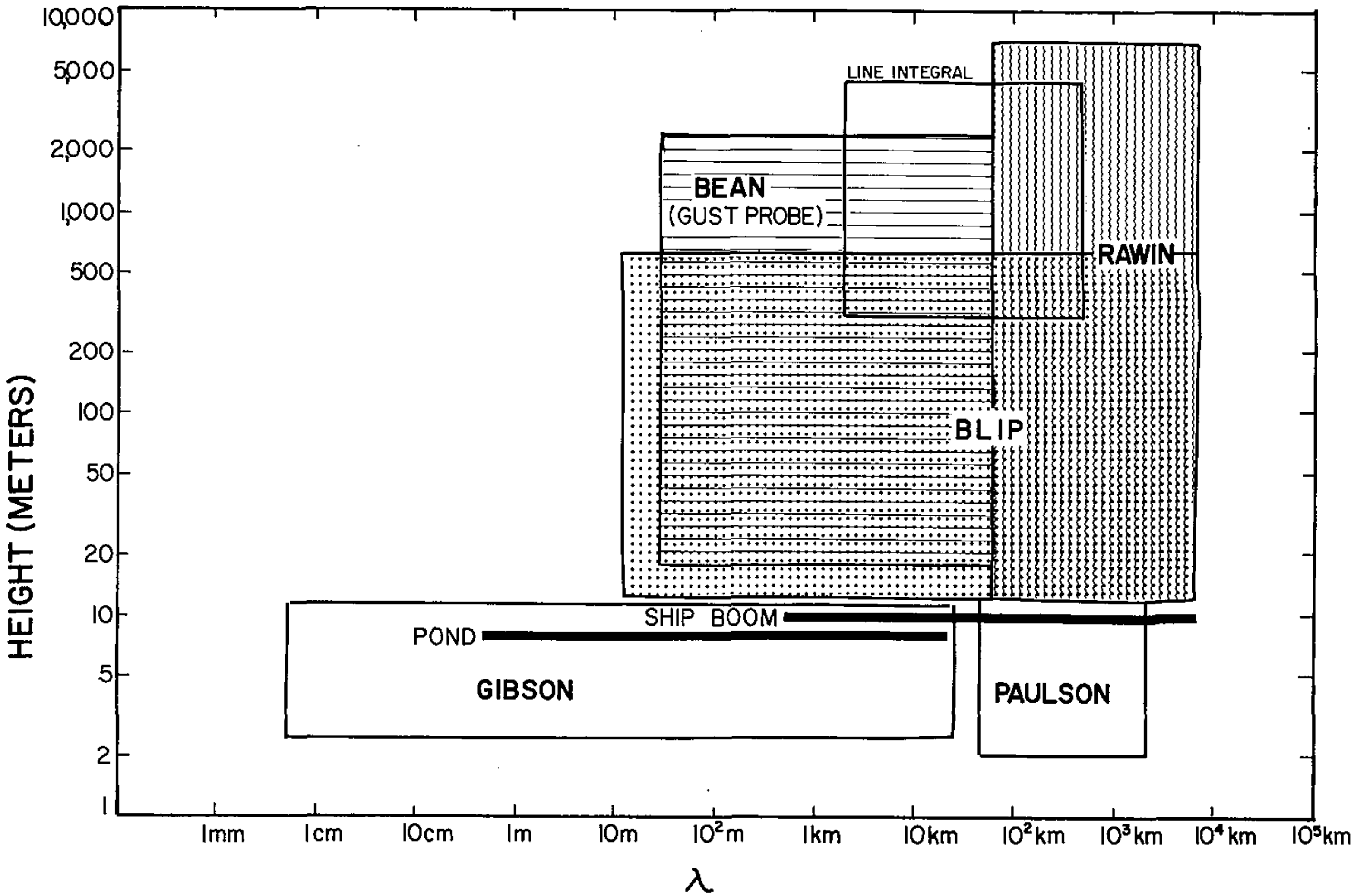

Figure 3. Wavelength (λ) band-width and height range of each measurement system discussed.

NOAA Research Flight Facility (RFF) DC-6 and by the NCAR Queen-Air carrying UBC turbulence instrumentation.

6. OSU-UBC TURBULENCE DATA

Figure 4, taken from Pond et al. (1971), shows the common logarithm of spectral density of the longitudinal (u), transverse (v) and vertical (w) components of wind velocity, measured by a sonic anemometer on FLIP. The spectral densities are normalized by multiplying by the frequency f and dividing by the variance of w. The abscissa is the logarithm of normalized frequency fz/U, where z is the height of measurement (varying from 8.1 to 8.6 m) and U the mean wind speed during each run. The 20 runs analyzed varied in duration from 25 to 75 min and include four that were made in the Pacific Ocean off San Diego in a pre-BOMEX trial.

The disturbance due to wave-induced motions of FLIP can be seen in all three components as a large excess of spectral density over a band of log-normalized frequency between -1 and -0.5. At higher frequencies all three normalized spectra fall off with a -2/3 slope corresponding to the Kolmogoroff -5/3 law (for un-normalized spectra), except that the w spectrum slope does not become this large until a log-normalized frequency of about 0.5 is reached, corresponding to a frequency of about 2 Hz or a wavelength of about 3 m.

The most interesting behavior of the spectra is that at lower frequencies. The w spectrum has its peak near the frequency of the wave effect, and decreases steadily with decreasing frequency. The u and v spectra have minima at frequencies below that of the wave effect, and rise to maxima at normalized frequencies below 10^{-2}, corresponding to wavelengths of the order of 1 to 2 km. The v component has less energy than the u component at wavelengths of the order of 100 m, but the v spectral peak is higher and occurs at a longer wavelength than that of u. On the face of it, the spectra would indicate near-isotropy at wavelengths of the order of 1 to 3 m, anistropic eddies with a predominance of energy in the u direction, i.e., primarily transverse vortices, at wavelengths in the neighborhood of 100 m, and possibly a predominance of longitudinal vortices at wavelengths of the order of a few km.

The average variance of the v component is also larger than that of u. Pond et al. (1971) consider this excess of v variance over u variance, concentrated in the low frequency portion of the spectrum, to be spurious, resulting from slow rotations of FLIP. However, in view of the possibility that such an effect could be associated with longitudinal vortices on this scale, any interpretation should perhaps be taken with reservations until it can be supported by additional evidence.

The OSU temperature and humidity spectra published by Phelps and Pond (1971) at 8 m on FLIP during the 16 BOMEX runs are reproduced in figure 5. The humidity spectrum closely resembles that of u without the FLIP motion effect. The temperature spectrum has its peak at normalized frequencies in the neighborhood of unity, and does not attain a -2/3 slope within the bandwidth of the data. This raises some doubt as to whether the inertial

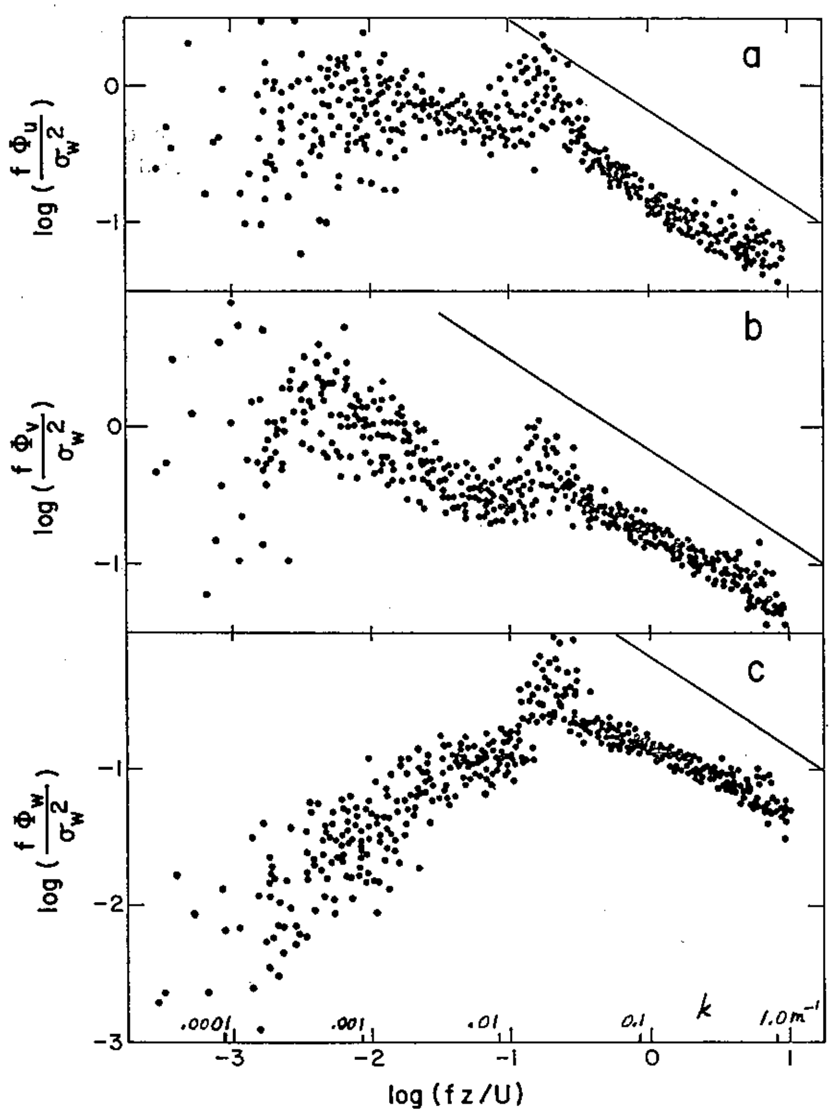

Figure 4. *Log_{10} of normalized spectral density of longitudinal (u), transverse (v), and vertical (w) components of velocity, as functions of log_{10} of nondimensionalized frequency; f is frequency (cycles sec^{-1}), ϕ_u, ϕ_v, ϕ_w are spectral densities, $\sigma_w{}^2$ is variance of w (used for normalizing all three components), z is height, and U is mean wind speed based on measurements by sonic anemometer at 8 m on FLIP (from Pond et al., 1971).*

subrange extends to wavelengths as large as 1 m at a height of 8 m over the sea, despite the slopes of the u, v, w, and humidity spectra in this region. It is well known that the slope characteristic of the inertial subrange extends to much lower wave numbers in one-dimensional spectra (Gifford, 1959).

The difference between the temperature and humidity spectra consists mostly of the absence in one and the presence in the other of fluctuations on the 100-m to several kilometer scales. This can be related qualitatively to the mean vertical profiles of θ and q shown in figure 2. The action of vertical eddy motions on the temperature field over height intervals of several hundred meters or more must be to transfer warmer air into a layer of minimum and nearly uniform potential temperature both from below and from above. This heat flux convergence must be compensated for by radiative loss at a sufficient rate to maintain the θ minimum, so that the potential temperature of air reaching the surface layer in the larger eddies cannot deviate much from that of the near-adiabatic mixed layer. Humidity differences, on the other hand, increase with the vertical distance over which they are taken, so that the action of the largest-scale vertical eddy motions on the mean humidity gradient produces the largest humidity fluctuations.

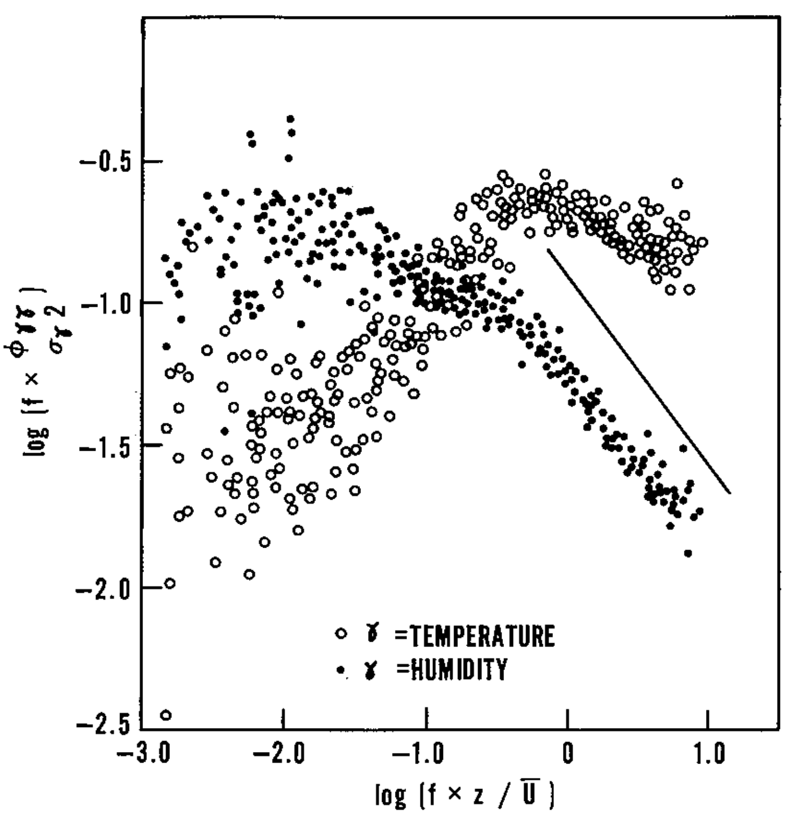

Figure 5. Log_{10} of normalized spectral density of temperature and humidity vs. log_{10} of nondimensional frequency at 8 m on FLIP (from Phelps and Pond, 1971).

7. SCRIPPS TURBULENCE DATA

The behavior of the temperature fluctuations at still higher frequencies has been studied by the group at Scripps Institution of Oceanography, UCSD (Gibson, Stegen, and Williams, 1970), based on measurements at heights of about 2, 4, 7, and 12 m on FLIP in the frequency band from 2 Hz to 2 kHz by means of a fine platinum resistance thermometer. Although analyses of only a small sample of their BOMEX data have been reported so far, it is evident that the Scripps group has documented the temperature and velocity spectra up to the viscous cutoff (Gibson, Stegen, and McConnell, 1970; Gibson, Stegen, and Williams, 1970). Their 2-m temperature derivative spectrum, reproduced in figure 6, had a slope of +1/3 from wavelengths of about 10 to 300 times the Kolmogoroff scale (which is about 1 mm). This is a necessary but not sufficient condition for an inertial subrange. Gibson and his coworkers found that the temperature derivative had a skewed distribution, opposite in sign to that of the u component of velocity, indicating anisotropy. They also found other unexpected properties of the temperature derivative spectra.

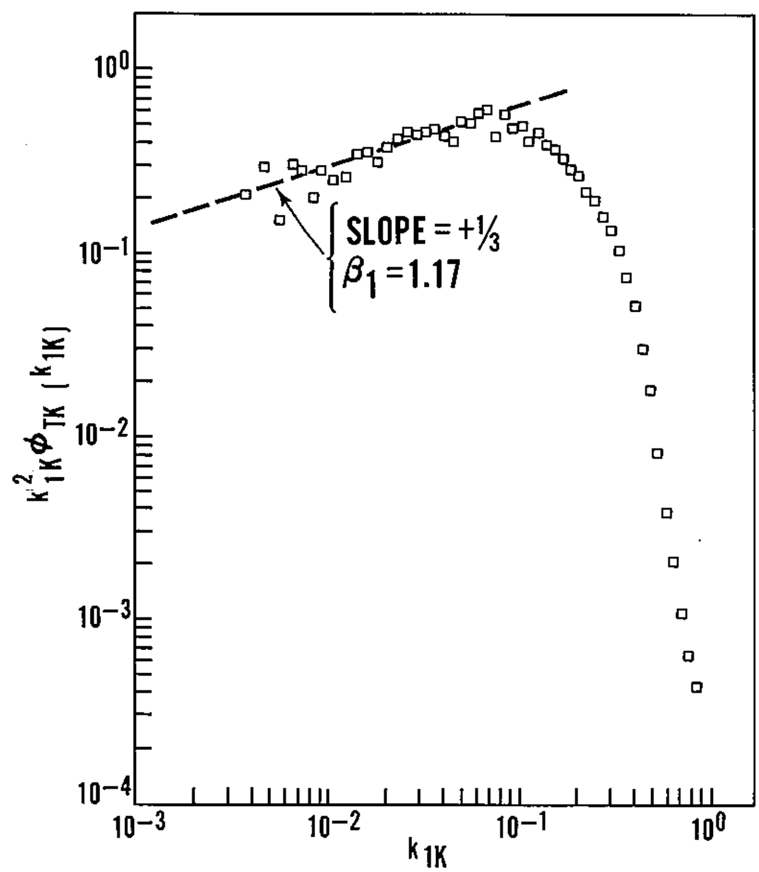

Figure 6. Log_{10} of spectral density of longitudinal spatial derivative of temperature vs. log_{10} of nondimensionalized wave number; k_{1K} is longitudinal wave number divided by Kolmogoroff scale based on measurements by UCSD fast resistance thermometer at 2 m on FLIP (from Gibson, Stegen, and Williams, 1970).

The anisotropy of the temperature derivative, and the possibility of a negative correlation of u and T suggest that some vertical eddy heat flux, and hence some production of temperature fluctuations, may occur within this band. From Phelps' and Pond's temperature spectra at a height of 8 m in the adjoining band, one might suspect that eddies with wavelengths of the order of 30 cm at a height of 2 m are not sufficiently far down the inertial cascade to fulfill the physical conditions of the Kolmogoroff hypothesis. A negative correlation of u and T could, of course, also arise from undulatory flow without vertical eddy transfer.

8. FLIP COSPECTRA

Figure 7, from Phelps and Pond (1971), shows normalized cospectra of w and T on a linear scale as a function of the log of normalized frequency from both the San Diego pre-BOMEX trials and BOMEX. The BOMEX cospectra do not fail to zero at the high frequency cutoff (wavelength about 1 m), suggesting that even at an altitude of 8 m the heat flux may indeed have extended into

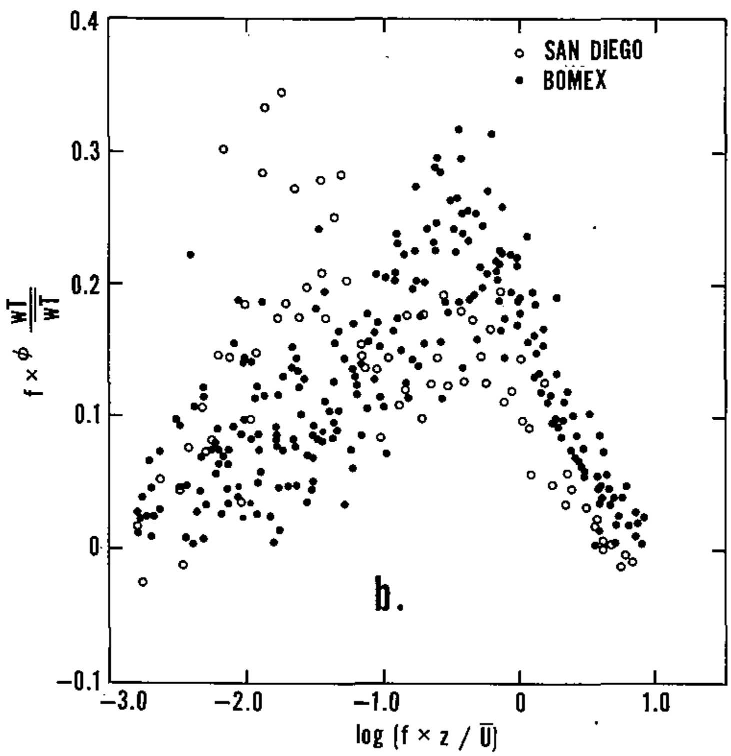

Figure 7. Normalized cospectrum of vertical velocity (w) and temperature (T) vs. log_{10} of non-dimensionalized frequency based on sonic anemometer and resistance thermometer at 8 m on FLIP (from Phelps and Pond, 1971).

the frequency band studied by Gibson and his coworkers. Figure 7 also shows that the sensible heat transfer at 8 m in the BOMEX environment was dominated by eddy scales in the tens of meters, near the peak in the w spectrum. There was no suggestion of thermal convection at larger scales.

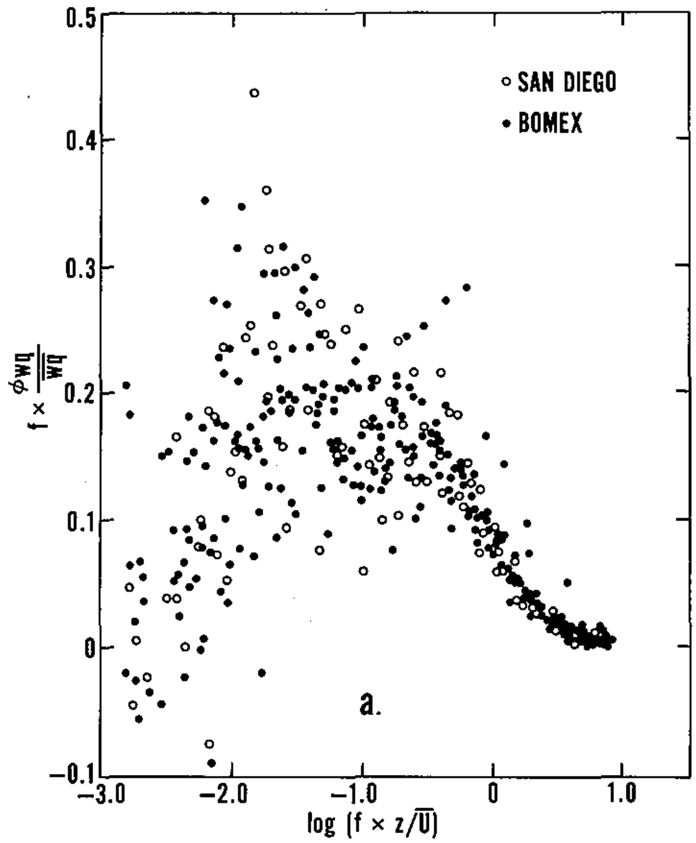

Figure 8. Normalized cospectrum of vertical velocity (w) and absolute humidity (q) vs. log_{10} of nondimensionalized frequency based on measurements on FLIP by sonic anemometer and Lyman alpha humidiometer (from Phelps and Pond, 1971).

Figure 8, also taken from Phelps and Pond (1971), shows the corresponding normalized cospectrum of w and q (although Phelps and Pond use q to represent absolute humidity, the normalization by $\overline{wq}$ makes this cospectrum essentially identical to that which would be obtained with specific humidity). Not suprisingly, w and q had maximum cospectral density at normalized frequencies about an order of magnitude lower than those of the wT cospectral peak. Despite much scatter of the data at low frequencies, it appears that the principal scales for vertical moisture transport at the 8-m level were from a few hundred meters to about a kilometer. Significant contributions to the water vapor flux were limited to normalized frequencies less than about 3, corresponding to wavelengths greater than about 3 m.

9. AIRCRAFT COSPECTRA

The NOAA RFF DC-6 aircraft, equipped with angle-of-attack vane and microwave refractometer (Friedman et al., 1970) obtained data on the vertical flux of water vapor at altitudes from 18 m to 2,400 m over the BOMEX square, with a frequency bandwidth from about 0.01 to 3 Hz. The data were low-pass filtered by both analog and digital methods with a cutoff at about 4 Hz to eliminate instrument boom vibration effects. At an air speed of about 93 m sec^{-1} this corresponds to a wavelength band from 30 m to 9 km. Figure 9, taken from Bean et al. (1971) of NOAA's Wave Propagation Laboratory, shows the averaged normalized cospectra of absolute humidity and vertical velocity for about four 10-min runs in each of four groups: at altitudes of 100 and 500 ft (30 and 150 m) and in the alongwind and crosswind direction at each altitude.

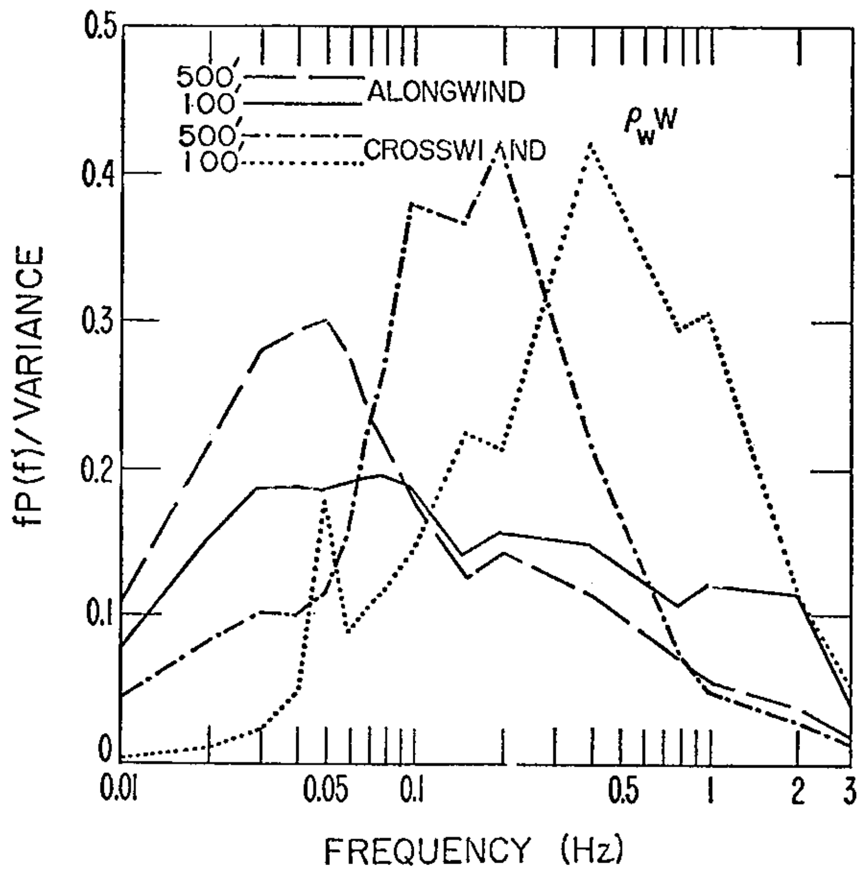

Figure 9. Normalized cospectrum of vertical velocity (w) and absolute humidity (ρ_w) vs. frequency based on measurements by NOAA DC-6 aircraft gust-probe/refractometer system (from Bean et al., 1971).

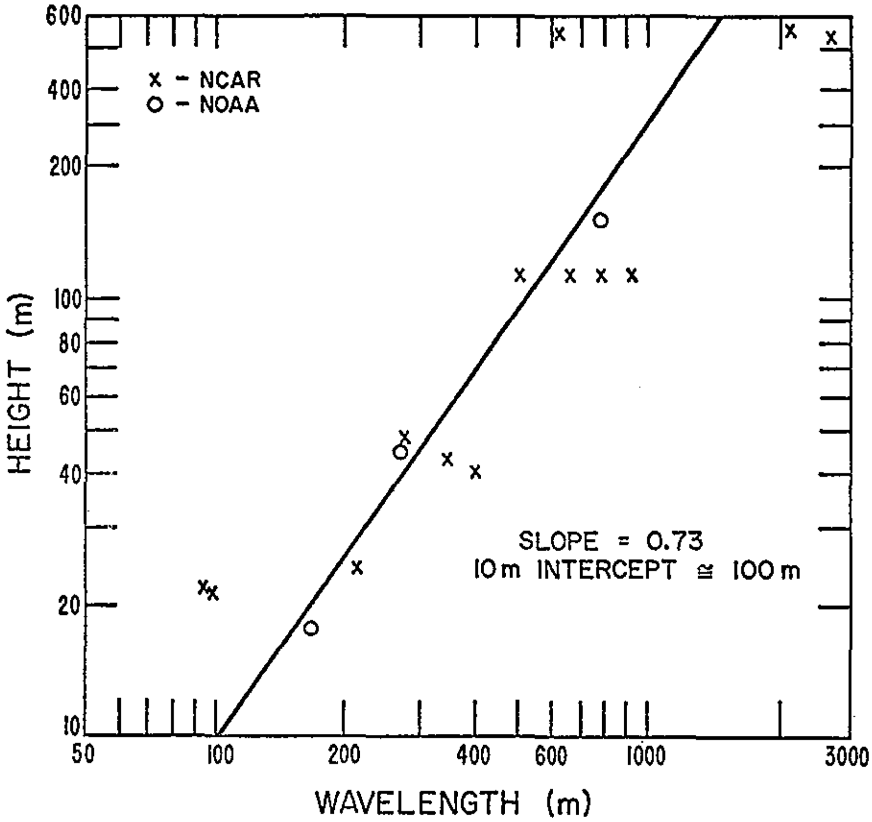

Figure 10. Wavelength of water vapor flux cospectral peak vs. height, based on measurements by UBC sonic anemometer and Lyman alpha humidiometer on NCAR Queen-Air aircraft (x) and by NOAA angle-of-attack vane and microwave refractometer on DC-6 aircraft (o) (from Bean et al., 1971).

The frequency of the peak of the cospectrum in the alongwind direction lay between .05 and .08 Hz, shifting only slightly with height. The corresponding wavelengths would be about 1.2 to 1.9 km, not more than a factor of two larger than those in figure 8 derived from the FLIP 8-m data.

The crosswind cospectra showed much sharper peaks, and the wavelength of the peak shifted with height, from about 250 m to 500 m as the height increased from 30 to 150 m.

The increase with height in the wavelength of the peak of the water vapor flux cospectra when measured in the crosswind direction was observed independently by the UBC group using the NCAR Queen-Air aircraft (Donelan, 1970). Figure 10, from Bean et al. (1971), shows the wavelengths of the crosswind cospectral peaks from both the NOAA and NCAR aircraft data in a log-log plot against height. The NOAA values are based on eleven to fourteen 5-min data samples at each altitude.

10. EDDY STRUCTURE AND REGULATION MECHANISM

The picture that emerges is one of highly elongated eddies carrying moist air up and dry air down, and having dimensions of the order of 1 to 3 km in the alongwind direction, or perhaps being even longer and oriented obliquely with respect to the wind. Their crosswind structure seems to consist of an eddy spectrum containing a range of crosswind scales, the scale that makes the principal contribution to the water vapor flux at each altitude increasing from about 100 to 200 m in the lowest 20 m to dimensions of the same order as the longitudinal wavelength as the cumulus base altitude is approached (about 500 m). These largest eddies may be in the nature of longitudinal roll vortices, with maximum energy in the crosswind component, although the evidence on this is unclear. These eddies, in turn, contain a broad spectrum of smaller eddies, apparently in the nature of transverse vortices or plumes with predominant energy in the longitudinal direction and feeding directly on the mean shear.

"Ramp" or "sawtooth" variations of temperature and humidity have been noted by several BOMEX participants. Gibson, Friehe, and McConnell (1971) have found slow-rise, rapid-drop sawtooth patterns in temperature time traces with a length scale of 10 m and less at a height of 2 m, which they have explained in terms of a train of transverse vortices. Phelps and Pond (1971) found similar patterns in humidity traces with length scales from tens to several hundred meters at a height of 8 m. Although the temperature and humidity traces had shown similar patterns in the San Diego pre-BOMEX trials, only the shortest of the humidity ramps in BOMEX were accompanied by a parallel variation of temperature. Examples from the BOMEX traces from these two groups are shown in figure 11.

In a rare verification of the validity of interpreting time traces in terms of spatial structure, Bean et al. (1971) have noted similar sawtooth patterns in the aircraft humidity data. Figure 12 shows vertical velocity and absolute humidity time traces taken in the upwind direction at 18 m (upper pair) and in the downwind direction at 46 m. Each trace represents about 2 km of travel. The steep dry fronts were always on the upwind side of the moist updrafts. There were four or five more or less distinct ramps of nearly equal amplitude in the 18-m humidity trace. In the 46-m trace, we see an example of the increase in wavelength by suppression or merging of some of the smaller features.

The close and ever increasing parallelism of the vertical velocity and humidity traces with increasing height despite the very small local vertical gradient of mean humidity at 46 m (cf. fig. 2), suggests that the buoyancy due to humidity is a significant driving force in generating these larger eddies. Donelan (1970), analyzing the relative contribution of sensible and virtual temperature to the production of turbulent kinetic energy, found that the positive contribution due to moisture flux not only exceeded that due to sensible heat flux in magnitude but was sufficient to overcome the negative effect of the downward heat flux on the larger scales.

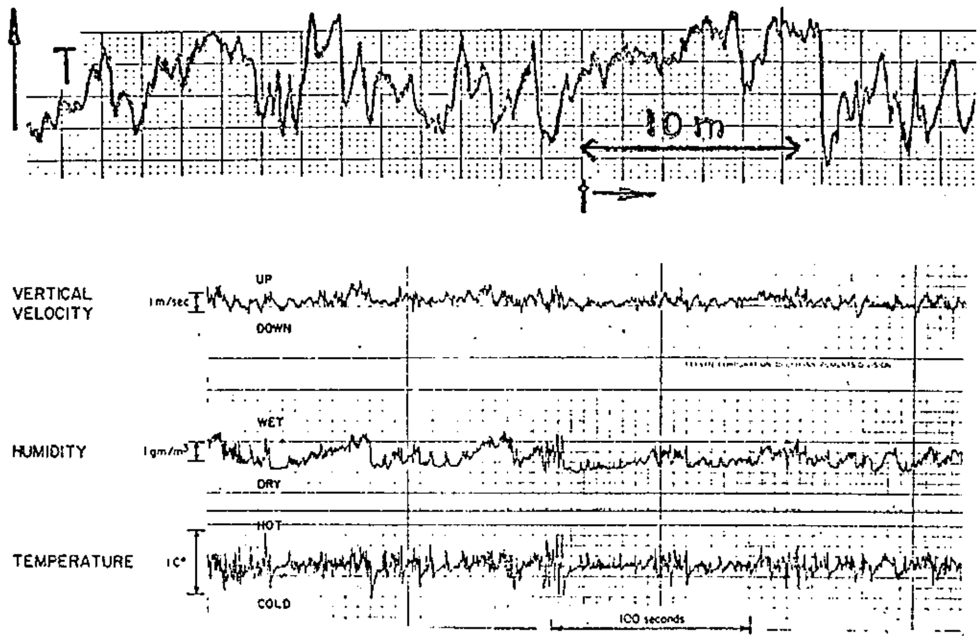

Figure 11. Sample time traces of temperature, vertical velocity, and humidity taken on FLIP. Top trace is from USCD fast resistance thermometer at 2-m level (from Gibson, Friehe, and McConnell, 1971). Lower three traces are from 8-m level (from Phelps and Pond, 1971).

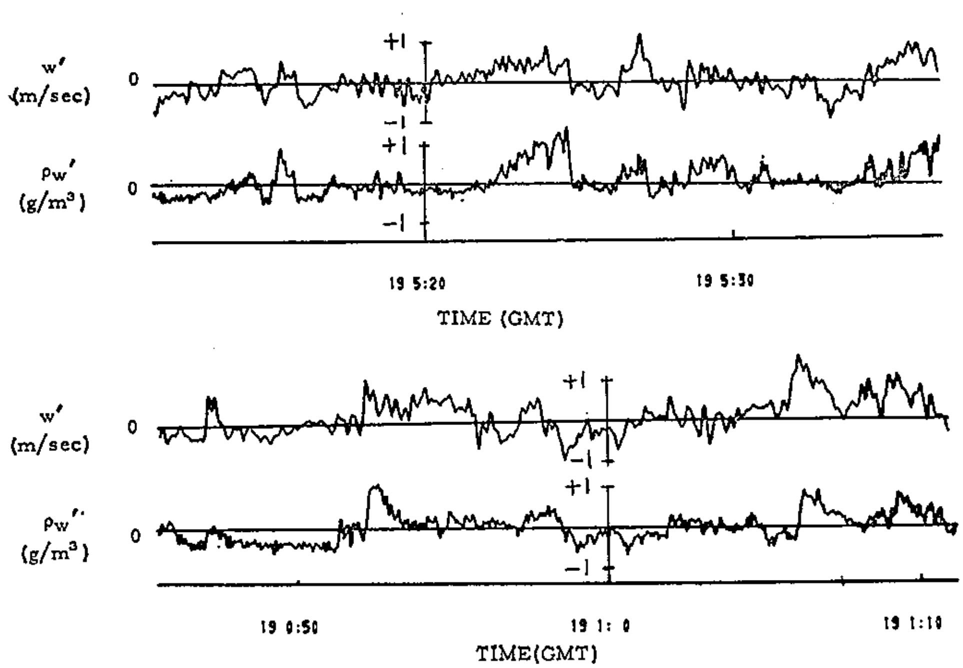

Figure 12. Time traces of vertical velocity (w') and absolute humidity (ρ_w') from NOAA gust-probe/refractometer system on DC-6 aircraft. Upper pair taken at 18 m heading upwind; lower pair at 46 m heading downwind (from Bean et al., 1971).

The data are consistent with the view that the mean temperature gradient, the turbulent temperature fluctuations, and the correlations of temperature with wind components and humidity are consequences of the balance between two opposing processes. One of these processes is the radiative one, which includes the heating of the upper ocean and cooling of the lower kilometer of the atmosphere and leads to an unstable temperature gradient over the sea surface. The other process is the evaporative cooling of the sea surface, which is regulated by the rate at which the boundary-layer-scale eddies supply dry air from above. If the large eddies should weaken, the lower atmosphere would become more unstable due, in the short term, primarily to infrared radiative energy loss from the air. If the large eddies should become too energetic, they would warm the air and cool the sea (the latter within narrower limits set by the destabilization and convective heat supply within the water) until the water vapor buoyancy is no longer able to overcome the thermal stability. On the boundary-layer scale, the downward sensible heat flux produced as a by-product of the sea-air evaporative energy transfer, and moist-buoyant convection, opposes the latter, and thereby operates as a negative feedback mechanism to regulate it.

11. FEASIBILITY OF BULK AERODYNAMIC METHOD

The UBC and OSU groups found that most of the time the stress computed from the uw covariance could be reasonably well approximated by $1.52 \times 10^{-3}U^2$, where U is the mean wind speed (fig. 13). They also found that the vertical flux of water vapor could be estimated quite well by $1.23 \times 10^{-3}U\Delta q$, where Δq is the difference between the saturation absolute humidity at the sea surface temperature and the observed absolute humidity at the height of measurement (fig. 14). They found, on the other hand, that $U\Delta T$ (where ΔT is the difference in temperature between the sea surface and the height of measurement) did not serve as a useful predictor for the sensible heat flux (fig. 15). I have plotted in fig. 16 the values of $\overline{w'T'}$ from Pond et al. (1971) in terms of cal cm^{-2} day^{-1} against $U\Delta q$ (gm cm^{-2} sec^{-1}) computed from the values of U and Δq given in their tables.

It appears that $U\Delta q$ is a better predictor of the sensible heat flux than is $U\Delta T$! This seems to confirm that, in the relatively undisturbed, delicately balanced tradewind region in which these observations were taken, the vertical temperature structure responds passively to the energy transfer process, dominated by evaporation even in the surface layer where the sensible and latent heat fluxes have the same sign. What we observe is that a positive deviation of the heat flux may be accompanied by either a positive or negative deviation of the temperature gradient, but, in either case, will generally be accompanied by a positive departure of the evaporation. Indeed, the negative feedback hypothesis requires that the longer-period fluctuations of heat flux lag behind the temperature gradient fluctuations on the order of a quarter cycle.

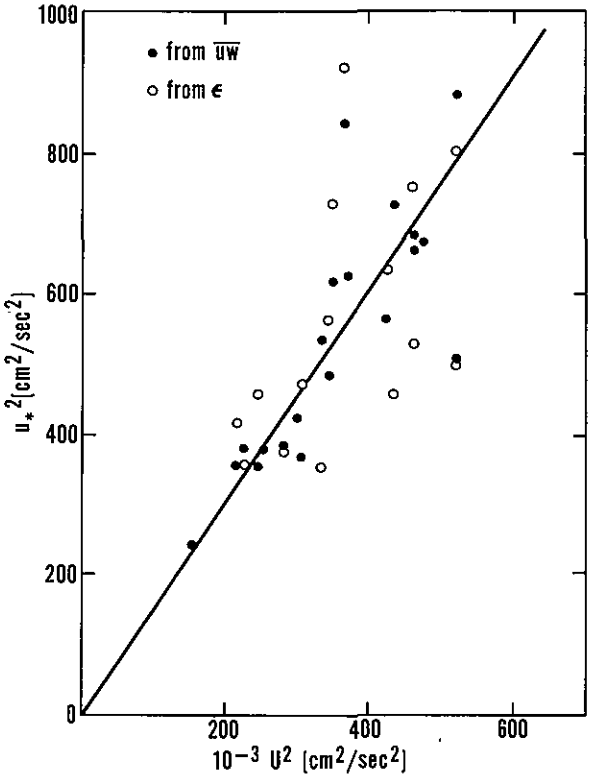

Figure 13. *Stress (u_*^2) vs. square of mean wind speed (U^2) at 8 m on FLIP (from Pond et al., 1971). The straight line corresponds to $C_D = 1.52 \times 10^{-3}$ in the formula $u_*^2 = C_D U^2$.*

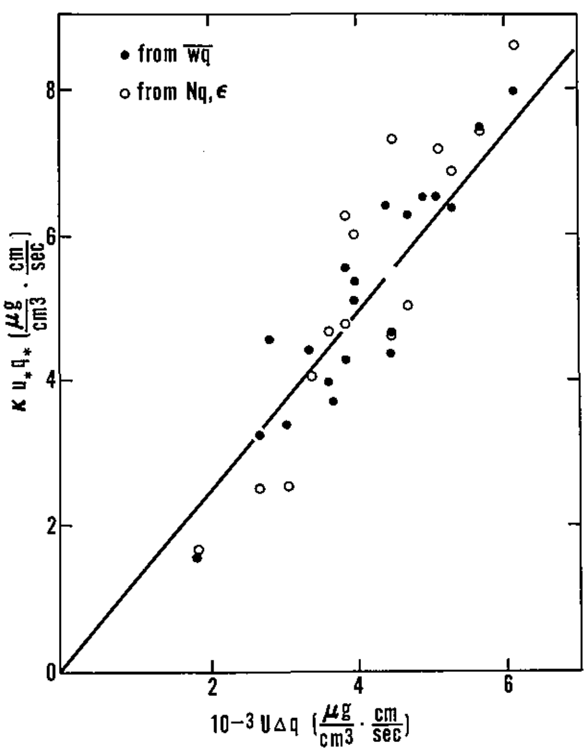

Figure 14. *Vertical flux of water vapor ($\kappa u_* q_*$) vs. product of mean wind speed (U) and absolute humidity deficit at 8 m relative to saturation at sea surface (Δq) on FLIP (from Pond et al., 1971). The straight line corresponds to $C_q = 1.23 \times 10^{-3}$ in the formula $\kappa U_* q_* = C_q U \Delta q$.*

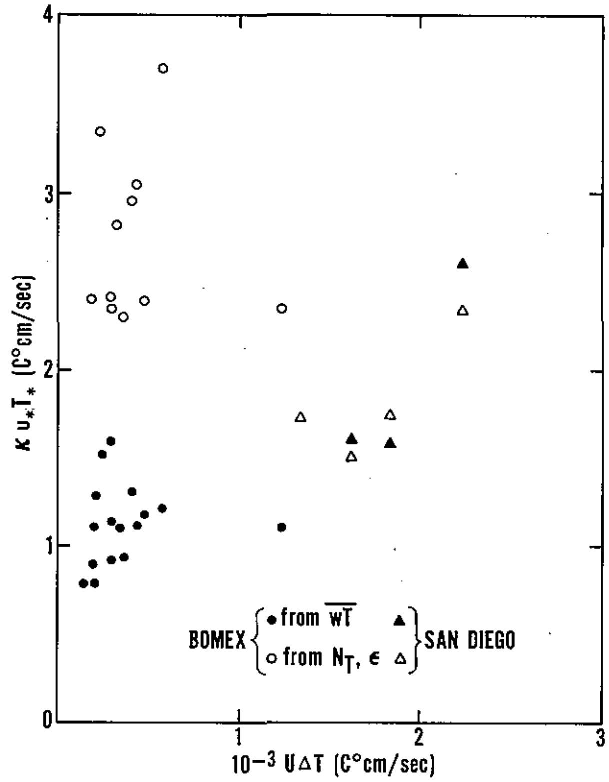

Figure 15. Vertical heat flux ($\kappa u_* T_*$) vs. product of mean wind speed (U) and temperature deficit at 8 m relative to sea surface (ΔT) on FLIP (from Pond et al., 1971).

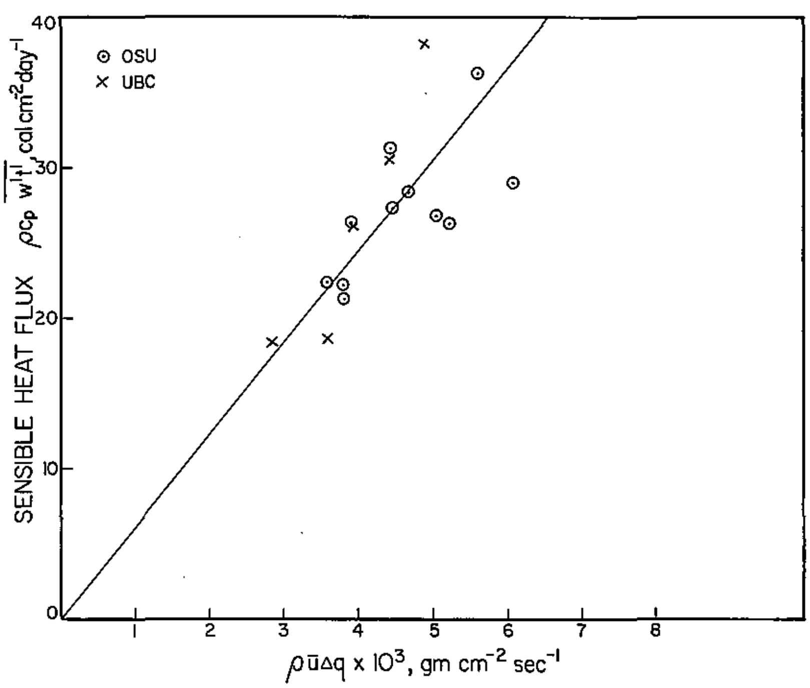

Figure 16. Vertical heat flux vs. product of mean wind speed ($\bar{u}$) and absolute humidity deficit at 8 m relative to saturation at sea surface (Δq) on FLIP (based on data from Pond et al., 1971).

Thus while the bulk aerodynamic coefficient for heat flux was not found to be constant, Pond et al. (1971) found the Bowen ratio to be fairly constant, varying only between 0.07 and 0.14, with an average of 0.10 in 16 BOMEX data samples.

12. TIME, SPACE, AND ALTITUDE VARIATIONS OF WATER VAPOR FLUX, AND COMPARISONS OF DIFFERENT METHODS OF MEASUREMENT

It is desirable to compare the water vapor flux measurements by the different methods. An intercomparison of the UBC Queen-Air and RFF DC-6 systems was carried out over station DELTA, at the southwest corner of the BOMEX square, on May 29, 1969, between 1715 and 1830 local time. Measurements of water vapor flux by the Queen-Air, at altitudes of 26, 49, 88, and 140 m, reported by Donelan (1970), when converted to surface evaporation rates, give 4.96, 4.27, 2.77, and 3.00 mm day^{-1}, respectively. DC-6 aircraft measurements at 18, 45, 90, and 135 m give 4.45, 4.14, 5.45, and 3.69 mm day^{-1} (average of two samples at each of the upper three levels). These and other nearly simultaneous data collected by the two aircraft show that any relative bias must be small compared to the scatter of the data, particularly at the lower altitudes.*

Unfortunately the samples of FLIP data analyzed by the OSU and UBC groups so far do not correspond to the times of aircraft runs, but there are several cases where data from these two measurement systems are separated by only a few hours. In figure 17 a variety of measured evaporation rates are shown as a function of time for the periods May 5-7 and May 10-12, 1969. These include values based on eddy flux data analyzed by the OSU and UBC groups (Pond et al., 1971), values based on mean profile measurements by the UW group on FLIP (Paulson et al., 1970), and values obtained by bulk aerodynamic computations for the times of FLIP eddy flux measurements, using U, Δq, and the empirical coefficients given by Pond et al. (1971).

Also shown in figure 17 are average values of the vertical flux of water vapor for groups of six NOAA aircraft runs at FLIP. The usual flight pattern consisted of six 5-min straight and level data runs, one alongwind and one acrosswind at each of three altitudes: 18 m, 45 m, and 150 m. Bean et al. (1971) have shown that the covariance of vertical velocity and absolute humidity has very little (of the order of 5 percent) systematic variation with height,

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*Comments regarding Queen-Air data on p. 38, *BOMEX Bulletin* No. 10, the BOMAP Office, 1971, are erroneous, due to erroneous plotting of Queen-Air data in fig. 6 of that article.

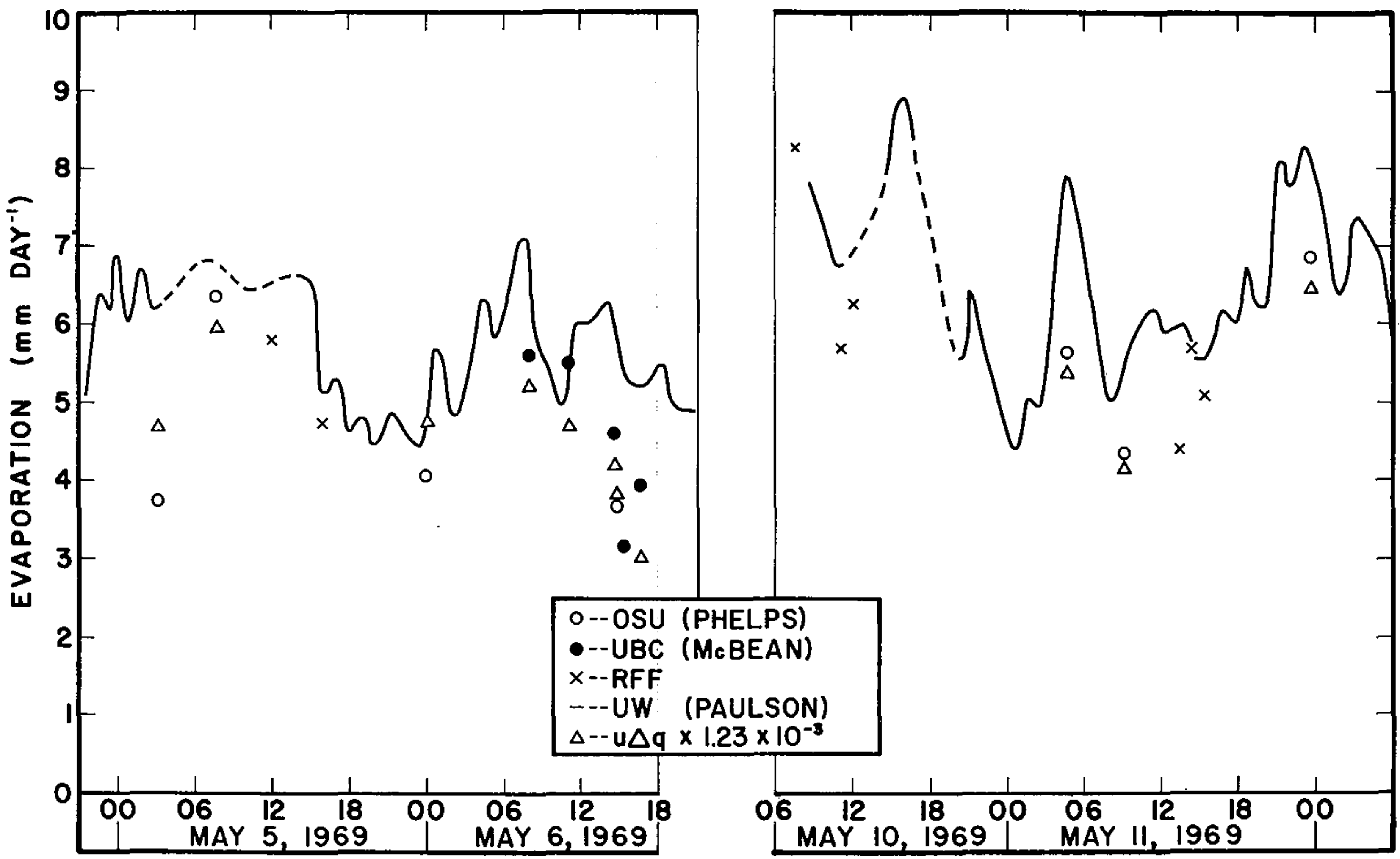

Figure 17. Time graph of evaporation rate measurements by Phelps and Pond of OSU (o) and McBean of UBC (Pond et al., 1971) (●) on FLIP using eddy covariance method, by Paulson, Leavitt, and Fleagle on FLIP using UW profile method (solid curve), by NOAA Research Flight Facility (RFF) DC-6 aircraft (X), and by bulk aerodynamic method applied to FLIP data of Pond et al. (Δ). May 4, 2123, to May 6, 2145, and May 10, 1045, to May 12, 0653, 1969 (local time).

while the variability from one sample to another at the same height, even on the same day, is considerable (of the order of 50 percent). Accordingly, a better estimate of the surface evaporation for a given time is obtained by averaging together the vertical fluxes computed from all six samples in a group than by using only the 18-m crosswind sample, which is probably the most accurate single sample.

Three features of this time-series plot are of interest. First, there are large variations over a factor of two in evaporation rate (4 to 8 m day^{-1}), shown by all the measurement systems over periods of the order of 20 to 40 hours, but not in a consistent diurnal phase. If this is typical, a very long sample would be required to isolate the diurnal variation of evaporation.

Second, the FLIP covariance, profile and bulk aerodynamic estimates, as well as the NOAA aircraft fluxes are very consistent with respect to these large-scale trends and show no evidence of relative bias, except that the profile estimates from Paulson et al. (1970) run about 25 percent higher than the other measurements. This does not necessarily indicate that the profile data are wrong, since the covariances may suffer from negative bias due to bandwidth limitations. In fact, this bulk aerodynamic coefficient (1.23 x 10^{-3}) based on the observed covariances is slightly lower than that derived from the widely used Deacon and Webb (1962) formula, which would give 1.42 x 10^{-3} for a 6 m sec^{-1} wind at 10 m and a still larger value for 8 m. Still, it is difficult to account for a negative bias, due to bandwidth truncation, greater than about 5 percent in either of the covariance systems. So perhaps the truth lies in between the covariance and the profile figures.

Third, the profile estimates, which have the best time resolution and continuity, show a persistent fluctuation with a variable period between 2 and 4 hours (averaging about 2.7 hours) and with a variable amplitude of the order of 1/2 mm day^{-1}. There is nothing in the descriptions of the design or operation of the UW profile measurement system, nor in the method of analysis of the data, that would suggest an artificial source of such an oscillation. Furthermore, whenever a series of several samples taken over a short span of hours is available from either FLIP or the aircraft (samples of the order of 1/2-hour durations or more on FLIP; groups of six consecutive 5-min runs on the aircraft), they show sample-to-sample variations of this same order. One is tempted to speculate that one of the processes in the feedback loop has a time constant such as to cause the balance between evaporative-convective stabilization and radiative destabilization to oscillate above and below the equilibrium point with just this period. An alternative inference would be advection of motion systems having dimensions of the order of 100 km in the alongwind direction.

Although FLIP was not operating in the BOMEX array during the period covered by the atmospheric and oceanic energy budget analyses summarized earlier, there is a link by which their mutual consistency can be checked. The NOAA gust-probe aircraft, whose measured fluxes have been seen to be quite consistent with those of the OSU-UBC team aboard FLIP, carried out a thorough water vapor flux survey of the BOMEX square on June 29, 1969. This was during the second (disturbed) interval for which the ocean heat budget was evaluated. Figure 18 shows, plotted on a map of the BOMEX array, the evaporation rates deduced from the total energy fluxes given in table 2 for the five ships during June 28 - July 2, using a Bowen ratio of 0.1. For this estimate, the average solar radiation observed at the three ships having pyranometers was used for the other two ships. Also shown are the evaporation rates based on the nine 10-min water vapor flux runs on June 29 and two runs on June 30. The broad spatial patterns of evaporation rates obtained by the two methods, showing a maximum in the southwest-central portion of the square, are in remarkably good agreement. The average values for the BOMEX area from the ocean heat budget for June 28 - July 2, based on the three ships having pyranometers, and the average value from the aircraft sampling for June 29 are both 6.1 mm day^{-1}.

Another consistency check can be made by comparing the vertical variation of vertical flux of water vapor, as compiled from the various aircraft runs, which were taken above the first few hundred feet, with that determined from the 10-mb slicewise integration of the residual (subgrid scale vertical flux divergence) of the atmospheric water vapor budget equation. These are shown in figure 19. The surface flux for the trial integration period lies in the middle of the scattered low-level aircraft data, but with increasing altitude the aircraft data decrease much more rapidly than the budget data. The aircraft fluxes approach zero in the cumulus cloud layer, while the budget data remain high through this layer, dropping slowly to zero above the tradewind inversion.

This is a plausible relationship when it is remembered that the aircraft fluxes do not contain contributions due to eddies larger than 9 km. The budget includes all scales up to 500 km. Figure 19 suggests that in the cumulus and tradewind inversion layers there is a considerable upward flux of moisture accomplished by the mesoscale eddies.

Average values of evaporation rate in mm day^{-1} for all measurements available by each method within each of the time periods May 5-12, June 20-26, and June 28 - July 2, 1969, are summarized in table 3. The difference between the aircraft and FLIP values appears to be due to sampling variations, since the FLIP samples selected for analysis favor times of relatively low flux.

13. SUMMARY AND CONCLUSIONS

Preliminary values for the sea-air flux of heat, water vapor, and momentum during BOMEX have been obtained by a number of different measurement systems and analytical schemes. The data agree generally in putting the evaporation rate during undisturbed tradewind conditions between 5 and 6 mm day^{-1}, and the Bowen ratio between .09 and .10. The evaporation, driven by the wind speed and dry air supply, controls the sensible heat flux, which is upward at the surface but downward in the boundary layer as a whole.

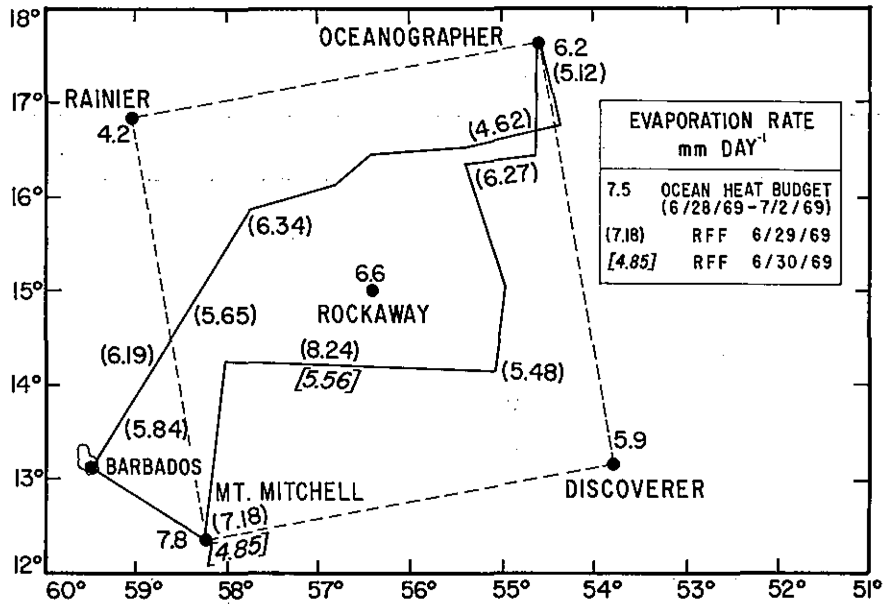

Figure 18. Map of BOMEX area showing evaporation rates (mm day^{-1}) from ocean heat budget for June 28 - July 2, 1969 (numbers without brackets); and from NOAA DC-6 aircraft measurements on June 29, (()) and June 20, 1969 ([]).

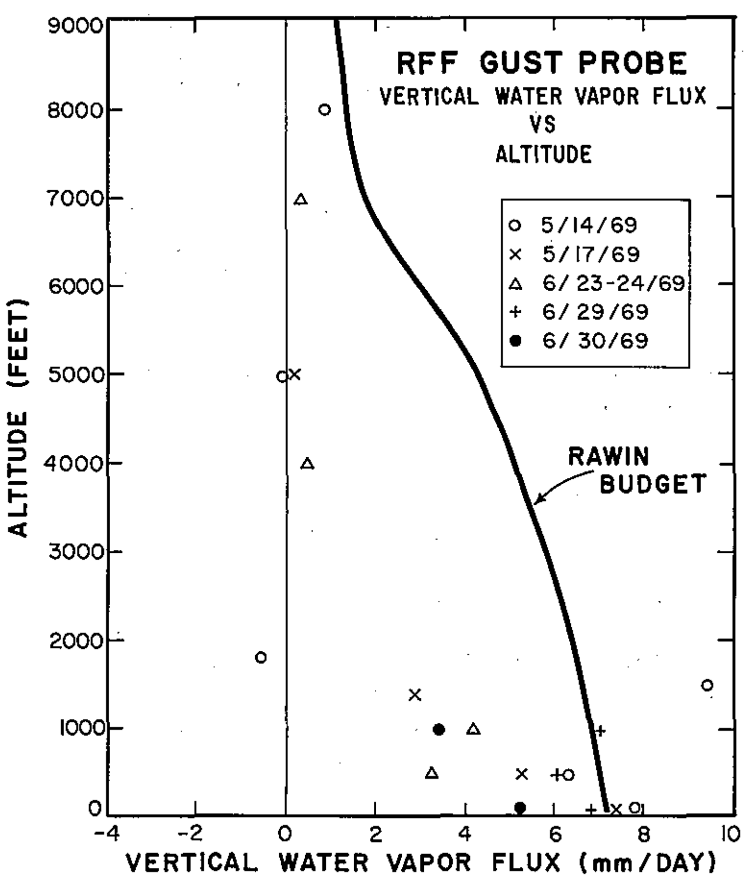

Figure 19. Vertical water vapor fluxes at altitudes above 150 m (500 ft) expressed in terms of evaporation rate (mm day^{-1}) from NOAA aircraft measurements on various dates and from the atmospheric water vapor budget analysis for the BOMEX 500-km square for the 5-day period June 22-26, 1969, based on rawinsonde data only.

Table 3. BOMEX average evaporation rates, mm day^{-1}

	May 5-12	June 20-26	June 28 - July 2
Atmospheric budget (rawinsondes)		6.0	
Ocean heat budget		5.8	6.1
Covariance: NOAA aircraft	5.8		6.1
Covariance: FLIP (OSU;UBC)	4.9		
Profile: FLIP (UW)	6.0		
Bulk aerodynamic (Deacon-Webb)	6.3	5.6	

The overwhelming importance of water vapor results from its three roles. First, because of its large latent heat, it is the principal cooling agent at the ocean surface, transfering energy in a nearly isothermal situation. Second, on the microscale in the atmosphere it contributes buoyancy that can promote convection when the temperature is very nearly uniform and ordinary thermal convection weak. Third, on the larger scales, because of its modest saturation density, it is readily condensed in the troposphere, reconditioning the air to receive a fresh supply of vapor and energy. In the latter process it liberates heat, which accumulates in the troposphere and has a stabilizing effect on the lower atmosphere. Each of these roles is regulated by a negative feedback mechanism.

The surface evaporation cools and destabilizes the upper mixed layer of the sea, bringing warm water to the surface and thus maintaining a nearly constant surface water temperature, even with widely varying evaporation rates.

The moist-buoyant convection on the boundary layer scale transfers sensible heat downward, a process that consumes kinetic energy. Its intensity is thus held within narrow limits.

The large-scale convective condensation and recycling of dry air is self-limiting by building up the stability of the troposphere.

Although only a few direct comparisons of analyzed data from the different measurement methods have been possible so far, it seems safe to say that present-day aircraft measurement techniques for turbulent flux are adequate to permit an atmospheric budget experiment to be designed with sea-air fluxes as measured inputs, and with the more difficult flux divergence and conversion terms for the upper layers as outputs.

14. ACKNOWLEDGMENTS

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