ESTIMATING MONTHLY-AVERAGED AIR-SEA TRANSFERS OF HEAT AND MOMENTUM USING THE BULK AERODYNAMIC METHOD

Steven K. Esbensen

and

Richard W. Reynolds[1]

(NASA-CR-163045) ESTIMATING MONTHLY-AVERAGED AIR-SEA TRANSFERS OF HEAT AND MOMENTUM USING THE BULK AERODYNAMIC METHOD (Oregon State Univ.) 35 p HC A03/MF A01 CSCL 04B G3/47

N80-28990

Unclas

26986

July 1980

Climatic Research Institute
Oregon State University
Corvallis, Oregon 97331

[1]Present affiliation: Climatic Analysis Center, NMC, NWS, NOAA, Washington, D.C., 20233

ABSTRACT

Air-sea transfers of sensible heat, latent heat, and momentum are computed from twenty-five years of middle-latitude and subtropical ocean weather ship data in the North Atlantic and North Pacific using the bulk aerodynamic method. The results show that monthly-averaged wind speeds, temperatures and humidities can be used to estimate the monthly-averaged sensible and latent heat fluxes computed from the bulk aerodynamic equations to within a relative error of approximately 10%. The estimates of monthly-averaged wind stress under the assumption of neutral stability are shown to be within approximately 5% of the monthly-averaged non-neutral values.

1. Introduction

Estimates of the transfers of sensible heat, latent heat, and momentum between the atmosphere and the oceans may provide important insight into the dynamics of interannual and decadal climate fluctuations. Sensible and latent fluxes combine with radiative fluxes to determine the major part of the net air-sea exchange. Momentum transfers play an important role in driving the ocean general circulation and particularly in determining the vertical structure of the upper ocean.

The first systematic evaluation of the sensible and latent heat transfers over the world ocean was performed by Budyko (1963). He computed long-term monthly-mean values of the near-surface wind speed and of the air-sea temperature and humidity differences and evaluated the fluxes by the well-known bulk aerodynamic formulas with constant transfer coefficients (see section 2).

Budyko's method depends on the validity of the following two conditions:

1. The covariance of the wind speed with either the air-sea temperature difference or the air-sea humidity difference must be less than the product of their respective mean values (except for cases in which the fluxes are small). This condition was found to be satisfied in several important climatic regions of the world oceans for limited time periods by Malkus (1962), Budyko and Gandin (1966), Kraus and Morrison (1966) and Fissel *et al.* (1977).
2. The transfer coefficients for sensible heat and evaporation must not be strongly dependent on either wind speed or air-sea temperature and humidity differences for typical conditions over the oceans. Budyko (1974) concludes that the stability of the surface boundary layer, as parameterized by the bulk Richardson number, is almost neutral except in weak wind conditions. As we will show in subsequent sections, recent developments in surface boundary layer theory support Budyko's conclusion.

The first systematic evaluation of air-sea momentum transfers on a global basis was presented by Hellerman (1967). His computations were based on the bulk aerodynamic formulas with a drag coefficient for neutral surface boundary layer conditions and wind data consisting of monthly-averaged wind roses with both wind direction and speed categories. The use of only one mean wind speed in each direction category of a monthly-averaged wind rose leads (see Hellerman, 1965) to underestimates of the computed monthly-averaged wind stress of 10 to 30 percent.

Recently there has been considerable interest in updating the classical estimates of Budyko (1963) and Hellerman (1967). An excellent example of this type of work is Bunker's (1976) estimates of the surface heat balance over the Atlantic Ocean. The approach used by Bunker differs from the classical approach since he computed the turbulent fluxes for each simultaneous measurement of surface meteorological variables. The computed fluxes within a given month are then summed, and finally are divided by the total number of observations. Data processing technology has made Bunker's approach feasible. However, more than thirty million ship reports would have to be processed at the present time to obtain global estimates of the turbulent heat and momentum transports. The classical method clearly has a number of practical advantages which will be discussed further in our final section.

The main purpose of this paper is to provide a basis for determining to what extent the classical method of estimating turbulent fluxes of sensible and latent heat from monthly averaged surface meteorological variables can be used for studies of interannual and decadal climatic variability. A unique feature of our study is the systematic exploration of the effects of boundary layer stability and the physics of the air-sea interface on the estimates of the sensible and latent heat fluxes using recent developments in boundary layer theory. We will also discuss the importance of stability effects for the estimates of air-sea momentum transfers. Computations will be based on data taken from ocean weather ships in the North Pacific and North Atlantic Oceans since these data are the best available long-term records of surface marine meteorological variables. In the final section, we will discuss the implications of our results for studies of climatic variability.

2. The Bulk Aerodynamic Calculations

The well-known bulk aerodynamic formulas for the sensible heat flux S and the latent heat flux LE at the air-sea interface are

$$S = \rho c_p C_H V_a (T_s - T_a) \tag{1}$$

and

$$LE = \rho L C_E V_a (q_s - q_a), \tag{2}$$

where the variables V_a, T_a and q_a are the wind speed, temperature, and specific humidity of the air at a given height (z_a) which is within the surface boundary layer. The variables T_s and q_s are the sea-surface temperature and the saturation specific humidity of air with temperature T_s and sea-level pressure p_s.[1] In this study we have used $z_a = 10$m and approximated the potential temperature difference $(\theta_s - \theta_a)$ by $(T_s - T_a)$. The variable ρ is the average air density in the layer from the surface to z_a; c_p is the specific heat of air at constant pressure; and L is the latent heat of evaporation of water. The quantities C_H and C_E are the so-called transfer coefficients for sensible heat and evaporation, respectively.

To estimate the fluxes from Eqs. (1) and (2), the transfer coefficients must be determined. Although these coefficients may be determined in a completely empirical manner using regression analysis, boundary layer similarity theory provides a rational, physically motivated way to minimize the dependence of C_H and C_E on empiricism. In particular, similarity theory allows the inclusion of the effects of the stability of the atmospheric surface boundary layer; the measure of stability being the Monin-Obukov length which is defined as

$$L = -\frac{(\tau/\rho)^{3/2}}{\frac{g}{\bar{T}_V}\,\overline{w'T'_V}} \tag{3}$$

where τ is the magnitude of the surface wind stress (see section 7), T_V is the virtual temperature, and g is the gravitational acceleration. Here the overbar indicates the Reynolds averaging of the turbulent product $w'T'_V$ where w is the vertical air velocity. Similarity theory also allows the effects of the roughness of the sea-surface to be parameterized by a roughness length z_o.

We have chosen to evaluate the transfer coefficients using the method proposed by Liu *et al.* (1979) with the exception that we have

---

[1]The variables V_a, T_s, T_a, q_s and q_a are all assumed to be averages over a time-scale at least as long as the lifetime of boundary layer convective elements.

neglected the difference between the "observed" sea-surface temperature and the skin temperature of the water at the air-sea interface. The effects of boundary layer stability are included using the Businger-Dyer model (see Businger *et al.*, 1971 and Dyer and Hicks, 1970). Also the Liu *et al.* model determines the roughness length z_o and the corresponding lengths for the profiles of temperature and specific humidity, z_T and z_q; the lengths are not assumed to be equal. There is no free convection limit for unstable conditions as considered by Deardorff (1972); however, calm conditions with large unstable air-sea temperature and humidity differences are rare over the oceans.

The dependence of C_H and C_E on wind speed and air-sea temperature differences computed by the Liu *et al.* method are shown in Fig. 1a and 1b for a relative humidity of 70%. A comparison of these results show that for any given wind speed and temperature difference, C_H and C_E differ by less than 10%. However, at wind speeds less than 10 m/s both C_H and C_E are strongly dependent on both wind speed and air-sea temperature differences. At wind speeds greater than 10 m/s, the stability of the boundary layer has very little effect on the transfer coefficients and the dependence on wind speed is weak.

The transfer coefficients proposed by Liu *et al.* (1979) agree well with those of Kondo (1975), and give a reasonable fit of the currently available data from the surface boundary layers of the ocean and atmosphere. However, as we will show in sections 4 and 5, the conclusions of this study do not depend crucially on the choice of transfer coefficients. This may seem paradoxical to the reader given the importance of the transfer coefficients in obtaining reliable estimates of the fluxes. However, this study is concerned with the *methodology* of estimating monthly-averaged and long-term climatological fluxes from the bulk aerodynamic equations. We believe our conclusions concerning the choice of method for obtaining flux estimates are valid for any physically reasonable choice of transfer coefficients.

3. Two Estimation Methods for Monthly-averaged Heat Fluxes

In this section we distinguish between the *sampling method* and the *classical method* for computing monthly-averaged transfers of sensible and latent heat at the sea surface from the bulk aerodynamic equations. In the sampling method, fluxes are estimated from the sample means of S and LE in Eqs. (1) and (2) for a given calendar month and geographic area. These sample means are defined as

$$\overline{S} = \rho c_p \overline{C_H V_a (T_s - T_a)}, \tag{4}$$

and

$$\overline{LE} = \rho L \, \overline{C_E V_a (q_s - q_a)}, \tag{5}$$

where the averaging operation defined by the overbar is

$$\overline{(\;)} = \frac{1}{N} \sum_{n=1}^{N} (\;)_n, \tag{6}$$

and N is the number of simultaneous measurements of T_a, q_a, T_s, and V_a. It should be noted that ρ, c_p, and L are not included in the averaging operation since these variations either play a negligible role in the surface flux estimates or can be incorporated using linear combinations of $\overline{S}$ and $\overline{LE}$ (e.g., Brook, 1978).

The classical method (see Budyko, 1963) uses the following expressions:

$$\hat{S} = \rho c_p \hat{C}_H \bar{V}_a \; (\bar{T}_s - \bar{T}_a), \tag{7}$$

and

$$\hat{LE} = \rho L \; \hat{C}_E \bar{V}_a \; (\bar{q}_s - \bar{q}_a), \tag{8}$$

where $\hat{C}_H$ and $\hat{C}_E$ are obtained by substituting the sample means of T_a, T_s, q_a, q_s, and V_a into the algorithm for C_H and C_E.

If measurements of surface meteorological variables are always simultaneous and dense in space and time, and if the resources for data processing are adequate, the sampling method provides the best estimate of the monthly-averaged flux. However, for non-simultaneous and sparse data sets, the classical method may have both practical and theoretical advantages as discussed in section 8.

4. Data

To quantitatively test the differences between the two methods, we have analyzed three-hourly marine meteorological observations from nine ocean weather ships. The study was limited to the 25 year period from 1948-1972 and the ships were selected to give good geographical coverage. Table 1 shows the ship positions and periods for which observations were available. The position of each ship is also shown in Fig. 2 with the ocean surface current regions from Sverdrup *et al.* (1942). As can be seen, the ships have sampled both eastern and western middle-latitude oceanic regions including the strong western boundary currents of the Kuroshio and Gulf Stream. Six of the ships have nearly continuous records for the entire 25-year period.

We processed the data to eliminate any observations that were obviously in error or in which the weather ship was not within 1 degree of the position given in Table 1. We also eliminated data when simultaneous observations of wind speed, wind direction, sea level pressure, and air, sea and dew point temperature were not available.

Table 1. Ocean weather ship stations studied.

Station	Latitude	Longitude	Period Studied
A	62°N	33°W	1948-1971
B	56½°N	51°W	1948-1972
D	44°N	41°W	1950-1972
H	36°N	70°W	1950-1954
J	52½°N	20°W	1951-1971
N	30°N	140°W	1948-1972
P	50°N	145°W	1950-1972
V	34°N	164°E	1956-1971
X	39°N	153°E	1948-1953

5. Long-term Climatological Estimates

In this section we see how closely the classical method approximates the sample mean of flux estimates from temporally dense observations taken within a given calendar month for a period of ten to twenty years. The principal result is shown in Fig. 3 for January and July. The classical method agrees with the method of sample means to within a few percent in both summer and winter and in all climatological regimes of the subtropical and middle-latitude oceans which we have investigated. The only exceptions to this conclusion occur when the fluxes are small. However, when the fluxes are small, percentage comparisons are meaningless and Fig. 3 shows the agreement to be within a few W m^{-2} in these cases.

The authors believe the above conclusions are not likely to be changed by any physically reasonable choice of transfer coefficients. To demonstrate this point, the fluxes shown in Fig. 3 were recomputed with the assumption that sensible heat, water vapor, and momentum are exchanged between the ocean and the atmosphere in the same manner. (Thus, in the Liu *et al.* (1979) algorithm $z_T = z_q = z_o$.) As shown in Fig. 4, the transfer coefficient for sensible heat using this unrealistic assumption has an extreme dependence on wind speed. The curves for C_E are similar.

The heat fluxes for January with this assumption are shown in Fig. 5. Again, the discrepancies between the classical method and the sample mean are only a few percent for the latent heat fluxes and about ten percent or less for the sensible heat fluxes. This result implies that the difference between Budyko's (1963) flux estimates and Bunker's (1976) estimates for the Atlantic Ocean are due primarily to data availability, data reliability and the use of different transfer coefficients. The differences are *not* due to the fact that Budyko chose the classical method and Bunker chose the sampling method.

Fig. 6 shows the comparison of the classical and sampling methods using only nighttime observations for January. The conclusions based on Fig. 3 are unchanged. Diurnal effects, whether they are real or instrumental, do not lead to significant differences between the two methods.

We now consider the importance of including boundary layer stability effects on long-term climatological estimates of the sensible and latent heat fluxes. The heat fluxes with the neutral transfer coefficients from Fig. 2 were recomputed using the sampling method. Fig. 7 shows a comparison of these results with the stability-corrected sampling method results. The figure shows differences of approximately ten percent. Budyko (1974) reached a similar conclusion by using a bulk Richardson number as the stability parameter for the boundary layer.

6. Monthly Averaged Estimates

The results from the previous section suggest that the classical method may also be successful in monitoring the interannual variability of the sensible and latent heat fluxes. To examine this possibility we calculated the heat fluxes for each January and July at all of the ocean weather ships using both the classical method and the sampling method. Since the results of the comparison are almost independent of the choice of month or ship, we have chosen to present only the January results at ships D and N representing significantly different heat flux regimes.

Fig. 8a shows the comparison for Ship D. Although the average heat fluxes are large (see Fig. 3), the interannual variability of the monthly-averaged fluxes is also large. Despite the observed variability, the classical method still gives estimates of the heat fluxes which are within ten percent of the values obtained by the sampling method. These results are consistent with the conclusions of Fissel *et al.* (1977) for Ship P. They performed cross-spectral analysis on a series of two-year segments of data at Ship P with constant transfer coefficients and found that disturbances with periods of less than a month contributed only ten percent of the biennially-averaged sensible heat flux and only five percent of the biennially-averaged latent heat flux.

The results in Fig. 8b for Ship N, show that the interannual variability of the heat fluxes in subtropical regions is also well resolved by the classical method. The classical estimate of the latent heat flux is systematically higher than the sampling estimate, but still within ten percent.

The comparisons in this section were based on the transfer coefficients shown in Fig. 1. However, we also computed the monthly-averaged fluxes with the unrealistic transfer coefficients shown in Fig. 4. Since these coefficients have a stronger dependence on wind speed, the sampling and classical methods are not in such close agreement. However, the average disagreement for the 25 year period was still less than 10% and never exceeded 20% in any individual month.

Finally, we wish to point out that the transfer coefficients themselves are subject to uncertainties of approximately 20 to 30% (e.g., Pond *et al.*, 1974). In view of this uncertainty, the classical method is a practical and reliable alternative to the method of sample means. The uncertainty underscores the need for independent methods for obtaining the heat fluxes such as the budget approach of Oort and Vonder Haar (1976).

7. Climatological and Monthly-Averaged Wind Stress Estimates

Although the classical method is successful for the estimation of sensible and latent heat fluxes, it cannot be used to compute the wind stress. As discussed in Hellerman (1965) a wind rose with both direction and speed categories is necessary to accurately determine the wind stress. In this section we re-examine Hellerman's result using the weather ship data and also examine the effect of stability on the computation.

The wind stress vector, $\underset{\sim}{\tau}$, can be determined from the bulk aerodynamic expressions

$$\tau_x = \rho\, C_D\, V_a\, u_x \tag{9}$$

$$\tau_y = \rho\, C_D\, V_a\, u_y$$

Here τ_x and τ_y are the east-west components and north-south components of $\underset{\sim}{\tau}$, respectively; u_x and u_y are the components of the wind; and C_D is the drag coefficient for momentum. The Liu *et al.* (1979) algorithm was used to compute C_D. For a relative humidity of 70%, the dependence of C_D on the wind speed is shown in Fig. 9 for several air-sea temperature differences.

The behavior of C_D is similar to the behavior of C_H and C_E (see Fig. 2) except that C_D increases significantly at wind speeds larger than 10 m s^{-1}. As described in Liu *et al.* (1979) momentum can be transferred by pressure forces acting on the wind as it flows over the waves as well as by frictional stresses at the surface. The transfer by pressure forces tends to be strongly dependent on wind speed. However, molecular diffusion is the only mechanism available for the transfer of sensible and latent heat fluxes at the air-sea interface. Thus, the dependence of C_H and C_E on wind speed tends to be less than the dependence of C_D for moderate to strong winds.

Using the Liu *et al.* algorithm with the stability correction, we computed the climatological τ_x and τ_y at each weather ship with the sampling method and with a wind rose with an average wind speed for each of the 16 direction categories. Fig. 10 shows that the "direction-only" wind rose consistently underestimates the wind stress when $|\tau_x|$ or $|\tau_y|$ is greater than 0.5 dynes/cm^2. This relative error can be as large as 25% (see τ_x for Ship B). Thus, Hellerman's wind rose conclusions for neutral coefficient wind stress also applies for stability-corrected coefficients. Therefore, the wind stress should be computed either by the sampling method or by a wind rose with several speed categories for each direction.

To clarify the importance of stability we recomputed the wind stress with the sampling method for a neutral C_D. The results are shown for the climatological τ_x and τ_y in Fig. 11. The importance of stability is small and affects the results by less than 10% except for small values of τ_x and τ_y. If we reexamine Fig. 7, we see the effect of stability generally produces a smaller relative error for wind stress than for sensible and latent heat fluxes.

The effect of stability on τ_x and τ_y was also examined on an interannual basis for each ship for January and July. The effect of stability was found to be small in each case. This comparison is shown in Fig. 12 for the January τ_x at Ship D. It is apparent that for many applications the atmospheric surface boundary layer may be considered to be neutral when computing wind stresses on time scales of one month or longer over the open ocean.

The small errors caused by the use of neutral coefficients can be further reduced by substituting monthly-mean values of (T_s-T_a) and (q_s-q_a) into the transfer coefficient algorithms in place of the air-sea temperature and humidity differences at each observation time. Fig. 13 shows a comparison between τ_x computed with monthly-mean temperatures and humidities and τ_x computed using temperatures and humidities at each observation time at Ship D during successive Januaries. The errors of less than few percent are typical of the results we found at every ship position.

8. Summary and Discussion

We have shown that monthly-averaged values of middle-latitude and subtropical surface marine meteorological data can be used in bulk aerodynamic formulas to obtain the monthly-averaged surface fluxes of sensible and latent heat. These results confirm and extend the work of Budyko (1963) and his coworkers by including recent developments in boundary layer similarity theory to give a more realistic representation of the physical processes which are important to air-sea transfers of sensible and latent heat.

Our results also confirm that neutral transfer coefficients give reliable wind stress estimates over the ocean. If greater reliability is desired, this is easily achieved through the use of monthly-averaged air-sea temperature and humidity differences for the determination of the drag coefficient.

Similar results for heat flux and wind stress estimates should be obtained in tropical regions. Although this study does not include tropical marine data, our conclusions are valid for the small Bowen ratios (S/LE) and the moderate to low wind speeds which characterize large portions of the tropical oceans (see for example Fig. 3b).

We believe that the classical method of using wind roses and monthly-averaged state variables to estimate heat fluxes and wind stresses has a number of advantages over computing weighted sample means. These may prove to be valuable in describing and monitoring interannual and decadal climate fluctuations.

a. Data processing requirements. The computational effort required to process monthly-averaged data from ship-of-opportunity and ocean weather ships is much less than the effort needed to process all the individual surface reports. Therefore, the revision of heat flux estimates with an imporved transfer coefficient or updated mean field is a relatively simple task with the classical method.

b. Statistical reliability. The classical method may provide a more reliable estimate of the heat flux when the number of observations is small. Since both the covariance and the correlations of V_a with either (T_s-T_a) or (q_s-q_a) are known to be relatively small, the non-linearities in the flux estimates using the sampling method may lead to more statistical uncertainty in estimating the long-term average heat fluxes than the classical estimators.

c. Utilization of heteorogeneous data. The classical method may provide a simple way of combining non-simultaneous data from a variety of observing platforms. For example, satellite data does not yet provide accurate estimates of the air-sea temperature and humidity differences, but it may soon give useful information on the surface wind speed. The use of the classical method removes the restriction that the observations of wind-speed temperature and humidity fields be simultaneous and thus provides a straightforward way to incorporate the satellite observations into the heat flux estimates.

ACKNOWLEDGEMENTS

We would like to acknowledge the participation of Mr. E. I. Tollerud in background studies which led to this work. We would also like to acknowledge discussions with Drs. J. W. Deardorff, W. T. Liu, and C. A. Paulson concerning the determination of the bulk aerodynamic transfer coefficients and would like to thank Drs. Deardorff and Paulson for their comments on a draft of the manuscript. Mr. T. McCart assisted with the data processing tasks.

This research was supported primarily by a grant from the National Aeronautics and Space Administration (NSG 5308). Funds were also provided by the National Science Foundation under Grant No. ATM 77-26406-03.

REFERENCES

Brook, R.R., 1978: The influence of water vapor fluctuations on turbulent fluxes. Bound.-Layer Meteor., 15, 481-487.

Budyko, M.I., Ed., 1963: Atlas of the Heat Balance of the Earth. Akad. Nauk SSSR, Prezidium. Mezhduvedomstvennyi Geofiz. Komitet. Also, Guide to the Atlas of the Heat Balance of the Earth, Transl. by I.A. Donehoo, U.S. Wea. Bur., WB/T-106, Washington, D.C.

________, 1974: Climate and Life. Acad. Press, 508 pp.

________, and L.S. Gandin, 1966: On determination of the turbulent heat exchange between ocean and atmosphere. Meteorol. Gidrol., 11, 15-25. [MGA 18.7-436]

Bunker, A.F., 1976: Computations of surface energy flux and annual air-sea interaction cycles of the North Atlantic Ocean. Mon. Wea. Rev., 104, 1122-1140.

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.

Deardorff, J.A., 1972: Parameterization of the planetary boundary layer for use in general circulation models. Mon. Wea. Rev., 100, 93-106.

Dyer, A.J., and B.B. Hicks, 1970: Flux-gradient relationships in the constant flux layer. Quart. J. Roy. Meteor. Soc., 96, 715-721.

Fissel, D.B., S. Pond, and M. Miyake, 1977: Computation of surface fluxes from climatological and synoptic data. Mon. Wea. Rev., 105, 26-36.

Hellerman, S., 1965: Computations of wind stress fields over the Atlantic Ocean. Mon. Wea. Rev., 93, 239-244.

__________, 1967: An updated estimate of the wind stress on the world ocean. Mon. Wea. Rev., 95, 607-626; see also Corrigendum (1968), 96, 63-74.

Kondo, J., 1975: Air-sea bulk transfer coefficients in diabatic conditions. Bound.-Layer Meteor., 9, 91-112.

Kraus, E.B., and R.E. Morrison, 1966: Local interactions between the sea and the air at monthly and annual time scales. Quart. J. Roy. Meteor. Soc., 92, 114-127.

Liu, W.T., K.B. Katsaros, and J.A. Businger, 1979: Bulk parameterization of air-sea exchanges of heat and water vapor including the molecular constraints at the interface. J. Atmos. Sci., 36, 1722-1735.

Malkus, J.S., 1962: Large-scale interactions. In, The Sea, Vol. 1, M.N. Hill, Ed., Wiley-Interscience, 88-294.

Oort, A.H., and T.H. VonderHaar, 1976: On the observed annual cycle in the ocean-atmosphere heat balance over the Northern Hemisphere. J. Phys. Oceanogr., 6, 781-800.

Pond, S., D.B. Fissel, and C.A. Paulson, 1974: A note on bulk aerodynamic coefficients for sensible heat and moisture fluxes. Bound.-Layer Meteor., 6, 333-339.

Sverdrup, H.U., M.W. Johnson, and R.H. Fleming, 1942: The Oceans. Prentice-Hall, 1087 pp.

Figure Legends

Fig. 1a. The transfer coefficient for sensible heat from Liu *et al.* (1979) as a function of wind speed at 10m above the sea surface for various air-sea temperature differences ($\Delta T = T_s - T_a$). The neutral surface boundary layer curve is also indicated. All curves assume a boundary-layer relative humidity of 70%.

Fig. 1b. Same as Fig. 1a except for latent heat.

Fig. 2. Positions of ocean weather ships used in this study superimposed on ocean surface currents from Sverdrup *et al.* (1942).

Fig. 3a. Climatological sensible and latent heat flux estimates for January for each ship computed by the sampling method (abscissa) and the classical method (ordinate). The dashed lines indicate relative errors of 10%. See text for details.

Fig. 3b. Same as Fig. 3a except for July.

Fig. 4. Transfer coefficients from Liu *et al.* (1979) for which $z_o = z_T = z_q$ (see also Fig. 1a).

Fig 5. Climatological sensible and latent heat fluxes as in Fig. 3a but computed with transfer coefficients for which $z_o = z_T = z_q$.

Fig. 6. Climatological sensible and latent heat fluxes as in Fig. 3a except that only nighttime observations are used.

Fig. 7. Climatological sensible and latent heat fluxes as in Fig. 3a except that values along the ordinate are computed by the sampling method with neutral transfer coefficients.

Fig. 8a. Monthly averaged sensible and latent heat fluxes computed by the sampling and classical methods for each January at Ship D.

Fig. 8b. Same as Fig. 8a except at Ship N.

Fig. 9. Same as Fig. 1a except for the drag coefficient from Liu *et al.* (1979).

Fig. 10. Climatological wind stresses computed from the sampling method (abscissa) and a "direction-only" wind rose (ordinate). See also Fig. 3a.

Fig. 11. Same as Fig. 10 except that the ordinate indicates the wind stress computed from the sampling method with neutral coefficients.

Fig. 12. Monthly-averaged τ_x computed by the sampling method for stability corrected and neutral momentum transfer coefficients for each January at Ship D.

Fig. 13. Same as Fig. 12 except τ_x $(\overline{\Delta T}, \overline{\Delta q})$ was computed from stability corrected momentum coefficients defined by the monthly-averaged air-sea temperature and humidity differences, $\overline{\Delta T} = (\overline{T_s - T_a})$ and $\overline{\Delta q} = (\overline{q_s - q_a})$.

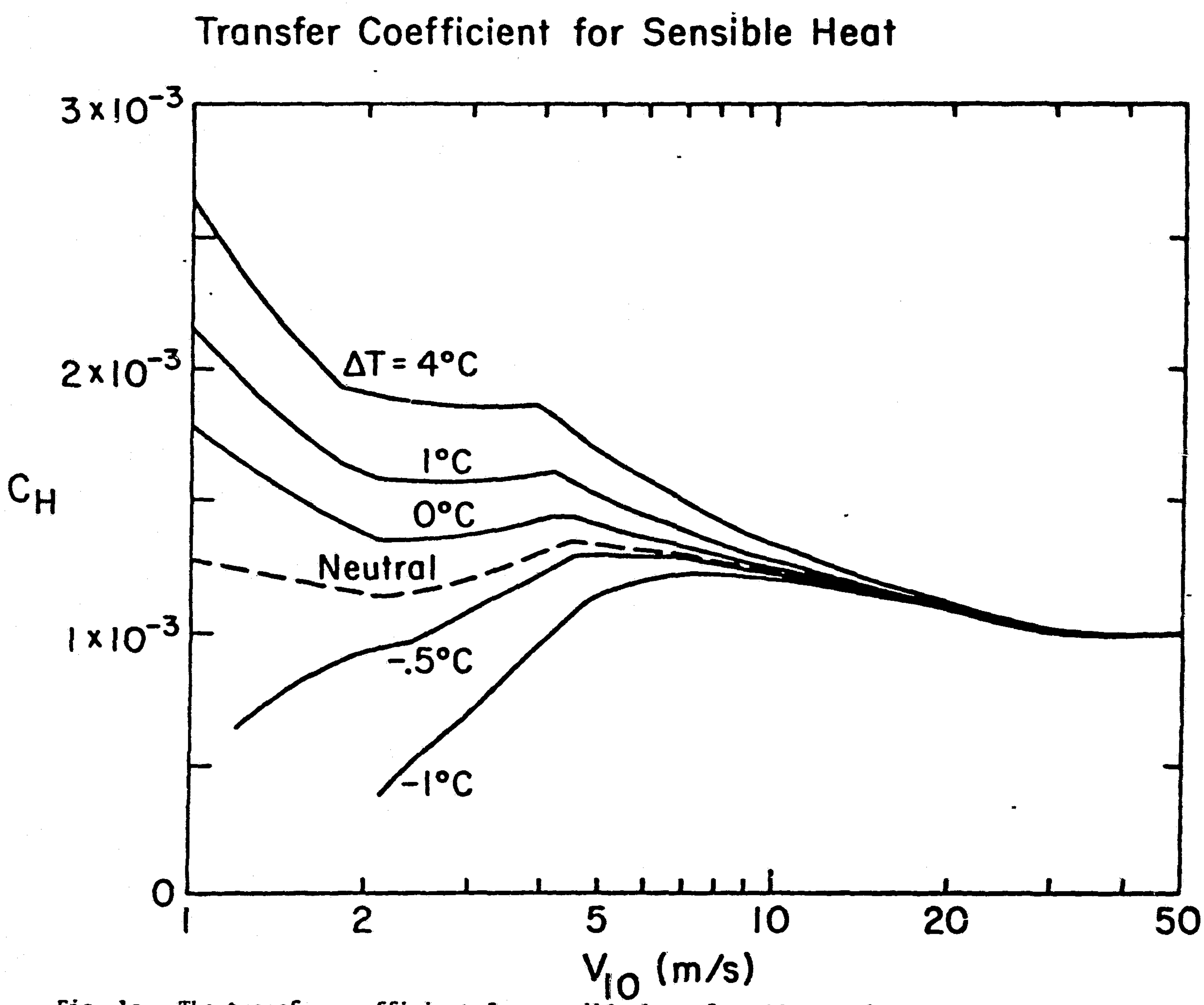


Fig. 1a. The transfer coefficient for sensible heat from Liu *et al.* (1979) as a function of wind speed at 10m above the sea surface for various air-sea temperature differences ($\Delta T=T_s-T_a$). The neutral surface boundary layer curve is also indicated. All curves assume a boundary-layer relative humidity of 70%.

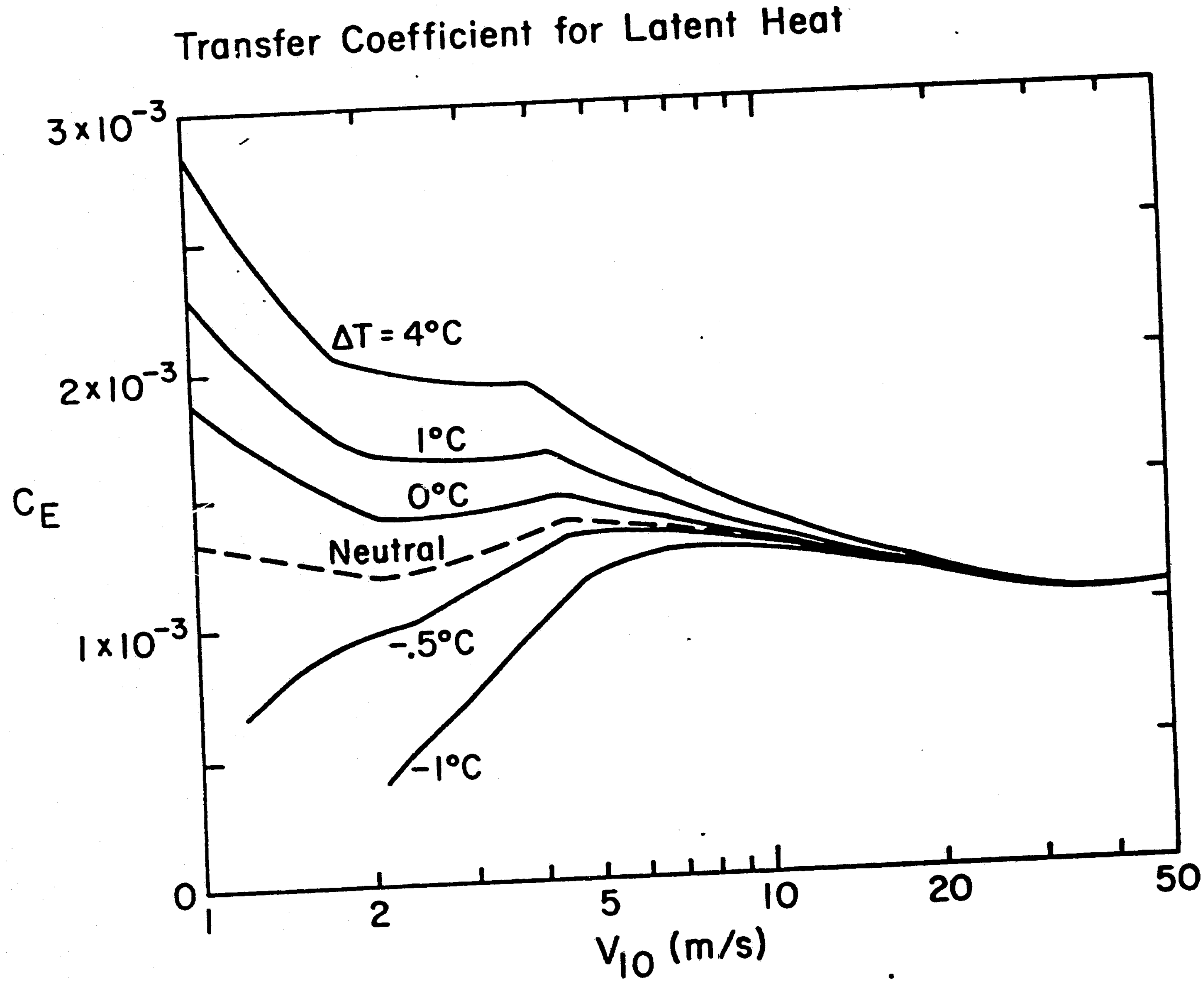


Fig. 1b. Same as Fig. 1a except for latent heat.

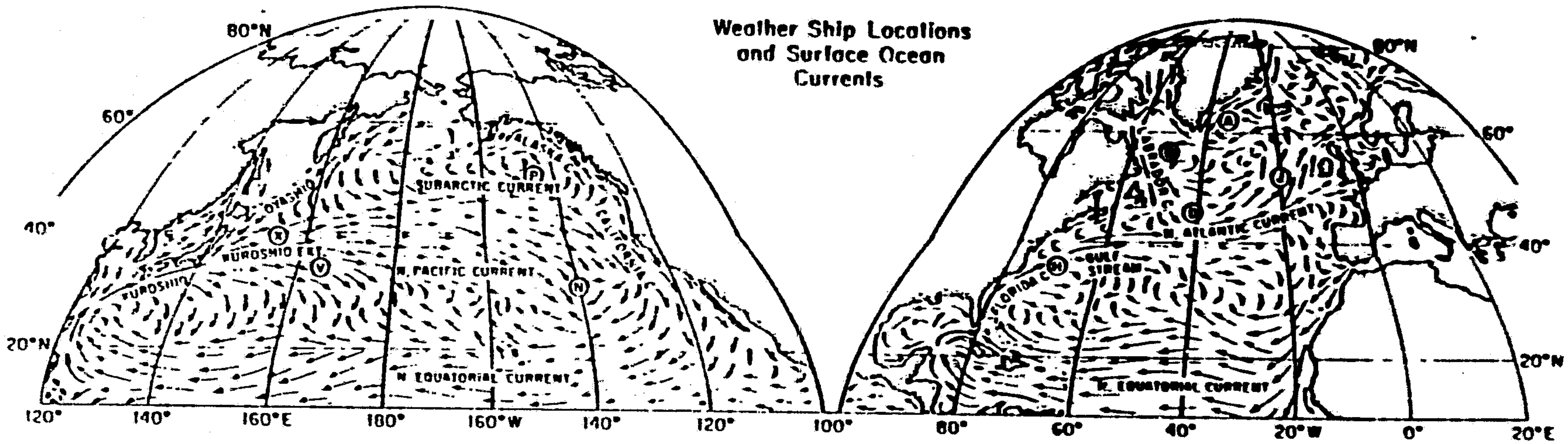


Fig. 2. Positions of ocean weather ships used in this study superimposed on ocean surface currents from Sverdrup *et al.* (1942).

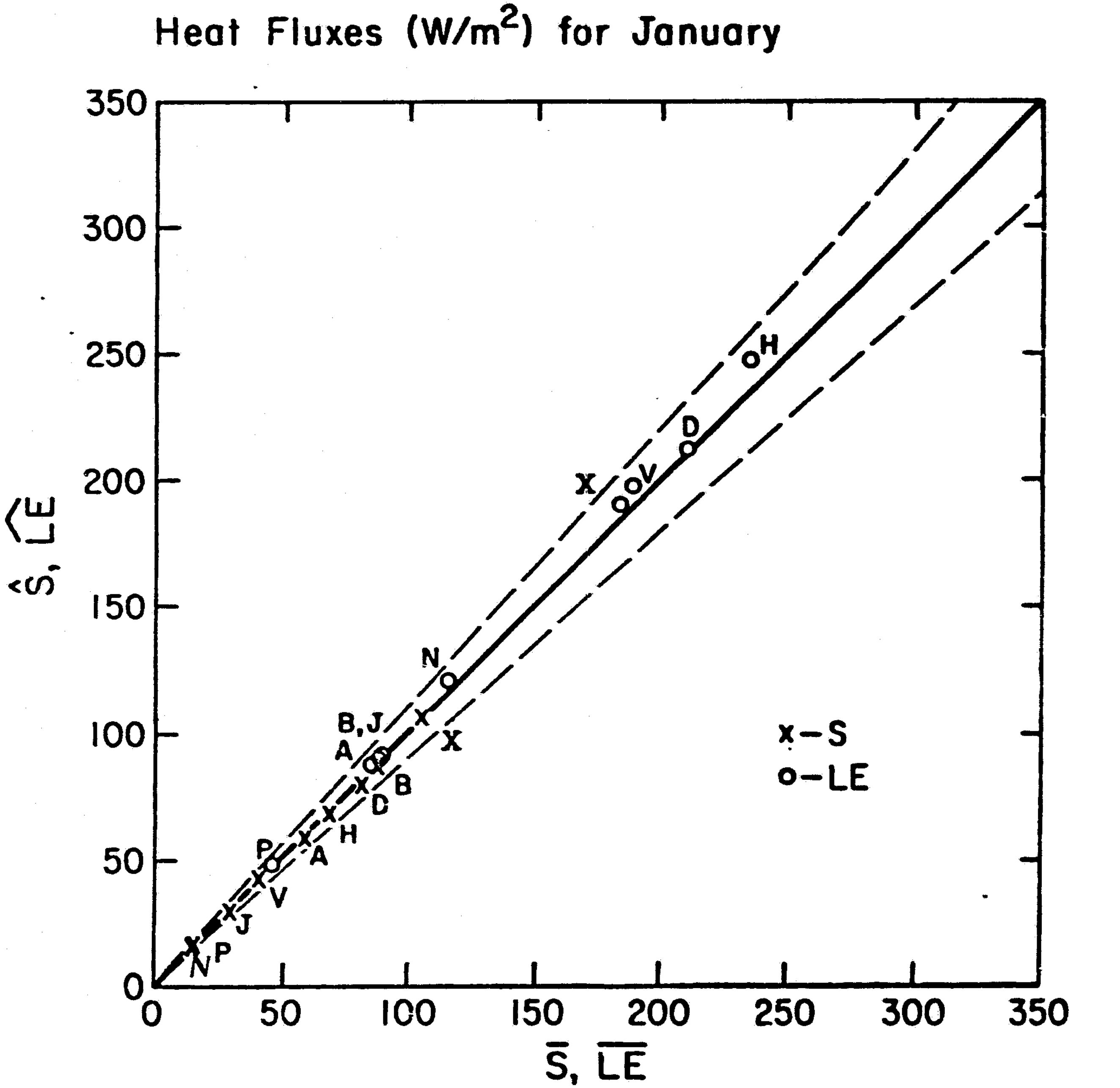


Fig. 3a. Climatological sensible and latent heat flux estimates for January for each ship computed by the sampling method (abscissa) and the classical method (ordinate). The dashed lines indicate relative errors of 10%. See text for details.

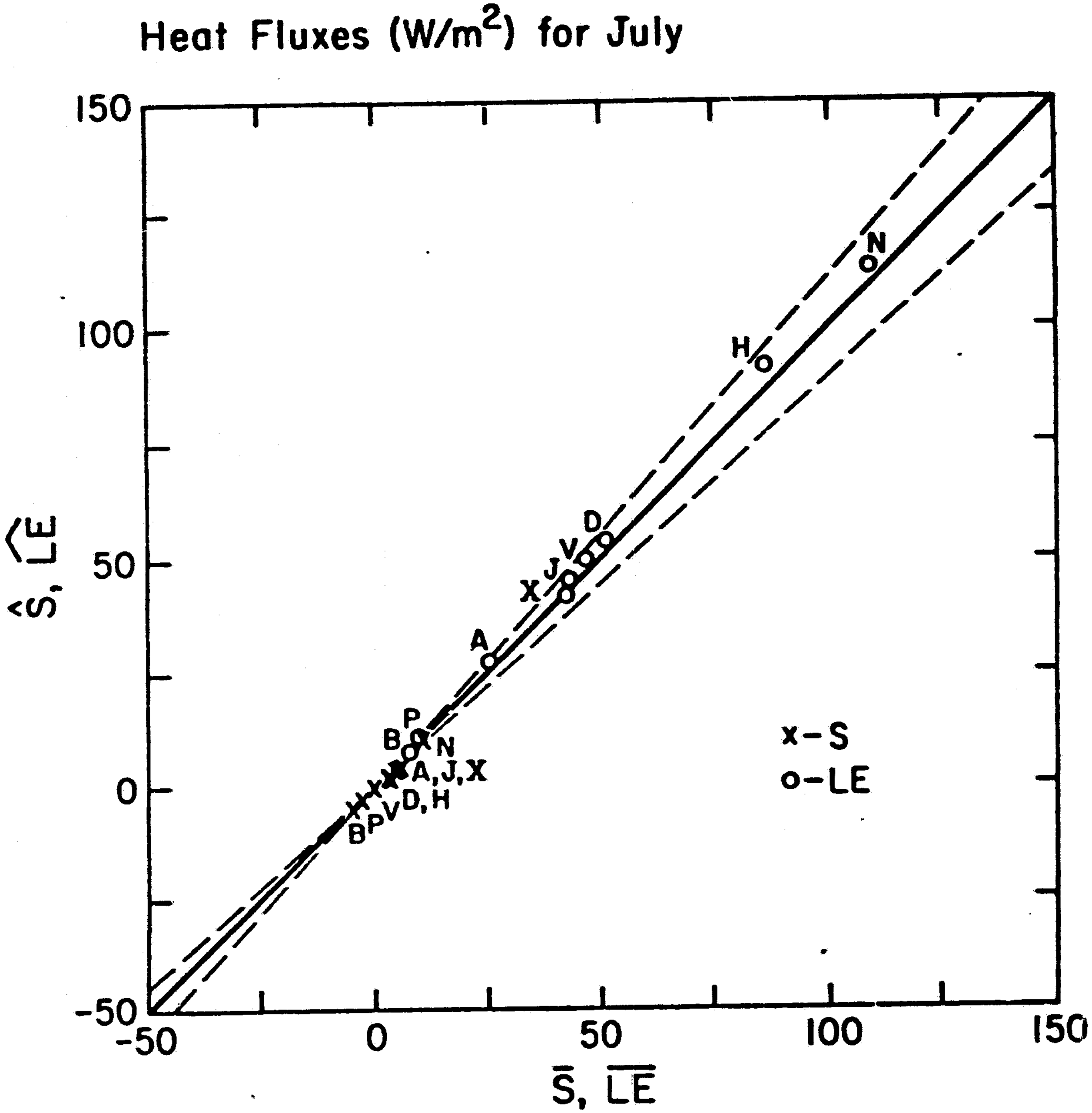


Fig. 3b. Same as Fig. 3a except for July.

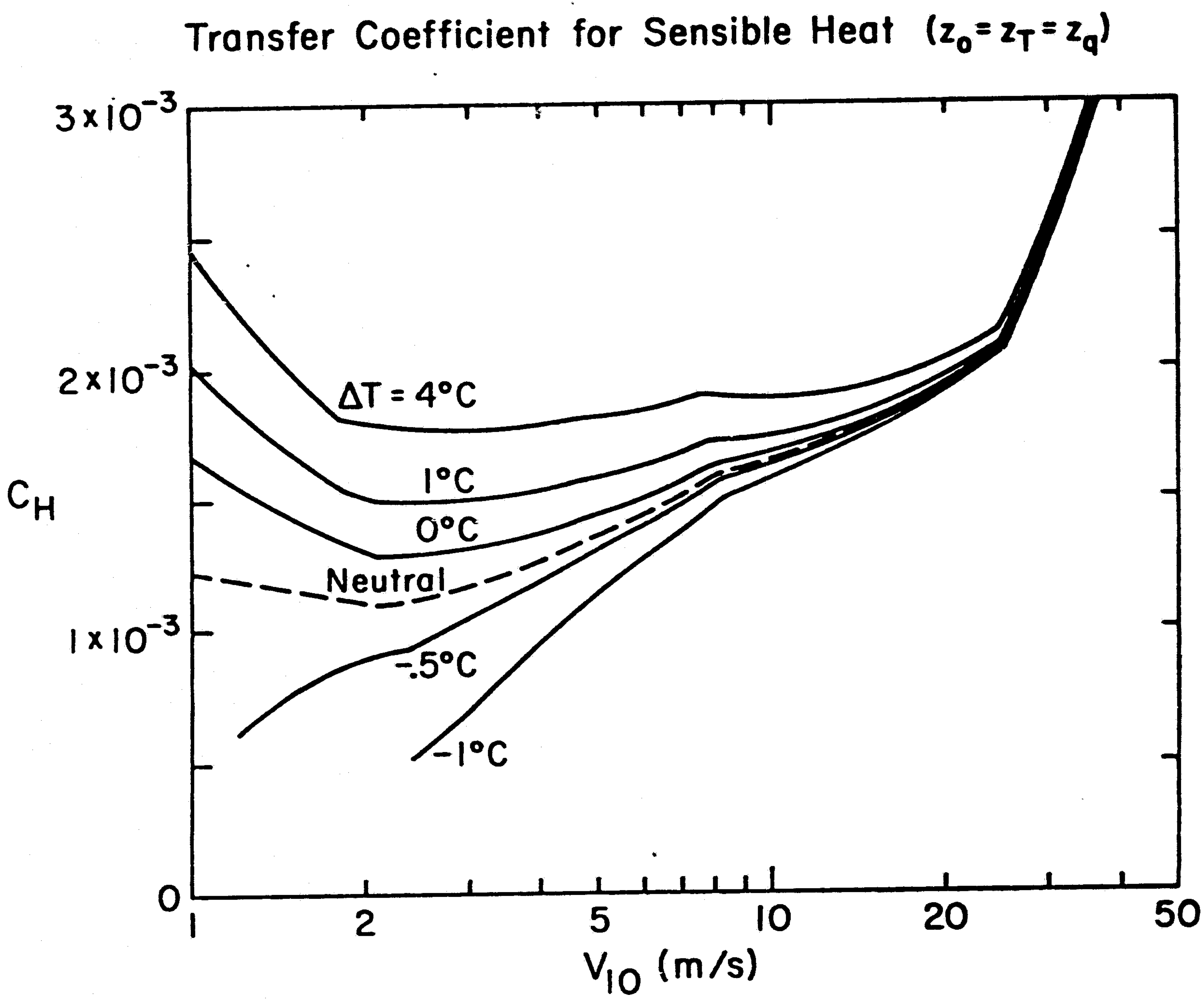


Fig. 4. Transfer coefficients from Liu *et al.* for which $z_o = z_T = z_q$ (see also Fig. 1a).

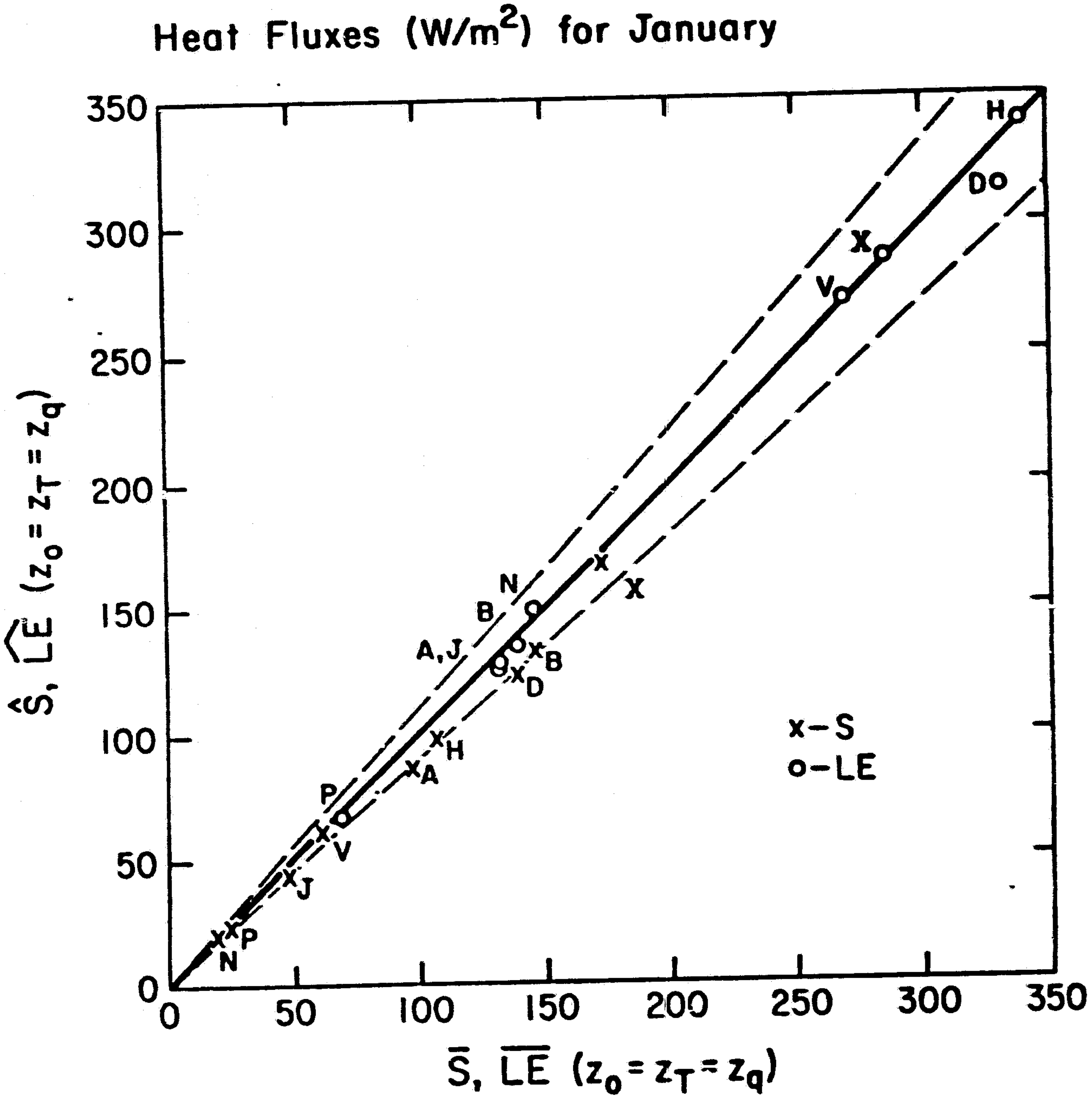


Fig. 5. Climatological sensible and latent heat fluxes as in Fig. 3a but computed with transfer coefficients for which $z_o = z_T = z_q$.

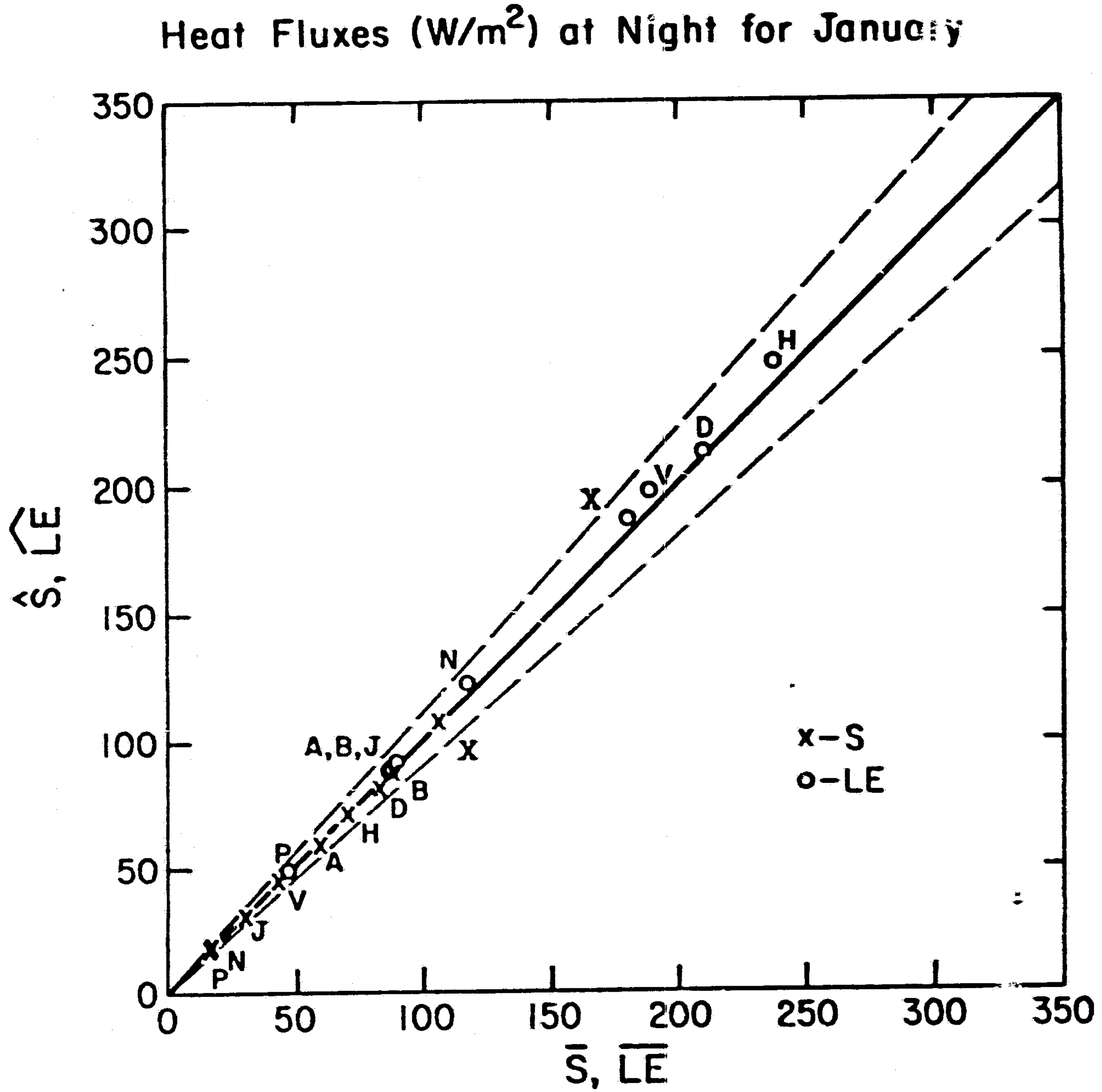


Fig. 6. Climatological sensible and latent heat fluxes as in Fig. 3a except that only nighttime observations are used.

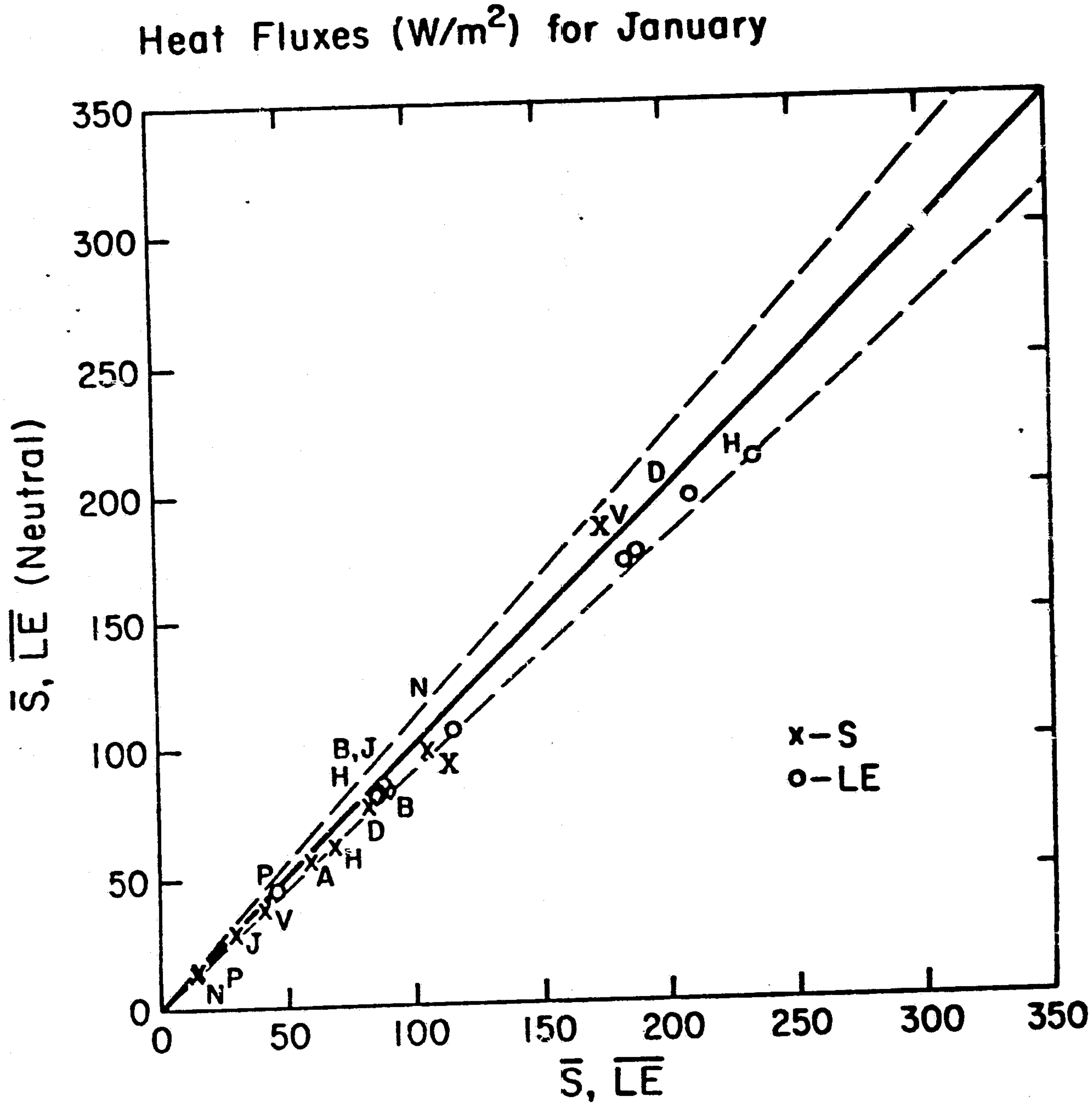


Fig. 7. Climatological sensible and latent heat fluxes as in Fig. 3a except that values along the ordinate are computed by the sampling method with neutral transfer coefficients.

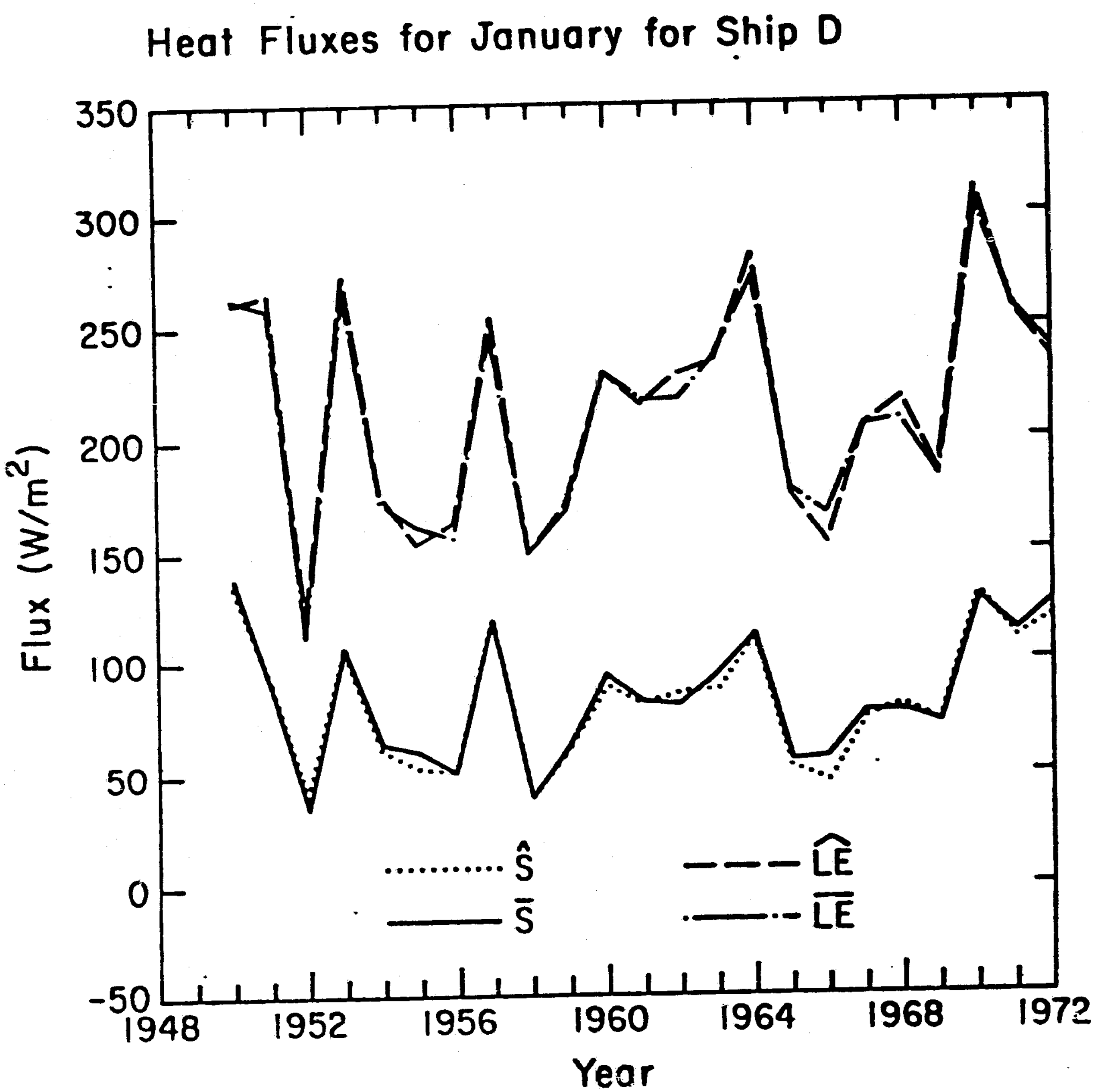
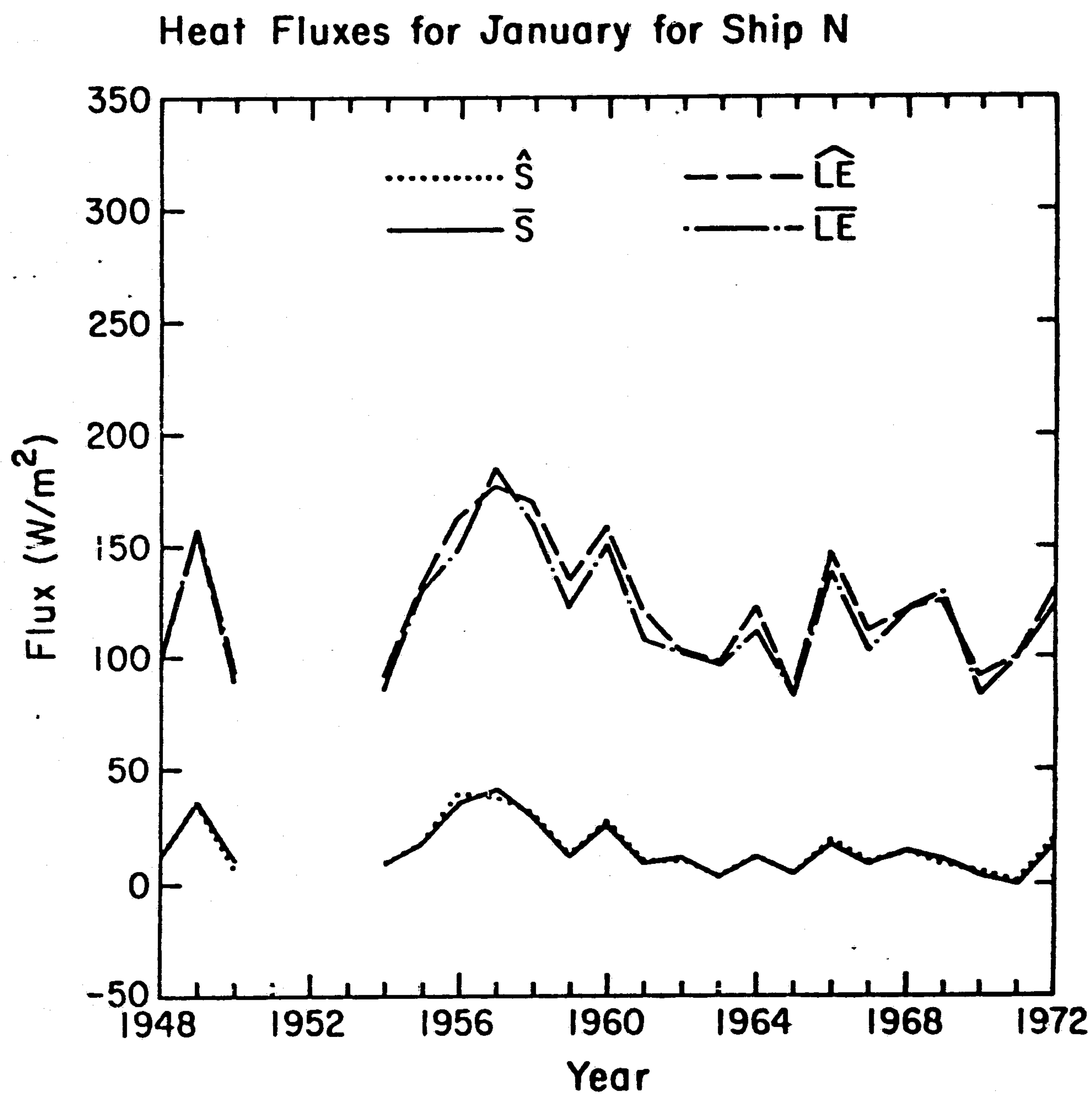


Fig. 8a. Monthly averaged sensible and latent heat fluxes computed by the sampling and classical methods for each January at Ship D.



Fib. 8b. Same as Fig. 8a except at Ship N.

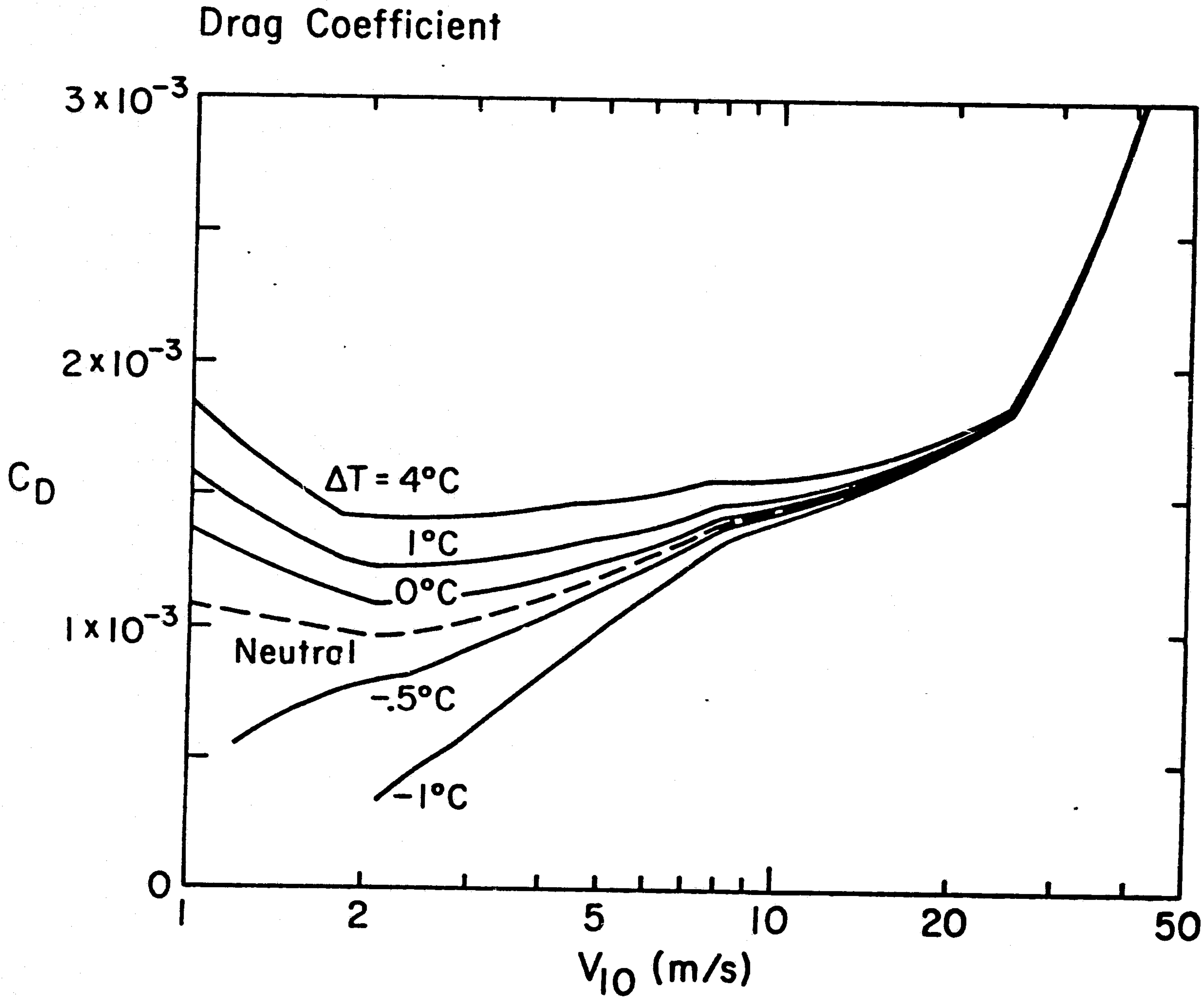


Fig. 9. Same as Fig. 1a except for the drag coefficient from Liu *et al.* (1979).

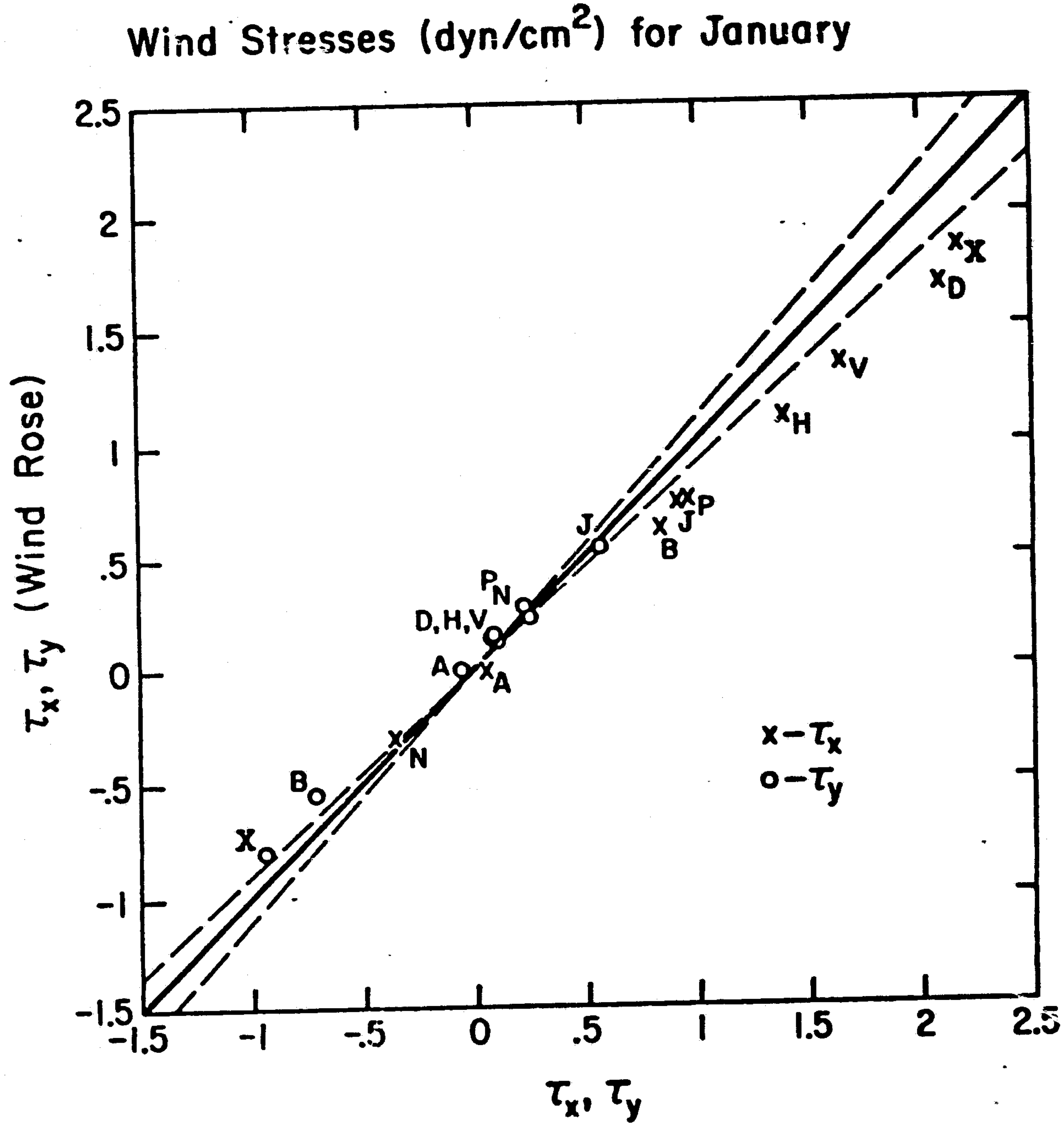


Fig. 10. Climatological wind stresses computed from the sampling method (abscissa) and a "direction-only" wind rose (ordinate). See also Fig. 3a.

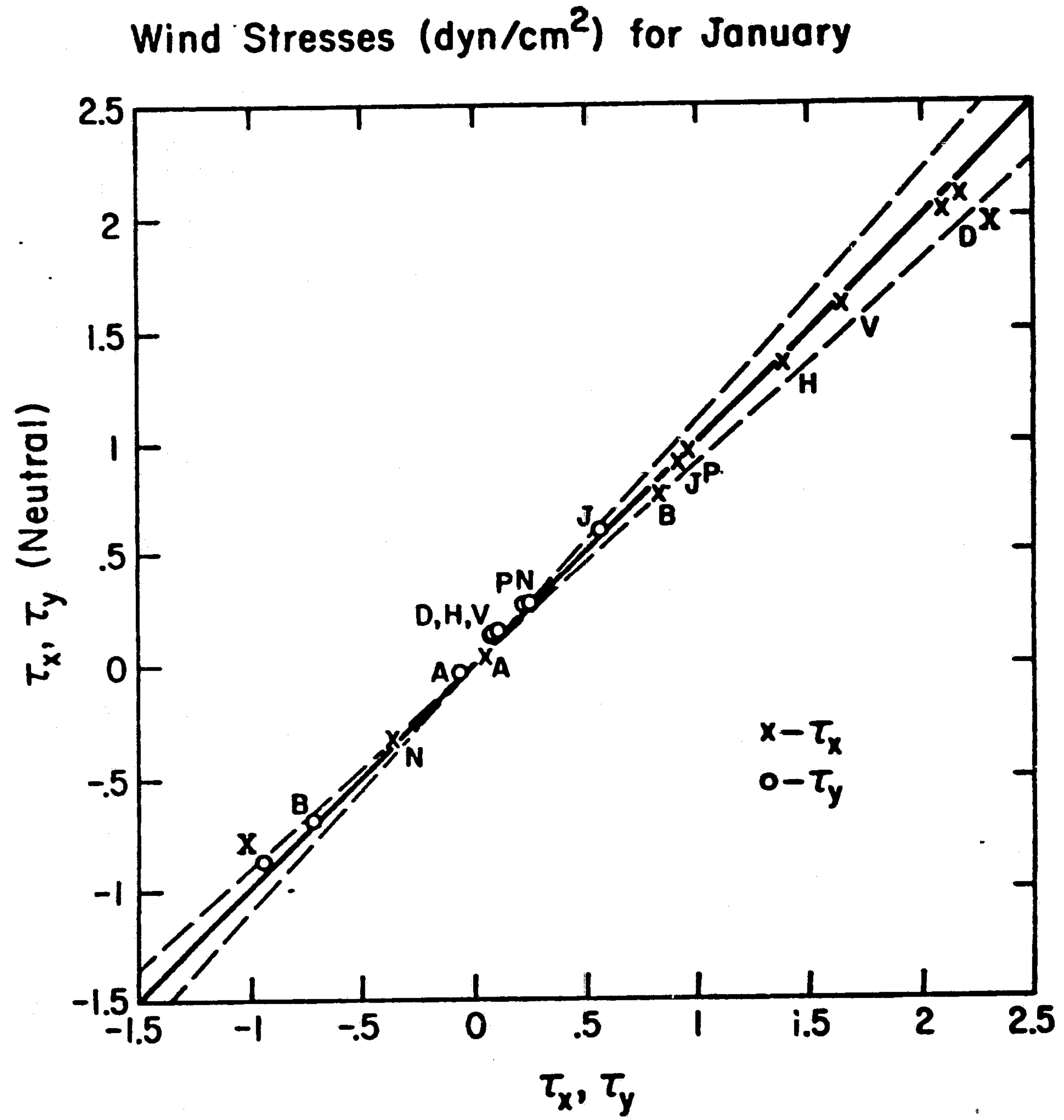


Fig. 11. Same as Fig. 10 except that the ordinate indicates the wind stress computed from the sampling method with neutral coefficients.

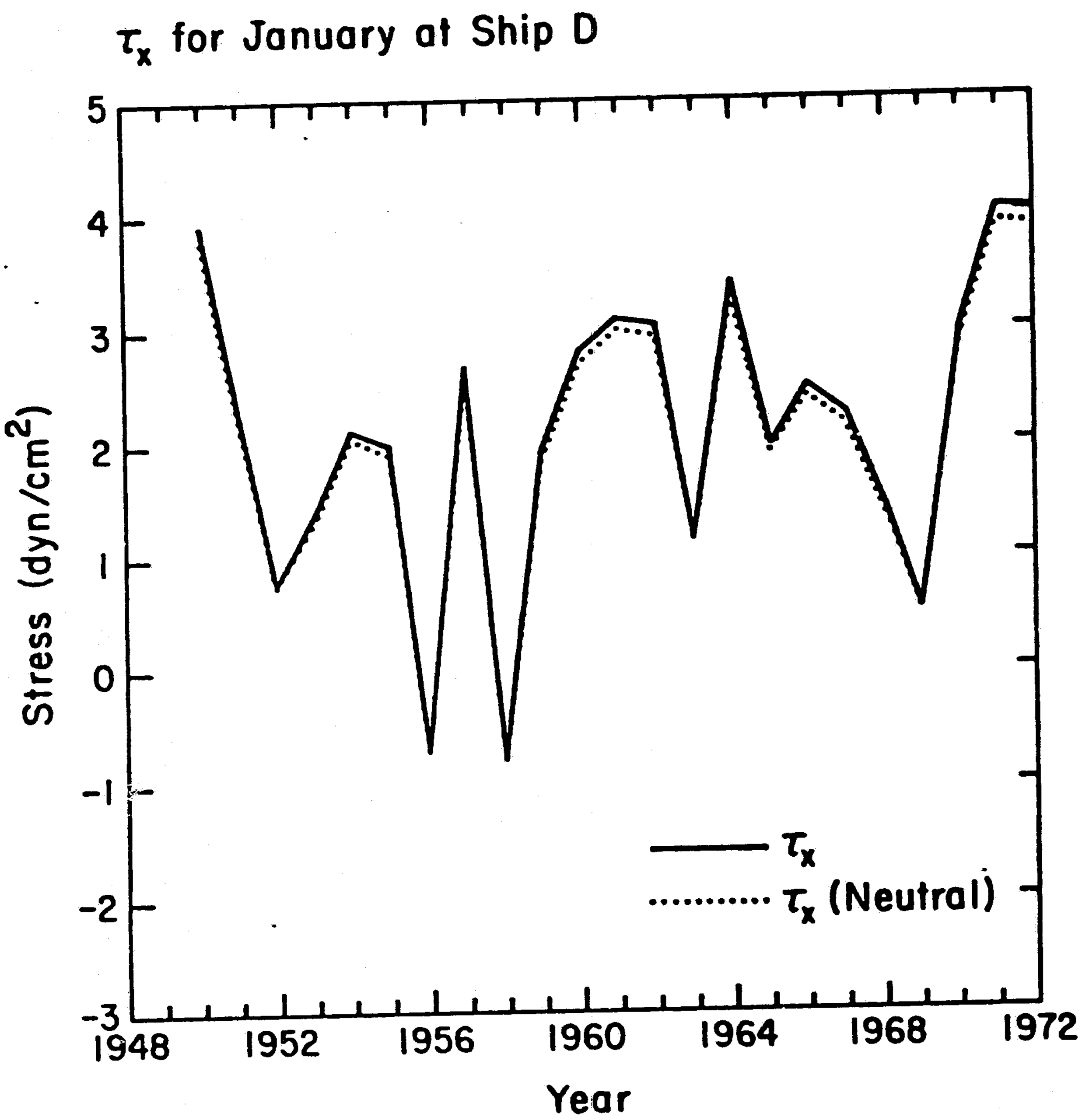


Fig. 12. Monthly-averaged τ_x computed by the sampling method for stability corrected and neutral momentum transfer coefficients for each January at Ship D.

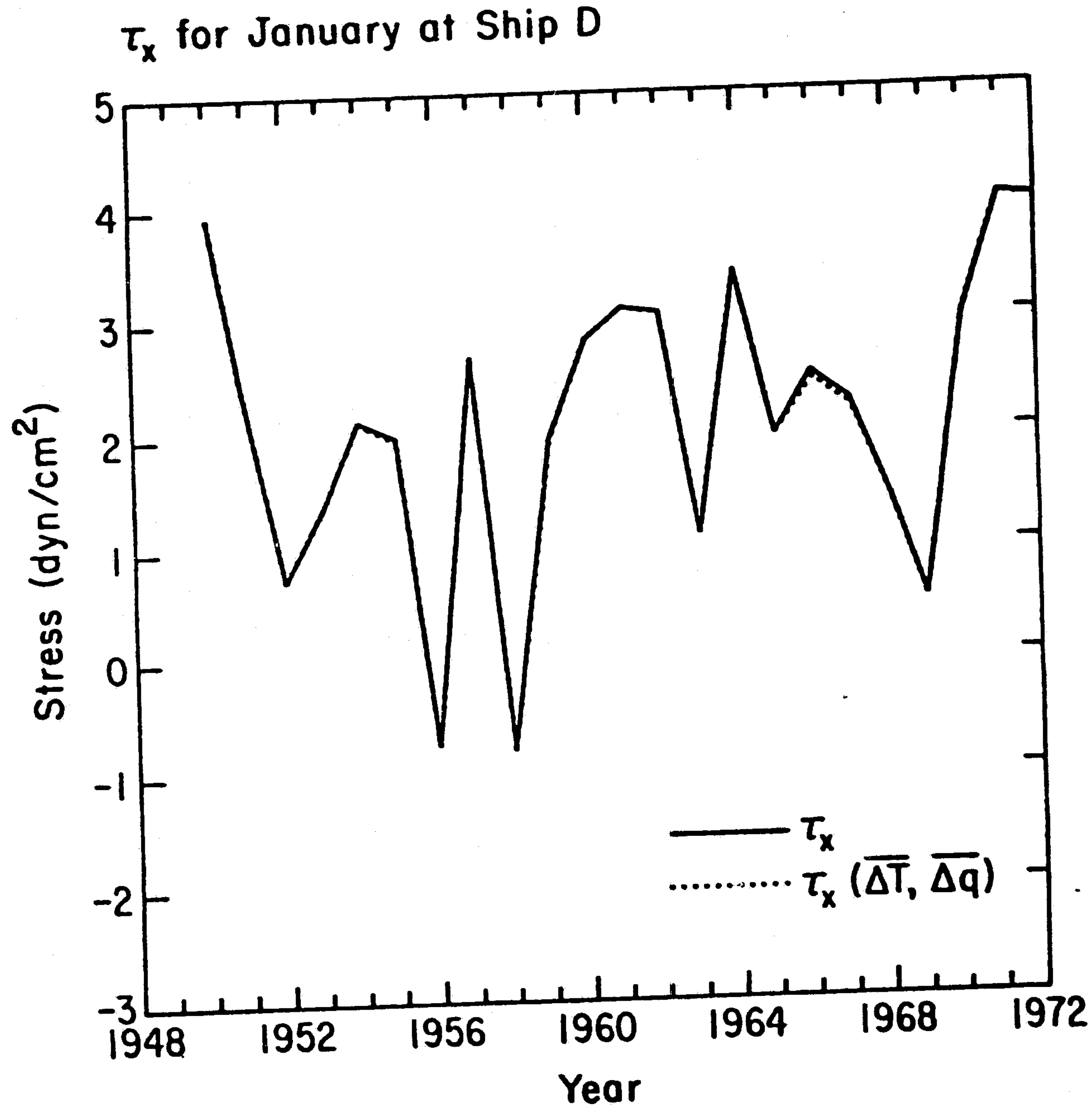


Fig. 13. Same as Fig. 12 except τ_x ($\overline{\Delta T},\overline{\Delta q}$) was computed from stability corrected momentum coefficients defined by the monthly-averaged air-sea temperature and humidity differences, $\overline{\Delta T}=(\overline{T_s-T_a})$ and $\overline{\Delta q}=(\overline{q_s-q_a})$.