

# Air–sea interaction processes in warm and cold sectors of extratropical cyclonic storms observed during FASTEX

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## SUMMARY

Mid-Atlantic measurements from the research vessel (RV) *Knorr* during the Fronts and Atlantic Storm-Tracks Experiment (FASTEX) are presented, and then used to examine the near-surface environment and air–sea interaction processes during the passage of ten frontal systems. This dataset includes measurements of the surface momentum, sensible-heat and moisture fluxes obtained from three different methods. The inertial dissipation (ID) drag coefficients from the RV *Knorr* are consistent with the ID data from other measurements in open-ocean storm environments. The covariance drag coefficients are generally larger than the ID values, indicating either the presence of flow distortion problems in the covariance data or a failure of the assumptions inherent to the ID technique at these higher wind speeds. Estimates of the wind speed dependence of the momentum, sensible-heat and latent-heat transfer coefficients are based on averaged values from the two methods. These measurements: (i) contribute significantly to the limited set of surface flux measurements for 10 m neutral winds in the 15–21 m s<sup>-1</sup> range; (ii) contain the only ship-based covariance flux measurements successfully obtained in an open-ocean, high wind speed, storm environment; and (iii) include coincident wave-height measurements.

The relationships between the surface layer and the synoptic atmospheric environment is examined using composites of atmospheric and oceanic surface-layer characteristics computed in ten storms for which the RV *Knorr* passed through the open-wave warm sector and the cold front. These composites show minima in the sensible- and latent-heat fluxes, and a maximum in the momentum flux, just before the frontal passage during the warm-sector peak in wind speed. A second momentum flux maximum of comparable magnitude occurs in the middle of the post-frontal regime. Though the warm-sector sensible-heat flux minimum is slightly negative, the sum of the two heat fluxes is positive, suggesting a positive impact on the synoptic development of these storms. Wave heights increase steadily from the eastern half of the warm sector to the frontal passage, remaining high through most of the post-frontal regime before decreasing. Differences between covariance and ID stresses are largest during the times bracketing the cold front when the wave heights and stresses are large. Differences between covariance and bulk stresses are greatest in the pre-frontal low-level jet, when the frequency of waves with periods of 6–9 s maximizes, and in the post-frontal regime where the wind direction veers. Systematic differences between stress direction and wind direction are observed.

KEYWORDS: Observations RV *Knorr* Storm composites Surface fluxes Surface layer

## 1. INTRODUCTION

Fluxes of momentum, heat, and moisture between the atmosphere and the ocean are important for climatological and synoptic processes. Direct measurements of these fluxes not only describe the air–sea interaction for that time and place, but also provide the data necessary to develop flux parametrizations for use in climate and weather forecasting models. Hence, great efforts have been made to obtain open-ocean surface flux measurements. Most of the successful air–sea flux measurement campaigns have been in the more benign tropical and subtropical low-to-medium wind speed regimes (e.g. Young *et al.* 1992; Chertock *et al.* 1993; Fairall *et al.* 1996, 2000; Webster *et al.* 2002). A few measurement campaigns have successfully collected data in the more difficult extratropical storm environments (e.g. Large and Pond 1981, 1982; Yelland and Taylor 1996; Hauser *et al.* 2003), although the number of data points acquired at wind speeds higher than 15 m s<sup>-1</sup> are limited. Furthermore, the most direct measurements are those using the covariance technique, and the number of covariance flux observations

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at high wind speeds are even fewer because of the difficulties involved in making such measurements in an environment with large surface waves.

Air–sea flux measurements are typically classified by stability and wind speed to make them useful for comparison to, and modification of, surface flux models based on Monin–Obukhov similarity theory. In this way, the dependence of the turbulent exchange coefficients on the stability and neutral wind speed can be quantified and incorporated into flux parametrization schemes. Brunke *et al.* (2003) recently compared the performance of such schemes, finding that the least problematic ones are from the Coupled Ocean–Atmosphere Response Experiment version 3.0 (COARE 3.0; Fairall *et al.* 1996, 2003), the European Centre for Medium-Range Weather Forecasting (ECMWF; Beljaars 1995), the Goddard Earth Observing System re-analysis version 1 (GEOS-1; Large and Pond 1981; Kondo 1975) and the University of Arizona (Zeng *et al.* 1998). However, classifying flux measurements with respect to physical atmospheric boundaries or regimes can provide greater physical understanding of the processes producing the flux environment and the significance of the fluxes for modifying atmospheric processes. Classifying flux data in this manner is much rarer. Large and Pond (1981, 1982) utilized time series of case-studies of data collected on a mast 10 km offshore of Nova Scotia and data from the Canadian Coast-Guard ship *Quadra* in the north-east Pacific Ocean, to obtain physical insights into the behaviour of fluxes in relationship to atmospheric changes and to provide an assessment of errors in their flux parametrizations. Their cases showed that much of the scatter in flux measurements displayed traditionally is not random but, instead, can be physically explained by various transitions, such as changes in wind and wave directions. The flux environment provided by a change in wind direction and wind speed in one case was also utilized by Rieder and Smith (1998) to show how the surface stress measurements respond to the transitional differences between various parts of the wave spectrum. In extratropical environments, the location of fluxes with respect to frontal systems emphasizes the process by which they are generated, and can also provide insight into their significance for storm development and frontal zone modification. Modelling studies have shown that fluxes in specific regions of storms have different impacts on the storm evolution (e.g. Reed and Albright 1986; Langland *et al.* 1995).

In this paper, we first present the surface flux measurements made onboard the research vessel (RV) *Knorr* during the Fronts and Atlantic Storm-Tracks Experiment (FASTEX). Although this high wind speed dataset has been used for parametrization development (Hare *et al.* 1999) and in comparisons of flux parametrizations (Brunke *et al.* 2003), a complete description of the data collection and processing has not been formally presented. The objectives of FASTEX were to collect data relevant to the development of midlatitude atmospheric frontal waves that impact Europe (Joly *et al.* 1999). Extensive measurements of the atmospheric boundary-layer structure, air–sea interaction processes, and ocean surface were obtained from four ships strategically placed in the central North Atlantic Ocean (35–65°N, 10–60°W) during January and February 1997. Two ships were specifically instrumented to make surface-layer flux measurements: the French RV *Le Suroît*, and the US RV *Knorr*. The data from RV *Le Suroît* have been reported elsewhere (Eymard *et al.* 1999; Brunke *et al.* 2003), and are often referred to as the Couplage avec l’Atmosphère en Conditions Hivernales (CATCH) dataset. CATCH was performed simultaneously and collocated with FASTEX. The surface-layer data from the RV *Knorr* are the focus of this paper. The RV *Knorr* measurements include the only open-ocean covariance flux measurements in high wind ( $> 15 \text{ m s}^{-1}$ ) conditions, other than those obtained for long-fetch directions from spar buoys within 50 km of the coast, i.e. within an atmospheric Rossby radius of deformation

( $\sim 100$  km; Smith 1980; Large and Pond 1981, 1982; Dupuis *et al.* 2003). During FASTEX, 10 to 20 storm systems passed each ship, with surface winds of  $15\text{--}30\text{ m s}^{-1}$  associated with each passage. Hence, the FASTEX oceanic environment consisted of frequent episodes of strong winds and large waves, with varying stability regimes dictated solely by the evolving synoptic conditions and ocean surface temperature.

After presenting this dataset in the traditional manner in sections 2 and 3, we then use it to examine how air-sea interaction processes are modulated by the storms, and how these processes in turn impact the atmospheric conditions important for the development of these storm systems, especially in the dynamically important warm-sector region. Specifically, we examine the surface-layer processes occurring in the warm sector and post cold-frontal baroclinic regimes of the extratropical cyclones encountered in the North Atlantic. A 'compositing' method is presented in section 4; this is used to facilitate the positioning of the physical processes relative to the surface cold front, the warm sector, or the post-frontal sector of the baroclinic zone. In section 5, we discuss the atmospheric surface-layer structure and its variation relative to the synoptically modulated conditions. In section 6, we describe the ocean-surface environment and discuss the air-sea interaction, again relating these interaction processes to the synoptic structure through the compositing. A discussion of the implications of the compositing results is presented in section 7, and conclusions are given in section 8.

## 2. MEASUREMENTS AND DATA PROCESSING

### (a) Instrumentation

The FASTEX campaign included the deployment of four ships, 15–18 buoys, and five aircraft. The work presented here involves the data collected onboard the Woods Hole Institute of Oceanography (WHOI) 85 m RV *Knorr*, deployed in the near-upstream (NUS) area of FASTEX in the central Atlantic Ocean (Joly *et al.* 1996).

An integral part of the RV *Knorr* deployment during FASTEX was the National Oceanic and Atmospheric Administration Environmental Technology Laboratory (NOAA/ETL) Air-Sea Interaction Group's suite of surface-layer turbulent flux instrumentation. This system has been developed during the past decade, and NOAA/ETL has deployed some form of the system on over 20 research cruises. The most vital element of the flux system is the ultrasonic anemometer/thermometer (Gill-Solent R2), which samples at over 20 Hz. Also of critical value to the measurement system is a fast-response infrared (IR) hygrometer (Ophir IR-2000), which provides *in situ* measurements of small-scale humidity fluctuations. Both of these instruments have proved to be rugged enough to withstand months in the harsh marine environment (Hare *et al.* 2000). During FASTEX, these instruments were mounted at the top of a sturdy, reinforced, pivoting, 12.5 m mast placed 3 m to the rear of the bow (Figs. 1(a) and (b)), allowing the collection of data with minimal flow distortion whenever the bow was pointed into the wind. The sensors were 15.5–19.2 m above the nominal water line. The hygrometer optics were cleaned with a remotely operated water spray when the data indicated that salt accumulation had become significant (Fairall and Young 1991). Shipboard direct covariance measurements of turbulent fluxes were possible from this system because of the routine application of a platform-motion measurement package (Systron-Donner, Motion-Pak) and a motion-correction algorithm (Anctil *et al.* 1994; Edson *et al.* 1998). The Motion-Pak consists of a slender package of three-axis accelerometers and rotation-rate sensors mounted in a canister in close proximity to the sonic anemometer. Details on the turbulent flux instruments and on the process for platform-motion correction can be found in Edson *et al.* (1998) and Fairall *et al.* (1997).

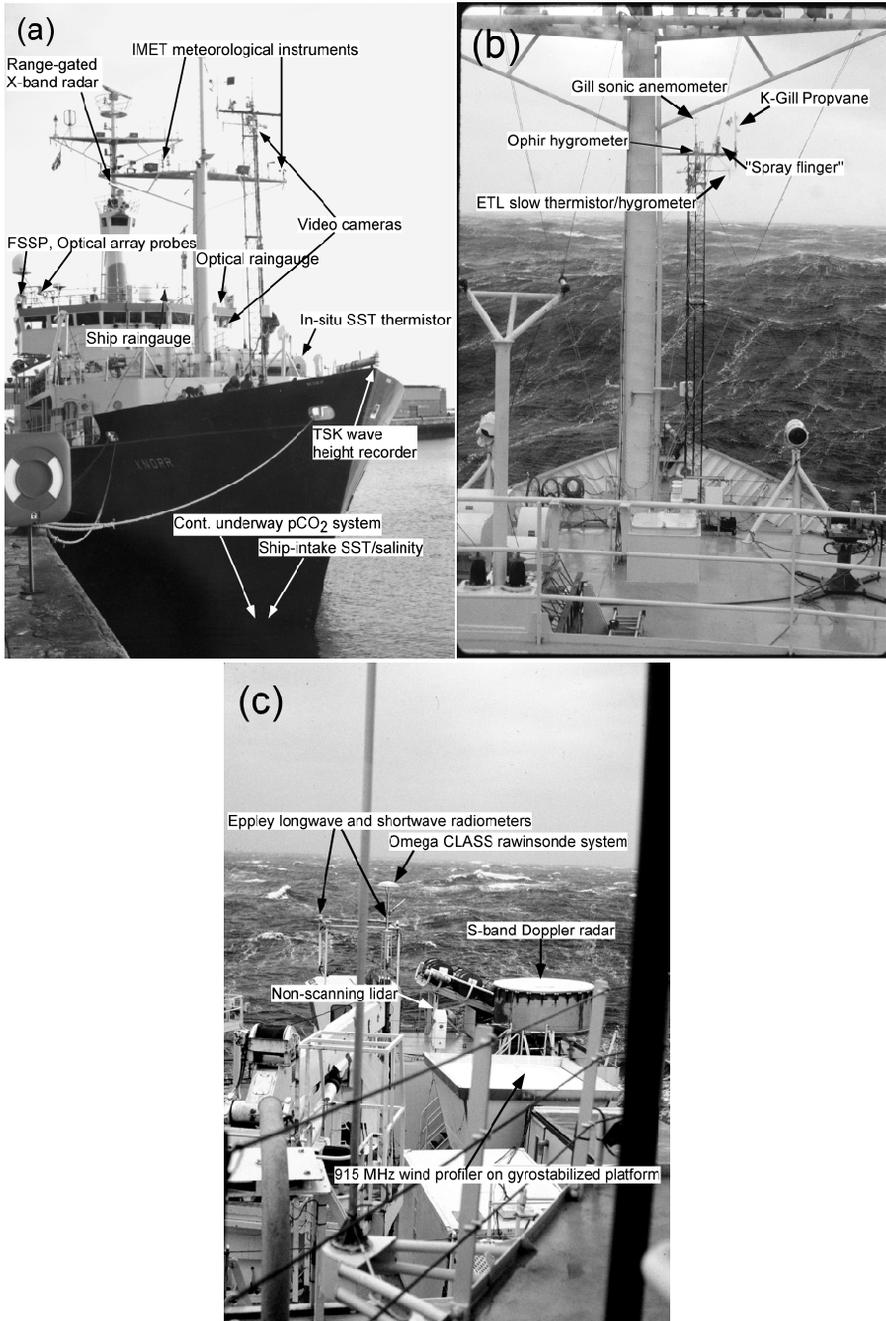


Figure 1. Placement of instrumentation onboard the RV *Knorr* as viewed from: (a) in front of the ship, (b) the bridge toward the bow, and (c) the bridge toward the stern. See text for further details.

Mean sensors included in the ETL suite are: air temperature and humidity (Vaisala) mounted on the bow mast, floating sea-surface temperature (SST) thermistor often referred to as the ‘sea snake’, and IR and solar radiative flux sensors (Eppley PIR and PSP, respectively). The WHOI improved meteorological (IMET) instruments (described at [http://www.whoi.edu/marops/research\\_vessels/knorr/sciequip\\_instrument.html](http://www.whoi.edu/marops/research_vessels/knorr/sciequip_instrument.html)) were mounted on a yardarm of the foremast located in front of the bridge (Fig. 1(a)). The IMET data, along with the ship navigation data, were distributed every 2 s by the onboard Athena data system; they were often redundant but helped fill gaps in the ETL data stream. Additional digital input to the data acquisition system included a gyroscopic compass and a global positioning system (GPS) receiver, deployed to monitor the ship manoeuvres. This integrated flux system enables the estimation of the turbulent fluxes of latent heat, sensible heat, and momentum using three established techniques: the direct eddy correlation (same as covariance), inertial-dissipation (ID), and bulk aerodynamic methods.

Other systems deployed on the RV *Knorr* provided critical information about the ocean surface and boundary-layer environment. An A. T. S. K. Corporation microwave wave-height meter (TSK wave-height recorder; described at <http://www.tsk-jp.com/tska/index.html>) provided by the Bedford Institute of Oceanography was deployed on the bow of the RV *Knorr* to measure wave statistics at a rate of 2 Hz. This instrument included the downward looking Doppler radar with a 13° beam width, which sampled a 1.8 m diameter footprint of the ocean surface from its location at the end of a short bowsprit (Fig. 1(a)), and a gimballed vertical accelerometer mounted below decks at the bow on the ship’s centreline. The ETL floating thermistor and the ship intake provided two *in situ* measurements of SST. In addition, subjective estimates of sea-surface conditions, including wave and swell heights and directions, were provided by the ship’s crew from the bridge at least every 4 hours. Precipitation was measured using a National Center for Atmospheric Research (NCAR) optical rain-gauge and a University of Kiel gimballed ship rain-gauge (Fig. 1(a)). Other *in situ* and remote sensors, not essential for this study but present on the RV *Knorr*, such as the Omega rawinsondes and vertically pointing Doppler S-band radar (White *et al.* 2000), are listed in Table 1. This ship-based observation system provides detailed depictions of the atmospheric boundary, surface layer and the ocean surface, all of which are essential in describing the influence of air–sea interactions on the genesis and development of the FASTEX storms.

### (b) *Data processing*

This study uses data from four main sources: the ETL motion-corrected flux package, the ship’s Athena system, the ETL mean measurement systems, and the TSK wave-height recorder. The first three produce the surface fluxes and the mean values; the final one produces the wave-height and wave-period data. Descriptions of the processing of the surface fluxes, the bulk meteorological data, and the TSK wave-height data follow. The processed datasets are available from <ftp://ftp.etl.noaa.gov/et7/anonymous/cfair-all/fluxdata/fastex/> or the FASTEX data archives (<http://www.cnrm.meteo.fr/dbfastex/>).

(i) *Surface fluxes.* After the velocity signals are corrected for ship motion, data from the first three sources are combined to produce a set of 10-minute processed data files, which include statistics, spectra, and means of all fast-response signals together with the mean meteorological and SST variables. We chose 10-minute segments in order to minimize the effects of weakly non-stationary conditions on the statistics. A final routine applies various corrections, computes the ID and bulk fluxes and the data-quality

TABLE 1. INSTRUMENTS ON THE RV *Knorr* DURING FASTEX

Instrument	Parameters measured	Height (m) a.m.s.l.	Sampling rate	Instrument location
Gill-Solent, R2, ultrasonic anemometer/thermometer	$u, v, w, ws, wd, \tau, H_s, H_1$ , turbulence spectra	19.2	20.83 Hz	bow mast
Ophir IR-2000 fast hygrometer	$q', H_1$	17.5	20 Hz	bow mast
Systron-Donner, Motion-Pak	sonic platform motion	0.5	10 Hz	bow mast
Vaisala HMP35 T/RH probe,	$T, RH$	17.3	10 s	bow mast
ETL floating thermistor	$T_s$	-0.05	(1 minute) 10 s	sea surface
Lowrance Global Positioning System	latitude, longitude, course, speed	N/A	2 Hz	internal laboratory
NCAR optical rain-gauge	precipitation rate	20	10 s	port side railing on
Kiel gimbaled ship raingauge	precipitation rate	20	(1 minute) 10 minutes	bridge roof centre railing on bridge roof
TSK microwave wave-height altimeter	$h_w, t_w$	7	2 Hz	bow
Eppley PSP and PIR	$Q_{si}, Q_{li}$	12	10 s	stern
IMET basic met (RM Young)	$ws, wd, P, T, RH$	20	(1 minute) 15 s	forward mast
IMET ship intake $T_s$	$T_s$	-4	15 s	bow
OMEGA Rawinsondes	$ws, wd, P, T, RH$	2-12 000	1-6 h	fan tail
ETL S-band radar	reflectivity, fall speeds	105-5000	30 s	stern
Non-scanning lidar	aerosol scatter, cloud-base height	30-8000	1 minute	stern
Gyro-stabilized 915 MHz wind profiler	$u, v, w, C_n^2$	500-4000	6 minutes	stern
FSSP, Optical array probes	particle size distribution	20	(30 minutes) 30 minute	starboard railing on bridge roof
Video cameras: forward and downward looking	sea state, visibility			port side of bridge window and bow mast
Gyrocompass	heading			internal laboratory
continuous underway $pCO_2$ system (NOAA/AOML <sup>a</sup> )	surface water and air $pCO_2$	-4 and 13.5	4.5 minutes	bow waterline
K-Gill system with 'spray flinger', dry/wet bulb psychrometers, Lyman-alpha hygrometer	$u, v, w, ws, wd, \tau, H_s, H_1$ , turbulence spectra	K-Gill: 19 spray flinger: 18.4	20 Hz	bow mast
NOAA/ATDD <sup>b</sup> fast hygrometer and $CO_2$ sensor	$q', H_1, CO_2$	18.4	10 Hz	bow mast
Range-gated X-band radar	directional wave height spectra	0	5 s	main mast

The second column shows the directly measured and derived parameters from each instrument. Column three shows the instrument heights for the *in situ* sensors and the sampling heights for the remote sensors. The sampling rate is given in the fourth column, with the minimum stored resolution of the averaged data given in parentheses if different than the sampling rate. The last column and Fig. 1 may be used to locate the instrument on the ship.

<sup>a</sup> Atlantic Oceanographic and Meteorological Laboratories.

<sup>b</sup> Atmospheric Turbulence and Diffusion Division.

$P$  is air pressure,  $RH$  is relative humidity,  $pCO_2$  is the partial pressure of carbon dioxide and  $C_n$  is the refractive index. See text for other details and definitions.

indices, and writes the final file of 10-minute averages. This program also averages the data to fixed 1 h time intervals, and computes the ID and bulk fluxes from 1 h means. The hourly averages provide a more stable estimate of the variances and turbulent fluxes.

Conditional sampling criteria are used to ensure that only high-quality products are included in the subsequent analysis. For each 10-minute period, data are retained only if: (i) the relative wind direction is less than  $\pm 45^\circ$  from directly onto the ship bow; (ii) the standard deviation of the ship's heading is less than  $8^\circ$ ; (iii) the standard

deviation of the ship's speed over ground is less than  $2 \text{ m s}^{-1}$ ; (iv) the standard deviation of the cross-ship motion correction is less than  $2 \text{ m s}^{-1}$ ; and (v) the rainfall rate is less than  $5 \text{ mm h}^{-1}$ . Criterion (iv) ensures the quality of the platform-motion corrections. There are additional requirements on the vertical velocity variance and the cross-stream velocity variance. These criteria are rather weak, allowing some outliers to remain. After conditional sampling, a total of 220 h of high-quality flux data remain within 10-minute segments. Only 10-minute data passing the quality-control checks were used in producing the 1 h averaged turbulence variables (e.g. covariances, variances and structure-function parameters).

The covariance turbulent fluxes are computed by converting the anemometer three-component velocities to fixed earth coordinates, correcting the fast time series for ship motion, and rotating the coordinate system to be aligned along the 10-minute mean flow. The covariance stress ( $\tau_c$ ), sensible-heat flux ( $H_{sc}$ ) and latent-heat flux ( $H_{lc}$ ) are then computed from:

$$\tau_c = \tau_{sc} = \rho_a u_*^2 = -\rho_a \overline{(u'_s w')}, \quad (2.1)$$

$$\tau_{cc} = -\rho_a \overline{(u'_c w')}, \quad (2.2)$$

$$H_{sc} = -\rho_a c_p u_* t_* = c_p \rho_a \overline{T' w'} = c_p \rho_a (\overline{T'_{sn} w'} - 0.51 T \overline{q' w'}), \quad (2.3)$$

$$H_{lc} = -\rho_a L_e u_* q_* = L_e \rho_a \overline{q' w'}, \quad (2.4)$$

where  $u_s$  is the wind component in the mean wind direction (streamwise component),  $u_c$  is the cross-stream wind component,  $w$  is the motion-corrected vertical velocity,  $T$  is the mean air temperature,  $T_{sn}$  is the sonic temperature,  $q$  is the specific humidity,  $\rho_a$  is the air density,  $c_p$  is the heat capacity of air at constant pressure,  $L_e$  is the latent heat of vaporization,  $u_*$  is the friction velocity, and  $t_*$  and  $q_*$  are the scalar flux scales for temperature and humidity, respectively. Primes indicate turbulent departures from the mean. Note that  $\tau_c$  is defined as the streamwise stress,  $\tau_{sc}$ , and does not include a contribution from the cross-stream covariance stress,  $\tau_{cc}$ .

For  $H_{sc}$ , the humidity contribution to the sonic temperature (the second term on the right-hand side of (2.3) was removed using the bulk latent-heat flux ( $H_{lb}$ ) computed from the COARE 3.0 scheme (Fairall *et al.* 1996, 2003) to estimate  $\overline{q' w'}$ . To remove effects of transducer delay, only times were used for which the temperature spectra showed a  $-5/3$  slope at high-frequencies, thereby reducing the amount of sensible-heat flux data compared to the momentum flux data. At times the  $H_{sc}$  showed non-zero values when the air–sea temperature difference was near zero. Comparisons of the covariance spectra to idealized spectra as a function of wind speed for these cases, suggested that a combination of motion-correction errors and flexing of the sonic anemometer produced these unrealistic values. Therefore, a bias correction which increased with increasing wind speed was applied to all the  $H_{sc}$  data presented in this study. This correction was determined from the heat flux offset from zero at near-zero air–sea temperature differences. At a wind speed of  $20 \text{ m s}^{-1}$ , this correction reduced  $H_{sc}$  by about  $9 \text{ W m}^{-2}$ .

For  $H_{lc}$ , the fast humidity fluctuations from the IR hygrometer were scaled by the ratio of the mean hygrometer humidity to the mean IMET humidity (i.e. increased by about 9%) for use in computations of the covariance and the standard deviation of humidity fluctuations. Using Kristensen *et al.* (1997), an additional scaling factor (1.0% increase) for the covariance was used to account for the physical separation of the hygrometer and the sonic anemometer, as the latter was 1.6 m above the former. Because the IR hygrometers detect water-vapour mass concentration, the  $H_{lc}$  are also corrected as per Webb *et al.* (1980).

Averaging the three components of the mast accelerometers for selected time periods during the cruise provides an estimate of the mean tilt of the mast with respect to the gravitational vector. This analysis shows that the mean tilt of the mast is toward the aft by approximately  $2^\circ$ , and to the port side by about  $1^\circ$ . Flow tilt angles, determined from the sonic-measured mean vertical and streamwise velocities averaged with respect to the relative wind direction, varied by from  $7^\circ$  to  $10^\circ$  from the horizontal. Experience has indicated that mean streamwise flow tilts greater than about  $15^\circ$  give questionable fluxes. Furthermore, these observed flow tilt angles are significantly greater than the  $2.3^\circ$  flow tilt angle for bow-on winds obtained from model flow-distortion studies of the RV *Knorr* (Moat and Yelland 1998; Yelland *et al.* 2002), suggesting that the true flow distortion may be greater than the modelled one. Therefore, we have estimated that the mean wind speed at the sonic anemometer location is decreased by about 4.5% rather than the 0.85% estimated by the modelling studies. This increase in airflow deficit of 3.65% may have resulted because the model study was unable to represent the bow mast (referred to as the lattice tower in the modelling studies) and the instrumentation on this mast; this is consistent with the mast and instrumentation effects on the RV *L'Atalante* seen by Weill *et al.* (2003) in computational fluid dynamics and physical simulations.

Furthermore, other studies have suggested that the flow distortion may vary significantly with relative wind direction (e.g. Dupuis *et al.* 2003). In an attempt to quantify this effect for the RV *Knorr*, we chose 21 periods where the non-bow-on winds were interspersed with the bow-on winds, and performed polynomial regressions using only the bow-on ( $\pm 10^\circ$ ) winds in order to obtain 'correct' winds during the times of non-bow-on winds. Though this procedure led to differences with a substantial scatter, the bin mean of these differences showed a nearly linear increase from 0% for bow-on winds to 3% at  $35^\circ$ . That is, the flow accelerated as it moved away from the bow. This 3% acceleration at  $35^\circ$  is about half that reported by Dupuis *et al.* (2003) using a simulation of the RV *L'Atalante*, and is consistent with those estimated from 'star' ship patterns for the RV *Ron Brown* (Webster *et al.* 2002), a vessel with a similar shape to the RV *Knorr*. More than 96% of the 'good' data from the RV *Knorr* used in this paper has relative wind directions within the range  $-12.5^\circ$  to  $+32.5^\circ$ , centred near  $+10^\circ$  (just to the right of the bow when viewing the bow from the ship). The wind speeds are corrected for this relative wind direction effect, retaining the 4.5% mean correction for all directions used. That is, bow-on winds were increased by about 5.5%, while winds with relative directions of  $\pm 35^\circ$  were increased by only 2.5%. The dataset has also been corrected for a 0.7 m upward flow displacement at the bow mast for all relative wind directions, as estimated by Yelland *et al.* (2002). These corrections for wind speed errors and vertical displacement primarily affect the ID fluxes and the mean winds. As the effects of flow-distortion on the turbulent eddies are unknown, no corrections were applied to the covariance statistics.

To compute the ID fluxes ( $\tau_{\text{ID}}$ ,  $H_{\text{SID}}$ ,  $H_{\text{IID}}$ ), the slope of each power spectrum within a reasonable range of inertial-subrange frequencies is compared with the canonical  $-5/3$  power law, and the spectral level within that narrow range of frequencies is used to compute the relevant structure function parameter. It is then straightforward to compute ID estimates of the turbulent scaling parameters for velocity ( $u_*$ ), temperature ( $T_*$ ), and humidity ( $q_*$ ) (see Fairall and Larsen 1986).

We normally use the clear-channel counts from the IR hygrometer as an index of clean optics. Past experience has indicated that the absolute calibration of the IR hygrometer is degraded as the optics become contaminated with salt and/or water. With clean optics the mean clear-channel counts are around 2800–3000 and the standard deviation is between 2 and 15. In experiments with predominantly calm weather we

reject humidity flux values when the mean counts are below 2500 and/or the standard deviation is above 20, although a threshold of 50 often provides acceptable data. FASTEX had predominantly stormy weather, so those criteria reject the majority of the data. To estimate the contamination effects on the fluxes, we carried out a linear regression of  $H_{lc}/H_{lb}$  (or  $H_{IID}/H_{lb}$ ) versus mean clear-channel counts for latent-heat flux. For the covariance flux there is a positive slope, and the ratio is 1.0 for mean counts greater than 2800; for the ID flux the slope is negative, and the ratio is 1.05 for mean counts greater than 2800. Therefore, we have used these slopes to adjust the  $H_{lc}$  and  $H_{IID}$  for mean counts  $<2800$ . At a mean clear-channel count of 2000, this amounts to a 15% increase in  $H_{lc}$  and a 19% decrease in  $H_{IID}$ . While an adjustment of this type violates the usual policy of never adjusting or rejecting direct flux values based on their agreement with the bulk model, we have accepted this compromise because it has essentially no effect on the average of the covariance and ID fluxes (i.e. they are adjusted in opposite directions). Users who find this adjustment overly objectionable can restrict their analysis to data with the standard deviation of clear-channel counts less than 20. An unadjusted version of the dataset is available on request.

(ii) *Bulk meteorological data.* True wind speed is computed from the sonic anemometer using the ship's Laser ring gyro, and the GPS 'speed over ground' and 'course over ground'; thus, it is interpreted as the speed relative to the fixed earth. Air temperature and humidity from the ETL system and the ship's IMET sensor were carefully compared with a hand-held wet- and dry-bulb psychrometer during the experiment. Based on these intercomparisons, the IMET values were deemed more accurate and provide the mean  $T$  and  $q$  values used in this study. At various times during the cruise the sea snake was removed from the water, in which case data from the ship's thermosalinograph were substituted. Thermosalinograph SST measurements were generally of lower quality due to the rough conditions, which caused the intake port to occasionally come out of the water (an example of bad thermosalinograph data can be seen in Fig. 11(c) near JD 8.95).

The long-wave radiative flux from the Eppley PIR unit was logged and computed, as per Fairall *et al.* (1998). The rain rate from the NCAR optical rain-gauge was used, as it appeared to be less affected by airflow over the ship's bridge; however, these data were only available after 2350 UTC 9 January. Prior to that time, values from the Kiel ship gauge are used, though the location of the Kiel gauge (see Fig. 1(a)) apparently led to underestimates of the precipitation amounts by a factor of 3–10 compared to the NCAR gauge. Hence, the editing of the surface fluxes as described above may not have removed all rainfall-contaminated cases before 9 January.

(iii) *TSK wave-height data.* By vertically integrating the Doppler signal every 0.5 s, the TSK microwave wave-height meter provided an objective measurement of the height between the instrument and the ocean surface (see Fig. 1(a)), internally corrected for vertical accelerations. One-hour time series of height are scanned to identify the local troughs and crests (Fig. 2). The height of an individual wave is computed as the mean of the differences between two consecutive troughs and the intervening crest. The measured wave period (encounter period) ( $t_{wm}$ ) is defined as the time between the troughs. To get the true wave period ( $t_w$ ),  $t_{wm}$  must be corrected for horizontal ship speed (speed-over ground,  $SOG$ ) and the relative angle between the ship's course over ground ( $COG$ ) and the waves' phase velocity direction ( $\phi_w$ ) using

$$t_w = 0.5 * [t_{wm} + \{t_{wm}^2 - (8\pi/g)(SOG)(t_{wm}) \cos(\phi_w - COG + 180^\circ)\}^{0.5}], \quad (2.5)$$

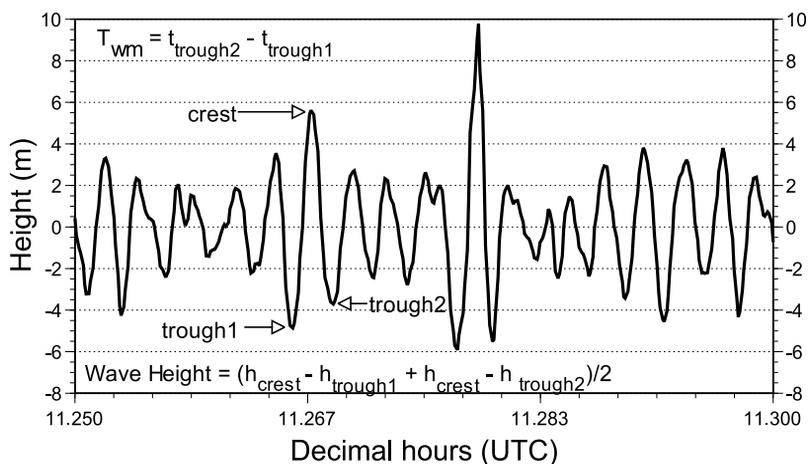


Figure 2. Methodology for calculating wave height,  $h$ , and wave period,  $t_{wm}$ . The solid curve shows the output from the TSK wave-height recorder from 1115–1118 UTC 2 January. See text for discussion.

where  $g$  is the acceleration due to gravity. Equation (2.5) was derived using the basic deep-water relationships between wavelength, wave period and phase velocity. Because  $\phi_w$  are only occasionally available from manual observations, and are therefore poorly known especially for different wave periods or wavelengths, and because much of the flux sampling from the RV *Knorr* was done while the ship was facing into the wind (and hence waves) and moving only very slowly forward,  $t_w$  was only computed for  $SOG < 2 \text{ m s}^{-1}$  and  $\phi_w$  was assumed to be the same as the wind direction. Errors in  $t_w$  are then less than 5% for all wave periods for  $\phi_w$  within  $35^\circ$  of the wind direction, and less than 2% for  $\phi_w$  within  $29^\circ$  of the wind direction and  $t_w > 3.4 \text{ s}$ . The  $\phi_w$  of the longer-period waves are the most likely to differ from the wind direction (e.g. Rieder *et al.* 1994), and these differences are rarely greater than  $60^\circ$  as shown in section 6. For differences of  $60^\circ$ , the maximum error in  $t_w$  for moderate or long periods ( $t_w > 6.4 \text{ s}$ ) is 6.1%.

Waves measured in this manner include both wind waves and swell, and only manually observed wave directions are available. The significant wave height ( $h_{sig}$ ) is computed as four times the standard deviation of the surface-height time series (e.g. Neumann and Pierson 1966), and the significant wave period ( $t_{sig}$ ) is the mean of the periods of the largest one-third waves. Wave-height power spectra were also computed using a fast-Fourier transform. The peaks of these spectra provided another estimate of the dominant wave period. Unpublished comparisons between the ship-mounted TSK wave-height recorder and wave buoys, indicate agreement when significant wave heights are less than 4 m. For larger  $h_{sig}$ , some comparisons indicate that the TSK sensor overestimates the wave heights by 0.5–2 m (F. Dobson, private communication; D. Stredulinsky, private communication) and others report that it underestimates them (J. Edson, private communication). However, these tests were done with the TSK-bearing ship moving at  $5\text{--}8 \text{ m s}^{-1}$  rather than nearly stationary as for the FASTEX sampling. The hour containing the 3-minute sample shown in Fig. 2 had a mean wind speed of  $20 \text{ m s}^{-1}$ , a  $h_{sig}$  of 6.9 m, a  $t_{sig}$  of 8.7 s, and a maximum wave height ( $h_{max}$ ) of 15.5 m. This maximum wave is seen at hour 11.278 in Fig. 2, and is embedded in waves with a large range of heights.

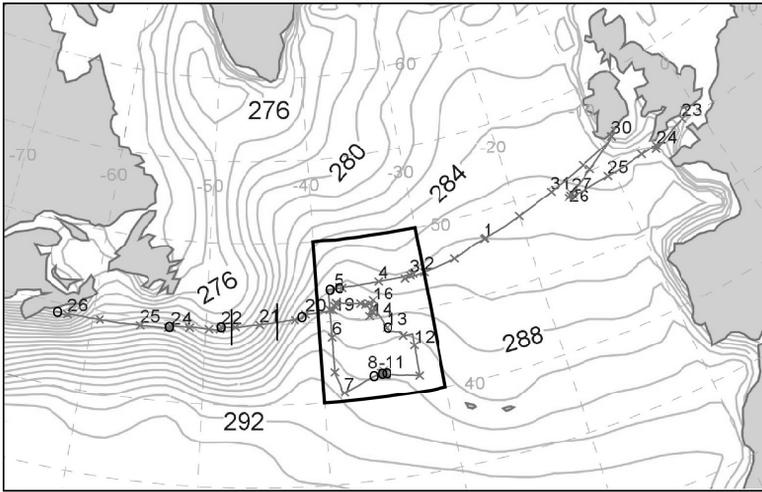


Figure 3. Track of RV *Knorr* (dark line) during the FASTEX experiment from 23 December 1996 to 26 January 1997, marked with an 'x' every 12 h, and labelled with days of the month at each 00 UTC. Also shown is the sea-surface temperature (SST) analysis (grey lines, K) from the European Centre for Medium-Range Forecasts on 7 January 1997. The observed strong SST gradient of the Gulf Stream 'wall' is bracketed by the two vertical lines along the ship track at times Julian day (JD)20.75 and JD21.67. The open circles show the location of the RV *Knorr* at the time of the cold-frontal passages discussed in sections 4–6 and listed in Table 2.

### 3. ENVIRONMENTAL DESCRIPTION

#### (a) Cruise track, sampling strategy and general environment

The collection of the data used in this study began on 23 December 1996 south-west of the UK, and continued until 27 January 1997 when the RV *Knorr* entered the port of Halifax, Nova Scotia (Fig. 3). The data 'shake-down' portion of the cruise during 22–27 December took place in the southern end of the English Channel and south-west of Ireland. Sampling started on the main portion of the cruise on 30 December after leaving Cork, Ireland. Before 20 January, the RV *Knorr* was in the Gulf Stream waters south and east of the strong SST gradient ('SST wall') to the north-west. The area defined by (40–50°N, 30–40°W) in the central Atlantic was called the FASTEX NUS domain. After 20 January, the RV *Knorr* headed toward Halifax west of the Gulf Stream approximately along 45°N latitude, stopping to sample storms as they occurred. The SST gradient at the SST wall was actually about twice as strong as that depicted by the ECMWF analysis shown in Fig. 3, since the SST measured by the RV *Knorr* on 20–21 January changed from 287.4 to 275.8 K from longitude 43.9°W to 48.5°W, while Fig. 3 indicates a SST change from 283.5 to 278.5 K for this same track. Hence, near 45°N, the Gulf Stream SST wall, bracketed by the vertical lines, is actually slightly farther west than suggested by the ECMWF analysis, and is better represented by the detailed fields given by Eymard *et al.* (1999).

Using weather forecasting material provided by the UK Met Office and the FASTEX Operations Centre, the ship was positioned to sample the strongest winds available for a given storm, and to maintain an approximate north–south line with the other FASTEX ships between 9–20 January. Generally, the RV *Knorr* was in the middle of the line. In order to maximize the amount of good-quality high wind speed data collected, an effort was made to only move the ship during periods of weak winds and high pressure. During the periods of stronger winds, the ship was turned into the wind

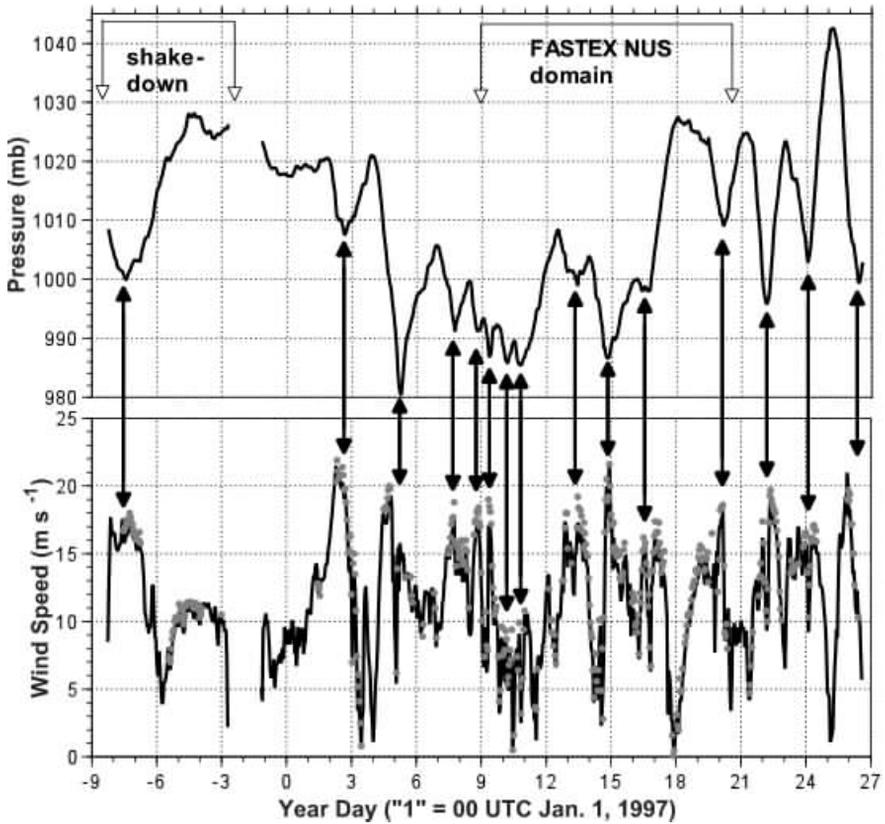


Figure 4. Surface pressure and 19 m wind speed from the RV *Knorr* during its FASTEX cruise. The wind speeds from both the Athena system (solid) and the NOAA/ETL sonic anemometer (dots) are shown. The double-headed arrows show the times of each storm or frontal wave. See text for details.

(and usually the waves), and measurements were made while the ship moved forward at  $1\text{--}2\text{ m s}^{-1}$  relative to the water. This strategy was used throughout the cruise, even during the portion after leaving the NUS and heading toward Halifax.

Of the 15 storms and frontal waves sampled by the RV *Knorr*, indicated by their pressure troughs and (usually) associated wind speed peaks in Fig. 4, seven were FASTEX intensive observation periods, while the others occurred either before the field program started on 9 January, or were storms forecast not to enter the FASTEX mesoscale sampling area near Ireland 18–48 h downwind of the RV *Knorr*. During the shake-down portion of the cruise, post-frontal high pressure dominated with moderate-to-strong winds from the European continent.

#### (b) Atmospheric surface layer

Over 150 hours of flux data in wind speeds greater than  $15\text{ m s}^{-1}$  (at a height of 19 m) were collected (Fig. 5), representing a significant increase in the direct covariance flux data collected under high wind speed, open-ocean conditions. Measurements were obtained in various stability regimes: from within the cold air ahead of highly occluded fronts, below the low-level jet in the warm sectors, and in the cold air behind cold fronts. In general, east of the Gulf Stream wall the sampled data were unstable with the air cooler than the sea surface; a few cases of weakly stable conditions were sampled.

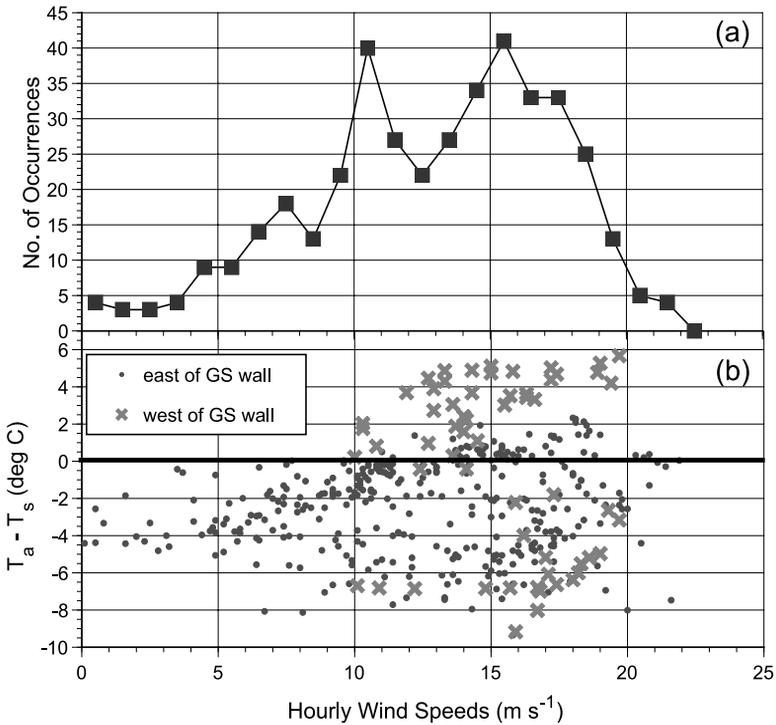


Figure 5. Characterization of the data sampled from the RV *Knorr* during its FASTEX cruise showing: (a) the number of occurrences of each hourly  $1 \text{ m s}^{-1}$  wind speed bin, and (b) the air-sea temperature difference associated with each hourly wind speed measurement. Data for 344 h were collected to the east of the Gulf Stream ‘wall’ (see text; dots in (b)), and for 59 h to the west of it (x in (b)). Only points for which the ship was stationary and facing into the wind with the sonic anemometer operational are shown.

Over the cold waters west of the Gulf Stream wall the distribution was bimodal, with warm-sector regimes providing mainly stable conditions and post-cold-frontal regimes providing unstable environments. Significant numbers of hours with wind speeds between  $15\text{--}22 \text{ m s}^{-1}$  were sampled in all of these stability conditions. For hours for which  $\tau_c$ ,  $H_{sc}$ , and  $H_{lc}$  were all available, the stability parameter  $z/L$  as a function of wind speed can be calculated, where  $z$  is the wind measurement height ( $19.2 \text{ m}$ ), and  $L$  is the Obukhov length defined by:

$$L = (Tu_*^2) / \{\kappa g(t_* + 0.61Tq_*)\}, \quad (3.1)$$

where  $\kappa$  is the von Kármán constant (0.4). Plotting  $z/L$  as a function of wind speed shows that the environments for most wind speeds  $> 10 \text{ m s}^{-1}$  were weakly unstable ( $-0.1 < z/L < 0$ ) east of the Gulf Stream wall, while both weakly unstable and stable conditions occurred to the west of it (Fig. 6).

In general, the surface streamwise stress increases rapidly with wind speed, with a modest scatter around this curve (Fig. 7(a)). Hourly average stresses as great as  $1.5 \text{ N m}^{-2}$  were observed. Many of the surface-flux parametrization schemes, including COARE 3.0, were developed from data with wind speeds below  $12 \text{ m s}^{-1}$ . This corresponds to a stress of about  $0.3 \text{ N m}^{-2}$ , or only 20% of the range observed in the FASTEX dataset. During FASTEX, a slight positive average cross-stream stress was observed for all wind speeds (Fig. 7(b)), implying a mean stress from the right when facing upwind. This component averages less than  $0.1 \text{ N m}^{-2}$ , though individual

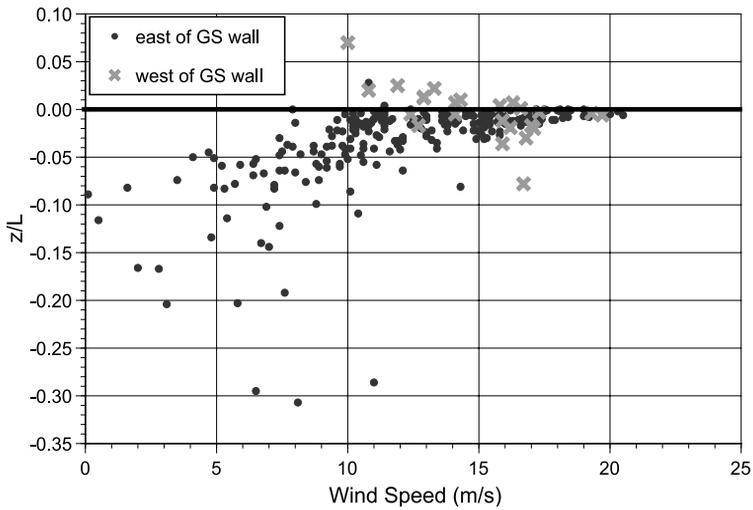


Figure 6. Scatter plot of stability parameter  $z/L$ , from the RV *Knorr* during its FASTEX cruise, where  $z$  is height and  $L$  is Obukhov length (see Eq. (3.1)), versus wind speed for the hourly samples from the RV *Knorr* for which the stress, sensible-heat, and latent-heat fluxes were all available, discriminating between observations east and west of the Gulf Stream ‘wall’ (see text). There are 289 points.

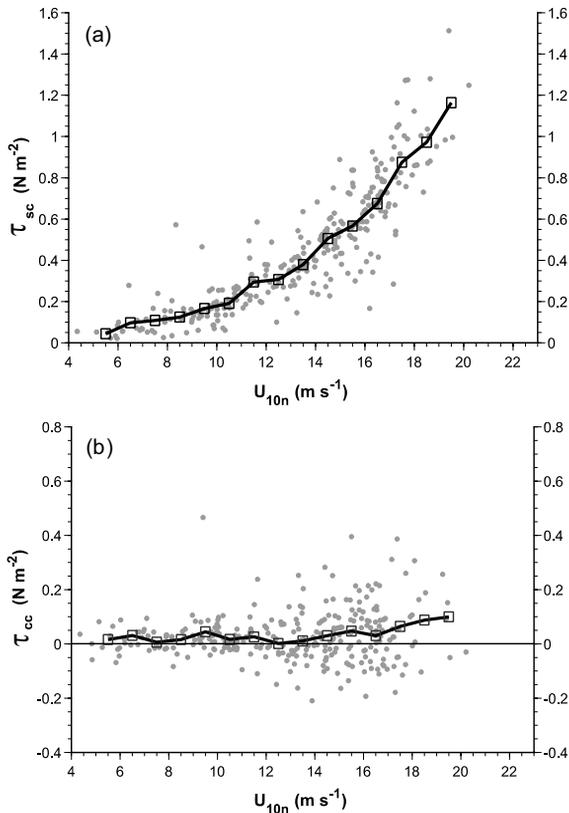


Figure 7. Scatter plots of hourly averaged values from the RV *Knorr* during its FASTEX cruise of: (a) streamwise covariance stress ( $\tau_{sc}$ ), and (b) cross-stream covariance stress ( $\tau_{cc}$ ) as functions of the 10 m neutral wind speed ( $U_{10n}$ ). Also shown are bin-averaged values for  $1 \text{ m s}^{-1}$  wind speed bins (squares).

points up to  $0.5 \text{ N m}^{-2}$  occur for wind speeds greater than  $8 \text{ m s}^{-1}$ , while the negative excursions are smaller but also tend to occur for the moderate to strong winds. This average positive cross-stream stress is believed to be physical, and has significance when relating the surface stress to the synoptic environment, as will be discussed later.

Since the stability is near neutral in these high wind speed conditions, the errors in both the streamwise and cross-stream directions due to anemometer tilt errors should be fairly small. With observed boundary-layer depths of 500–1500 m (not shown) and a mast tilt in the along-ship (approximately streamwise) direction of  $2^\circ$  (see subsection 2(b)(ii)) and for stability of  $-0.2 < z/L < 0$ , the results of Wilczak *et al.* (2001, their Fig. 2) indicate that the error in the streamwise stress should be approximately 12–16%. The  $1^\circ$  mast tilt in the cross-ship (approximately cross-stream) direction suggests a cross-stream stress error of 6–8%.

The 10 m neutral drag coefficients ( $C_{Dn10}$ ), sensible-heat transfer coefficients ( $C_{hn10}$ ), and moisture transfer coefficients ( $C_{en10}$ ) are computed from both the covariance and inertial dissipation fluxes using the methods described by Fairall *et al.* (2003). As discussed by Fairall *et al.* (2003), for a neutral 10 m wind ( $U_{10n}$ )  $> 12 \text{ m s}^{-1}$ , the  $C_{Dn10}$  for the covariance technique is greater than that for the ID technique (Fig. 8(a)), and there is uncertainty as to which is correct.

The FASTEX  $C_{Dn10}$  values from the ID technique are in good agreement with those from the ID technique of numerous other studies. The shaded area in Fig. 8(a) encompasses the ID curves from Large and Pond (1982), Anderson (1993), Yelland *et al.* (1998) and Dupuis *et al.* (2003). It also encompasses the curves obtained from the buoy-based covariance measurements of Smith (1980), Large and Pond (1982) and Dupuis *et al.* (2003), the first two studies using the Bedford Institute buoy 10 km off the coast of Nova Scotia and the last study using the ASIS (Air–Sea Interaction Spar) buoy 50 km from shore in the small, semi-circular Gulf of Lyon. Although only the long-fetch wind directions were used for these curves, it is unclear whether these coastal-zone measurements truly represent open-ocean conditions, because the low-level airflow together with the ocean wave conditions, and thereby the surface stresses, are possibly affected by mesoscale atmospheric phenomena and wave reflections attributable to their coastal location. Only the measurements of Eymard *et al.* (1999) and Dobson *et al.* (1994) produced ID curves similar to the FASTEX covariance curves, though these ID estimates were not corrected for flow distortion, while nearly all those within the shaded area were. The flow-distortion corrections will generally reduce the  $C_{Dn10}$  for a given  $U_{10n}$ , as shown by Dupuis *et al.* (2003).

Since this FASTEX dataset represents the only ship-based covariance flux measurements, and possibly the only covariance flux measurements, in an open-ocean storm environment regardless of platform (depending on the interpretation of the representativity of the coastal buoy-based covariance flux measurements), it is difficult to ascribe a reason for the difference between the FASTEX covariance and ID  $C_{Dn10}$  values. It is possible that flow-distortion effects on the turbulent eddies have led to this difference, as the effect of flow distortion on the covariance flux measurements appears to be greater than for the ID technique (e.g. Edson *et al.* 1991), but the specific effects on turbulent eddies are unknown, as are the specific effects on the turbulence at the sonic anemometer location of the RV *Knorr*. However, the ID assumptions of an empirical imbalance term, a varying Kolmogorov ‘constant’, and a zero pressure-term effect on the kinetic-energy balance lead to errors, especially at higher wind speeds. Janssen (1999) showed that the perturbation-pressure term in the kinetic-energy balance became significant at higher wind speeds and larger swell, thereby greatly increasing the  $C_{Dn10}$  estimates from the ID technique. Curve J99 in Fig. 8(a) shows that Janssen’s pressure-term correction of the

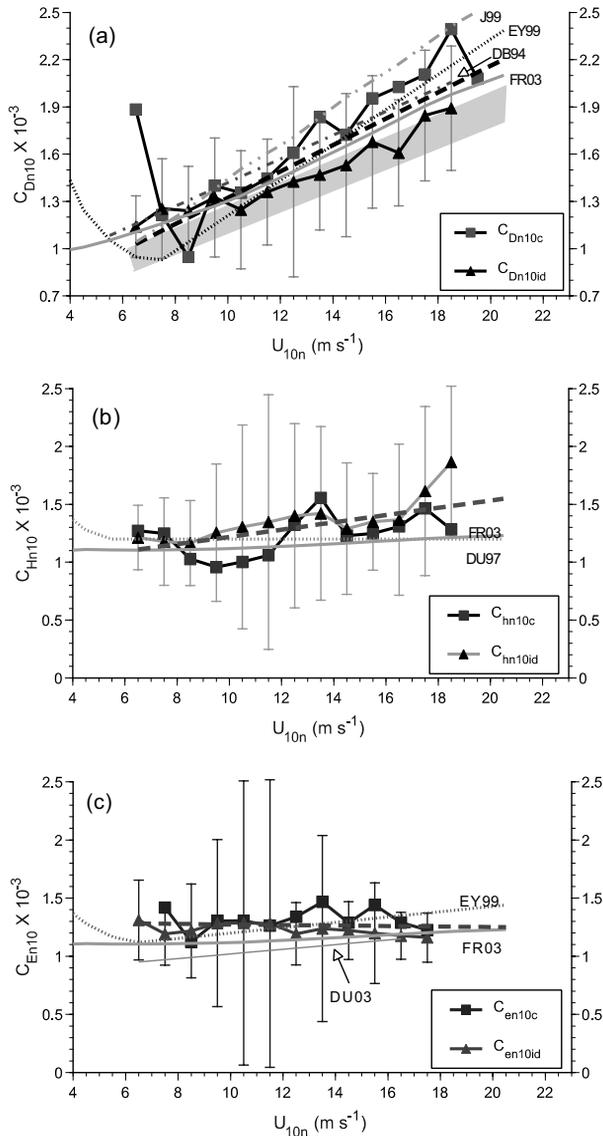


Figure 8. The bin-median values of the 10 m neutral transfer coefficients from the RV *Knorr* during its FASTEX cruise, for: (a) momentum, (b) sensible heat, and (c) latent heat, as functions of the 10 m neutral wind speed ( $U_{10n}$ ) from the covariance (squares) and inertial dissipation (ID) (triangles) techniques. The linear fits to the average of the two methods are shown as heavy dashed lines. Only bins with at least 18 ten-minute data points are shown, and the error bars show one standard deviation about the ID values. Additional curves are from Fairall *et al.* (2003; FR03, grey solid), Eymard *et al.* (1999; EY99, dotted in (a) and (c)), Dupuis *et al.* (1997; DU97, dotted in (b)), Janssen (1999; J99, dash-double dotted in (a)), Dupuis *et al.* (2003; DU03, thin solid in (c)), and Dobson *et al.* (1994; DP94, dash-dotted in (a)). The shaded area in (a) represents the region of numerous other curves, primarily from ID measurements as described in the text.

ID results from Yelland and Taylor (1996) produces  $C_{Dn10}$  values that are slightly greater than the covariance values measured during FASTEX. Before correction, the ID results from Yelland and Taylor (1996) were located within the shaded area in Fig. 8(a). Hence, while the agreement between the  $C_{Dn10}$  from the flow-distortion corrected ID technique and the buoy-based covariance measurements suggests that the FASTEX ship-based covariance fluxes may be too large, possibly due to unknown effects of flow distortion, the assumptions in the ID technique at high winds as represented by the Janssen (1999) study suggest that the FASTEX covariance fluxes may be closer to the truth than the ID fluxes; this implies that the buoy-based covariance flux measurements may have been affected by coastal effects. As a further complication, the Janssen (1999) results are also a contentious issue and a source of scientific debate (Janssen 2001; Taylor and Yelland 2001a).

Until the above issues are resolved, we feel that the current best estimate of the  $C_{Dn10}$  from the FASTEX data is represented by the average of the two methods. The linear fits to the average curves are represented by the heavy dashed lines in Figs. 8(a) to (c) and are given by:

$$C_{Dn10} = (0.0768U_{10n} + 0.603) \times 10^{-3} \quad 6 < U_{10n} < 19 \text{ m s}^{-1}, r^2 = 0.92, \quad (3.2a)$$

$$C_{hn10} = (0.0298U_{10n} + 0.922) \times 10^{-3} \quad 6 < U_{10n} < 19 \text{ m s}^{-1}, r^2 = 0.58, \quad (3.2b)$$

$$C_{en10} = (-0.0011U_{10n} + 1.297) \times 10^{-3} \quad 6 < U_{10n} < 18 \text{ m s}^{-1}, r^2 = 0.01, \quad (3.2c)$$

where  $r^2$  is the correlation coefficient of the regression to the bin averages.

The surface flux scheme of Fairall *et al.* (2003) predicts  $C_{Dn10}$  between the two observed FASTEX curves close to the average. The RV *Knorr* data suggest a slight increase in  $C_{hn10}$  with  $U_{10n}$ , though the scatter around the mean values gives us only low confidence in this trend. Note that the  $H_{sc}$  used in the calculation of the  $C_{hn10}$  are corrected for a bias as described in subsection 2(b)(ii). If uncorrected  $H_{sc}$  were used, the slope of (3.2b) is only slightly larger. The  $C_{en10}$  from the RV *Knorr* is essentially constant near  $1.3 \times 10^{-3}$ , slightly higher than that predicted by the model of Fairall *et al.* (2003).

### (c) Ocean surface characteristics

The atmospheric surface layers sampled by the RV *Knorr* during FASTEX were either over the relatively warm waters of the mid-Atlantic Gulf Stream east of the Gulf Stream wall, or in the much colder waters of the Labrador current west of the Gulf Stream. In the former area the observed SSTs were 11–18 °C, while in the latter they were 2–6 °C, producing different stability distributions in each region (see Figs. 3, 5 and 6).

During the cruise, the significant wave height was approximately constant at 4 m for wind speeds less than 10 m s<sup>-1</sup>, and generally increased with increasing wind speed for stronger winds, reaching values of about 6.5 m for a 22 m s<sup>-1</sup> wind (Fig. 9(a)). However, there is a large scatter around this mean curve. The 4 m wave heights for the lower wind speeds represent the constant swell conditions in the mid-Atlantic Ocean. The maximum wave heights averaged nearly 12 m for winds near 20 m s<sup>-1</sup>, but they are widely scattered around the bin-averaged curve, including one of 14.5 m for a 10 m s<sup>-1</sup> wind speed. A maximum wave height of 15.7 m occurred for a wind of 19.3 m s<sup>-1</sup> on 4 January. One nearly as high (15.5 m) is shown in Fig. 2. Neither of these is included in Fig. 9(a), as covariance stresses were not obtained for these hours. The mean period of the significant waves was approximately 8.2–8.7 s for weak winds, it had a minimum

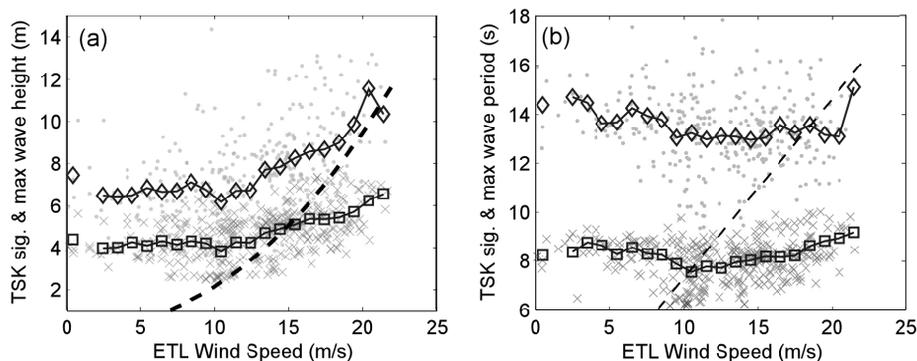


Figure 9. Scatter plots using data from the TSK wave-height recorder on RV *Knorr* during its FASTEX cruise of: (a) significant (x) and maximum (dots) wave heights, and (b) significant (x) and maximum (dots) wave periods, as functions of wind speed. Bin averaged values for  $1 \text{ m s}^{-1}$  wind speed bins are squares and diamonds, respectively. The wave heights and wave periods in equilibrium with the given wind speed are shown as bold dashed curves (from Taylor and Yelland 2001b). Only hours for which covariance stresses were obtained are shown. See text for further details.

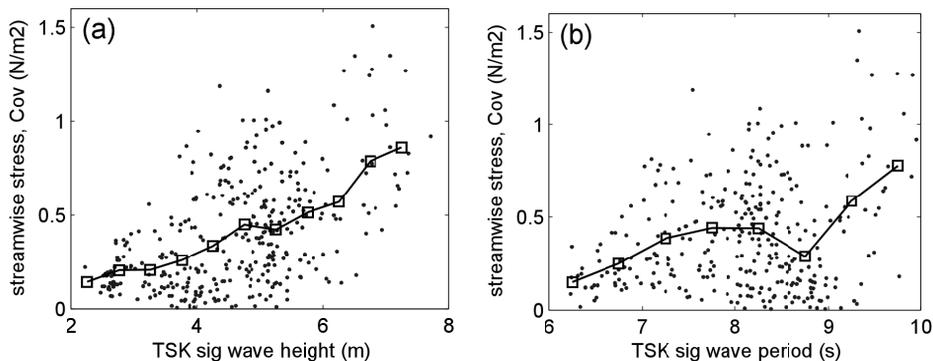


Figure 10. Scatter plots using data from the RV *Knorr* during its FASTEX cruise (see text) of hourly values of streamwise stress as a function of: (a) significant wave height, and (b) significant wave period. Bin-averaged values are shown as squares.

of about  $7.5 \text{ s}$  for  $10\text{--}11 \text{ m s}^{-1}$  winds, and slowly increased to  $9.2 \text{ s}$  for  $22 \text{ m s}^{-1}$  winds (Fig. 9(b)). Estimates of the significant wave heights and periods of waves in equilibrium with the local winds, show that the waves were rarely in equilibrium with the local winds. Hence, equilibrium relationships are not useful for describing the wave characteristics in the FASTEX environment. In particular, the equilibrium characteristics underestimate the significant wave heights in weak winds since there is nearly always a significant swell. For the strongest winds, the equilibrium characteristics are closer to the characteristics of the tallest waves rather than the significant waves, suggesting that the atmospheric conditions producing the strong winds are too transient to allow the local waves to reach equilibrium. The equilibrium wave conditions were calculated using the relationships given in the appendix of Taylor and Yelland (2001b).

The relationship between the wave characteristics and the measured surface stress is not strong in the data. There is an indication that the surface stress increases with significant wave height (Fig. 10(a)), though the scatter of the hourly values around this curve is significantly greater than the scatter around the curve for the stress as a

function of wind speed in Fig. 7(a). The relationship between the mean period of the significant waves and the surface stress is even more complex (Fig. 10(b)). In addition, there appears to be no obvious relationship between the wave characteristics and the cross-stream stress (not shown).

#### 4. COMPOSITING METHOD

In this and the next two sections, we show that the surface-layer and ocean characteristics that showed a large scatter in the traditional presentations in section 3, have clear relationships to the synoptic environment. In order to place the observations in a storm-relative framework, the beginning of the warm sector (WSB), the surface cold-frontal passage (CFP), and the end of the post-frontal baroclinic regime (PFE) were defined from basic meteorological parameters. The near-surface specific humidity was used as a key parameter to define the WSB and the PFE. The surface increase of specific humidity at the WSB, and the disappearance of the humidity perturbation at the end of the PFE were clear in every case (e.g. Fig. 11(a)). The CFP was defined by the surface wind direction shift (Fig. 11(a)). Therefore, the warm-sector region is defined as the time between the increase in the surface specific humidity and the wind-direction shift, while the post-frontal baroclinic zone regime is defined as the period between the wind-direction shift and the end of the decrease in specific humidity. Typical changes in air temperature, wind speed, surface pressure and precipitation were often seen (e.g. Figs. 11(b), (c), and (d)), though these were not used to define the transitions. Note that in Fig. 11 the time increases from right to left along the abscissa, in order that the warm sector and post-frontal air have the same spatial relationship to the cold front as when observing a storm system from a satellite image (e.g. Fig. 12).

With these definitions, statistical composites of storm-relative atmospheric parameters, surface fluxes, and wave characteristics were computed for the RV *Knorr*. The composites were temporally normalized using the duration of the warm-sector region for each case. Again, the abscissas of the composites are defined so the normalized time  $\tilde{t}$  increases with real time, and the warm sector is to the right and post-frontal air to the left of the cold front. Hence the warm sector occurs for a  $\tilde{t}$  from  $-1$  to  $0$ , and the post-frontal region occurs for  $\tilde{t}$  from  $0$  to  $1$  (though the post-frontal region in most cases extended only from  $0$  to  $\sim 0.5$ ). The difference in duration between the post-frontal region and the warm sector led to fewer samples during the latter half of the normalized post-frontal region.

Mathematically, the compositing can be expressed as:

$$\tilde{\lambda}'_k(i) = \langle \lambda_j(i) - \lambda_{\text{wsb}}(i) \rangle, \quad (4.1)$$

$$\tilde{wd}'_k(i) = \langle wd_j(i) - wd_{\text{cfp-1h}}(i) \rangle, \quad (4.2)$$

$$\tilde{\alpha}_k(i) = \langle \alpha_j(i) \rangle, \quad (4.3)$$

where  $\lambda$  represents air temperature ( $T_a$ ), specific humidity ( $q_a$ ) and wind speed ( $ws$ );  $wd$  is the wind direction;  $\alpha$  represents all other composited variables such as  $\tau_c$ ,  $\tau_{\text{ID}}$ ,  $\tau_b$ ,  $H_{\text{sc}}$ ,  $H_{\text{SID}}$ ,  $H_{\text{sb}}$ ,  $H_{\text{lc}}$ ,  $H_{\text{IID}}$ ,  $H_{\text{lb}}$ ,  $h_{\text{sig}}$ ,  $h_{\text{max}}$ . Angled brackets  $\langle \rangle$  indicate averaging over the observation times  $t_j$  that produce normalized times,  $\tilde{t}_j$ , within the  $k$ th normalized time interval of length 0.1 through:

$$\tilde{t}_j(i) = \{t_j(i) - t_{\text{cfp}}(i)\} / \{t_{\text{cfp}}(i) - t_{\text{wsb}}(i)\}, \quad (4.4)$$

where  $i$  is the index for each frontal case, and the primes represent storm-perturbation quantities. For each normalized time interval  $k$ , the mean of each variable over the

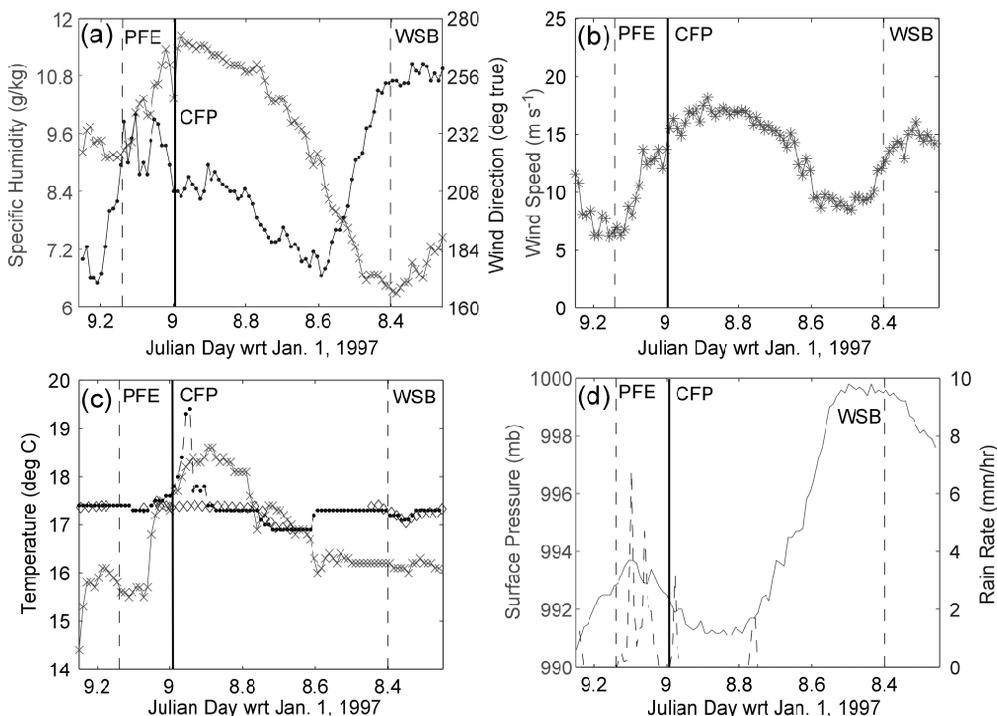


Figure 11. Time series from the RV *Knorr* during its FASTEX cruise (see text) for case 4 (Julian day 8.25–9.25) of: (a) specific humidity (x) and wind direction (dots); (b) 19 m wind speed; (c) air temperature (x), and sea-surface temperatures from both the thermosalinograph (dots) and the sea snake (diamonds); and (d) surface pressure (solid) and rain rate (dashed). The onset of the warm sector (WSB), the cold-frontal passage (CFP), and the end of the post-frontal regime (PFE) are marked by dashed and solid vertical lines. Note that time runs from right to left in order to place the warm sector to the right of the cold front.

ten cases (for  $i = 1 : 10$ ) are computed. For various reasons, some  $\tilde{t}_k$  for some variables have fewer than ten cases. The temporal normalization (4.4) removes biases resulting from the physical size and translation speed of the storm. By first obtaining average values for each normalized time and each storm, and then averaging the ten storms, equal weight is given to each synoptic case regardless of the number of measurements for each case.

The storms' movements past the RV *Knorr* and the other FASTEX research vessels produced a north-east to south-west time-to-space adjusted 'track' for each ship through each storm (Fig. 12). The orientation and path of the storm determined the obliqueness of the ships' tracks relative to the frontal cloud bands. For example, the tracks were nearly orthogonal to the surface cold front in cases 3 and 6, while they were nearly parallel for cases 4 and 8. Since the RV *Knorr* had to pass through the warm sector of an open frontal wave in order that the data be useable in this compositing method, only ten out of the 15 cases shown in Fig. 4 were used, and these are listed in Table 2. The duration of the warm sectors averaged 17.7 h, ranging from 3.1–43.4 h. The post-frontal region was less than half that in duration. Because  $\tilde{t}$  is defined using the duration of the warm-sector region, most cases have no data beyond  $\tilde{t} = 0.5$ . The first seven cases were obtained south and east of the Gulf Stream SST wall, while the last three cases were obtained in the colder waters to the north and west.

While normalizing using the duration of the warm sector is appropriate for studying the warm sector, a normalization of the post-frontal region is probably better carried out

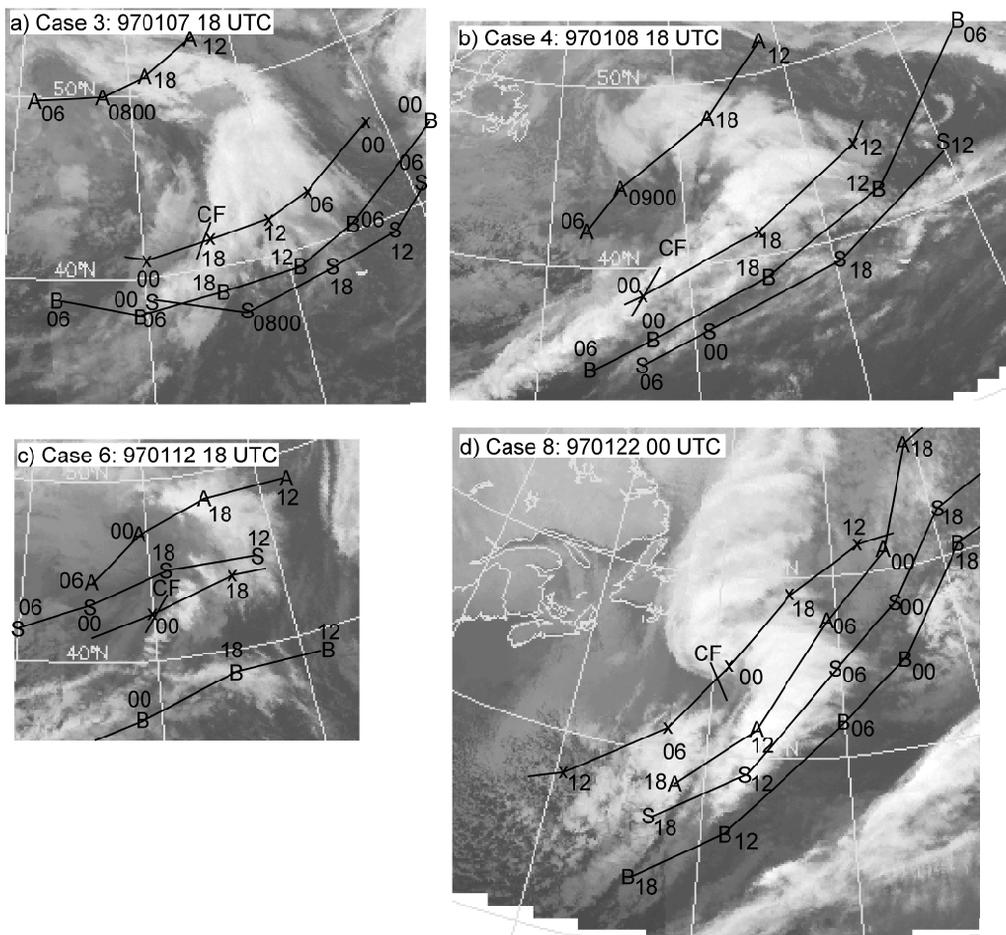


Figure 12. Time-to-space converted tracks during FASTEX of the research vessels (RVs) *Knorr* (x), *Aegir* (A), *Victor Bugaev* (B), and *Suroit* (S) for cases 3, 4, 6 and 8. The overlaid infrared satellite image corresponds to a time shortly before RV *Knorr* passed through the surface cold front. The time next to each symbol marks the hour (UTC) at that location. See text for details.

TABLE 2. CHARACTERISTICS OF FASTEX CASES USED FOR COMPOSITING THE RV *KNORR* DATA

Case number	Warm-sector duration (h)	Cold-frontal passage (decimal Julian day)	Post-frontal baroclinic regime duration (h)	System phase velocity ( $\text{m s}^{-1}/\text{deg}$ )	15-minute, 19 m LLJ <sup>a</sup> wind speed max ( $\text{m s}^{-1}$ )	Average sea-surface temperature ( $^{\circ}\text{C}$ )
1	16.44	4.885	3.48	18.4/210	21.5	15.0
2	3.12	5.190	15.84	11.6/225	19.1	15.4
3	18.62	7.776	6.58	18.7/254	19.6	17.4
4	14.28	8.995	3.48	33.4/240	19.0	17.4
5	3.50	9.425	2.76	26.4/233	21.3	17.3
6	7.92	12.99	4.80	25.0/258	18.6	15.4
7	43.44	20.21	16.08	23.2/229	20.0	14.5
8	15.36	22.05	13.20	27.4/234	18.3	3.4
9	28.32	24.08	12.48	25.7/266	18.0	4.0
10	25.68	26.37	5.52	27.3/234	22.0	3.2
Average	17.7	N/A	8.4	23.7/238	19.7	16.1 and 3.5 <sup>b</sup>

<sup>a</sup>Low-level jet.

<sup>b</sup>The two averages for the last column are the average sea-surface temperatures for the first seven and last three cases, respectively.

using the duration of the post-frontal period, though this would lead to a change in time-scales at the cold front. Though not shown, such a scaling has also been carried out; the results are similar to those to be shown using the warm-sector time-scaling for the entire storm period, except that data extend to  $\tilde{t} = 1$ .

Synoptic storm systems provide the principal source of variability in low-level, extratropical, atmospheric structure. Furthermore, numerous studies show repeatable mesoscale structure at low levels in the warm sector and near the cold fronts of oceanic midlatitude storms, such as a plume of greater specific humidity, higher temperatures, and a low-level jet (LLJ) (e.g. Browning and Pardoe 1973; Hobbs *et al.* 1980; Wernli 1997; Ralph *et al.* 2004). Since these features are generally dynamically linked, we expect that the composites will show systematic changes in the principal atmospheric parameters. Because, in a bulk sense, the surface fluxes are related to the principal atmospheric parameters, we also expect that the flux composites will show systematic changes in the surface-layer fluxes relative to the storm systems. Departures from this systematic behaviour of the fluxes will indicate the effects of other processes, such as changes in the surface wave characteristics and SST.

## 5. COMPOSITE SURFACE LAYER

### (a) Atmospheric

The composites of the basic surface-layer parameters of air temperature, specific humidity, wind speed, and wind direction show regular variations relative to the location of the surface front (Fig. 13). The air temperature shows an increase of about 4 degC from the WSB ( $\tilde{t} = -1$ ) to just before the CFP ( $\tilde{t} = 0$ ), decreasing behind the cold front. The specific humidity increases by nearly 4 g kg<sup>-1</sup> within the warm sector, peaking just before the CFP. The wind speed shows the surface-layer manifestation of the classical LLJ shortly before the frontal passage, with an increase of about 8.5 m s<sup>-1</sup> from the WSB. The maximum composite wind speed (not shown) was about 17 m s<sup>-1</sup> in the warm sector, though the maximum 15-minute 19 m wind speeds during the warm sector averaged 19.7 m s<sup>-1</sup>, and ranged from 18–22 m s<sup>-1</sup> (Table 2). The wind speed initially drops at the CFP, but then increases to another peak in the post-frontal region. An examination of associated satellite images showed no obvious post-frontal circulation or frontal features, suggesting that this post-frontal peak is most likely due to enhanced vertical mixing because of the weaker post-frontal stability. In the transition from the eastern edge of the warm sector, the wind direction initially has a more westerly component than at the CFP, and then a slightly more easterly component. This implies that the surface-layer flow is diffluent at the eastern edge of the warm sector, and then becomes slightly confluent from near the middle of the warm sector to just before the CFP. Note that the onset of the wind speed increase occurs west of the eastern edge of the warm sector (near  $\tilde{t} = -0.9$ ) as defined by the specific humidity. Hence, the thermodynamic and kinematic definitions of the warm-sector region are not exactly coincident. The main storm-generated variations are qualitatively present in all cases, as indicated by hourly standard deviations that are smaller than the amplitudes of the storm-generated variations.

Though the wind direction changes by 100° after the CFP, the ship-relative wind direction (not shown) changes only by about 5°, since the orientation of the ship was adjusted as the wind shifted. This minimizes directional differences of flow-distortion effects when examining the flux changes between the warm sector and post-frontal sector.

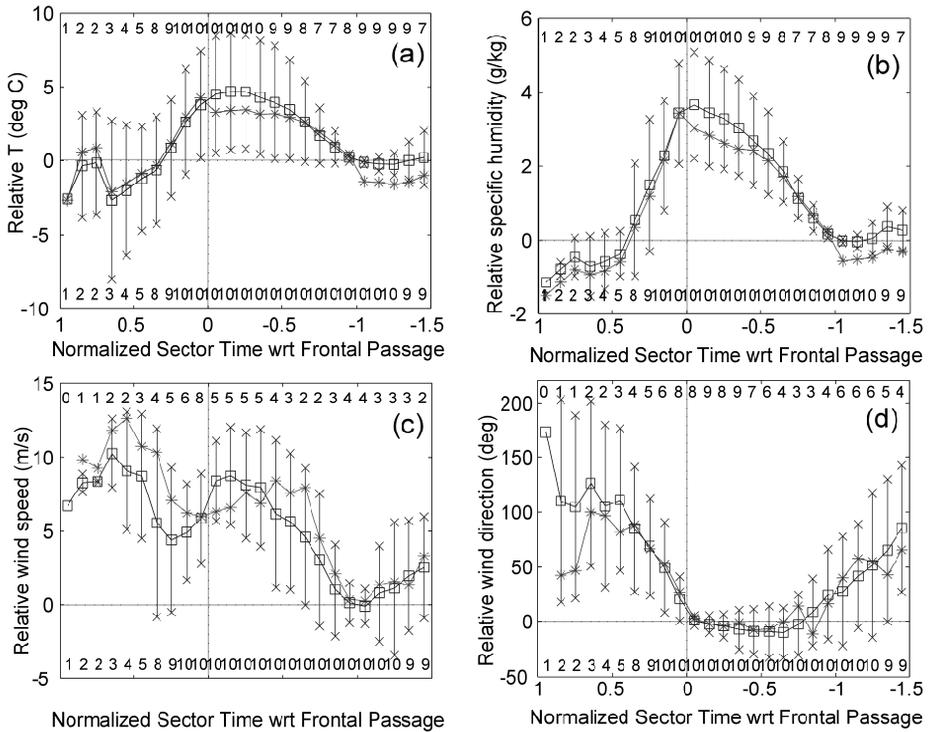


Figure 13. Composite relative values during FASTEX of: (a) temperature, (b) specific humidity, (c) wind speed, and (d) wind direction, with respect to the cold-frontal passage, from the RV *Knorr*. Values are relative to their values at the onset of the warm sector, except for wind direction which is relative to the value 1 h before frontal passage. The stars show the data from the ETL (see text) sensors and the squares those from the ship's Athena data system. The 'x's show  $\pm 1$  standard deviation of the Athena data. The numbers along the top and bottom of each frame show the number of cases that contributed to each composite bin for the ETL and Athena sensors, respectively. See text for details.

### (b) Oceanic

The significant wave heights increased from about 3.5 m in the eastern half of the warm sector to about 5.3 m at the time of frontal passage (Fig. 14(a)) and in the post-frontal region. The maximum wave heights ( $h_{\max}$ ) were about 5.5 m in the eastern half of the warm sector, increasing to about 9 m at the time of frontal passage. The ratio  $h_{\max}/h_{\text{sig}}$  abruptly increased from about 1.5 in the eastern part of the warm-sector region to about 1.75 during the time bracketing the CFP. The period of the significant waves was at a minimum of 7.5 s in the middle of the warm sector, and reached maxima of 8.3 s at the CFP and 8.5 s just east of the warm sector (Fig. 14(b)). Hence, wave growth and an increase of the wave period occur in the western half of the warm sector, and a systematic physical mechanism, perhaps due to a combination of the changing swell and wind directions in the vicinity of the front, causes occasionally taller waves. Note that the changes in wave height and period do not occur until after the onset of the increase in the wind speed at  $\tilde{t} = -0.9$ .

Since high-frequency (small period) waves tend to respond more quickly to changes in the wind than low-frequency (large period) waves (e.g. Rieder and Smith 1998), we find it useful to split the wave data into wave-period bins of: less than 3.4, 3.4–6.4, 6.4–9.5, and greater than 9.5 s. Using deep-water gravity-wave relationships, these period

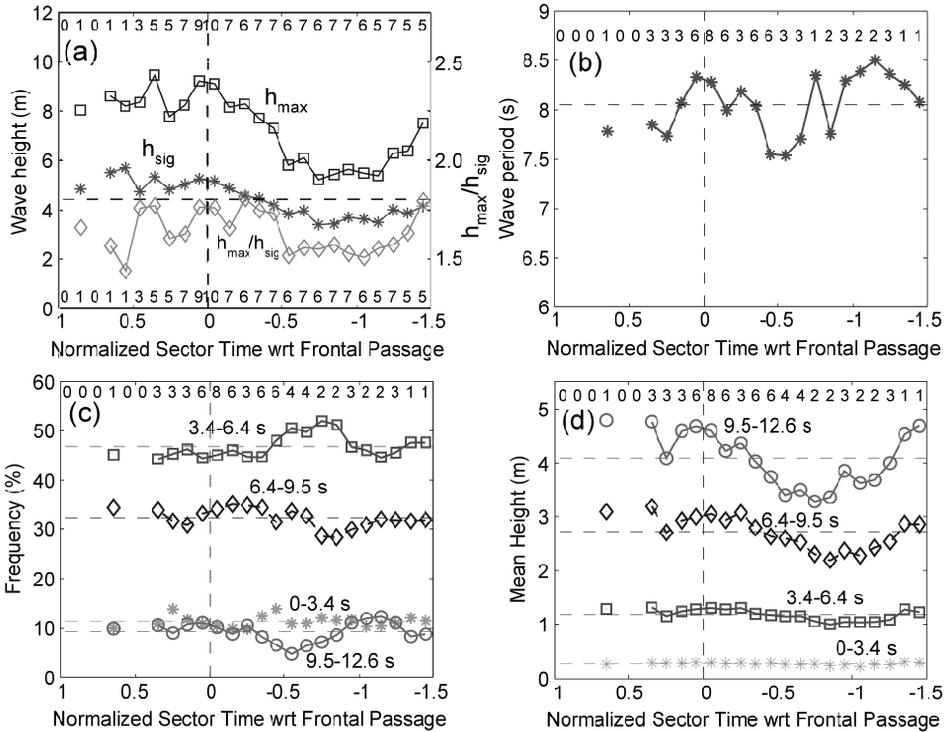


Figure 14. Composite values of: (a) significant (stars) and maximum (squares) wave heights and their ratios (diamonds), (b) significant wave period, (c) wave frequency for different wave-period bins, and (d) wave height for different wave-period bins, with respect to the cold-frontal passage from the RV *Knorr* during FASTEX (see text). The wave-period bins in (c) and (d) are 0–3.4 s (stars), 3.4–6.4 s, (squares), 6.4–9.5 s (diamonds), and 9.5–12.6 s (circles).

bins correspond to wavelength bins of: < 18, 18–64, 64–141, and > 141 m, respectively. We find the following:

- The frequency of the waves with relatively short 3.4–6.4 s periods peaks in the eastern half of the warm sector as the wind speed begins to increase, and then decreases as the winds continue to increase (Fig. 14(c));
- The frequency of occurrence of waves with periods >6.4 s increases in the western two-thirds of the warm sector at the expense of the waves with periods <6.4 s (Fig. 14(c));
- The mean height of the waves with periods >3.4 s increases in the warm sector (Fig. 14(d));
- The onset of the height increase and the peak in height occurs earlier (later) for the shorter (longer) period waves than for the ones with longer (shorter) periods. The longest-period waves have their maximum wave height in the post-frontal regime or east of the warm sector. These composite wave period changes are consistent with the case shown by Rieder and Smith (1998).

## 6. COMPOSITE SURFACE FLUXES

Turbulent fluxes were determined through the covariance technique, the inertial dissipation technique, and the bulk formulas of Fairall *et al.* (1996, 2003). The covariance

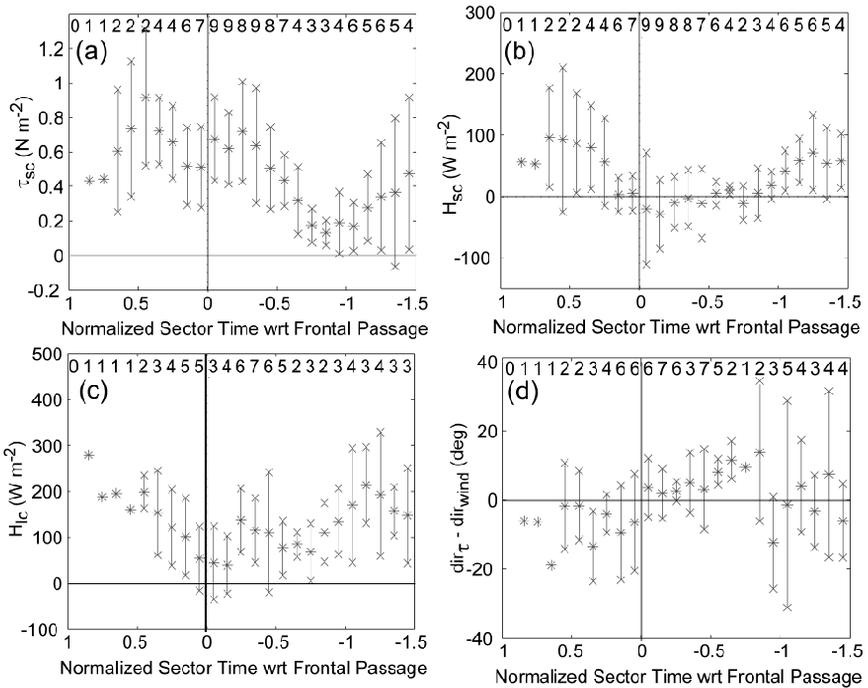


Figure 15. Composite values of: (a) stress, (b) sensible-heat flux, and (c) latent-heat flux determined from the covariance method with respect to the cold frontal passage from RV *Knorr* during FASTEX (see text); (d) shows the composite of the difference between the stress direction and wind direction. A three-point running mean was applied to the stress components before the stress direction was calculated. The vertical bars show  $\pm$  one standard deviation.

values show the increase of stress beginning at  $\tilde{t} \approx -0.75$  (Fig. 15(a)), shortly after the time the wind speed increases (Fig. 13(c)) and at the time the wave characteristics respond (Fig. 14). A peak stress of  $0.7 \text{ N m}^{-2}$  occurs shortly before the CFP at the time of maximum surface-layer wind speed, the peak in the occurrence of the 6.4–9.5 s period waves and near the peak in heights of these waves. Comparably high, or even slightly higher, values of stress occur in the post-frontal regime, roughly corresponding to the secondary wind speed maximum and the end of the change in wind direction.

The covariance sensible-heat flux ( $H_{sc}$ ) is a maximum in the post-frontal regime and before the WSB (Fig. 15(b)). Within the warm sector,  $H_{sc}$  decreases slowly as the front approaches the ship, becoming slightly negative just before the passage of the surface cold front. Qualitatively, this is consistent with the warming of the pre-frontal air through horizontal advection and surface-layer fluxes producing a stable environment nearest to the front, and hence negative  $H_s$  (e.g. Bond and Fleagle 1988). The latent-heat flux ( $H_{lc}$ ) also decreases within the warm sector (Fig. 15(c)) as the specific humidity increases (Fig. 13(b)). However, the specific humidity does not increase sufficiently to produce a negative  $H_{lc}$ . Hence, in contrast to  $H_{sc}$ ,  $H_{lc}$  remains positive. The maximum  $H_{lc}$  occurs just before the onset of the warm sector and at the very end of the post-frontal regime.

The streamwise and cross-stream covariance stress components can be combined to compute a stress direction. If the stress is due entirely to wind waves, then the stress direction should be the same as the wind direction. However, previous observations have noted that the stress and wind directions are often not the same, and that these directional

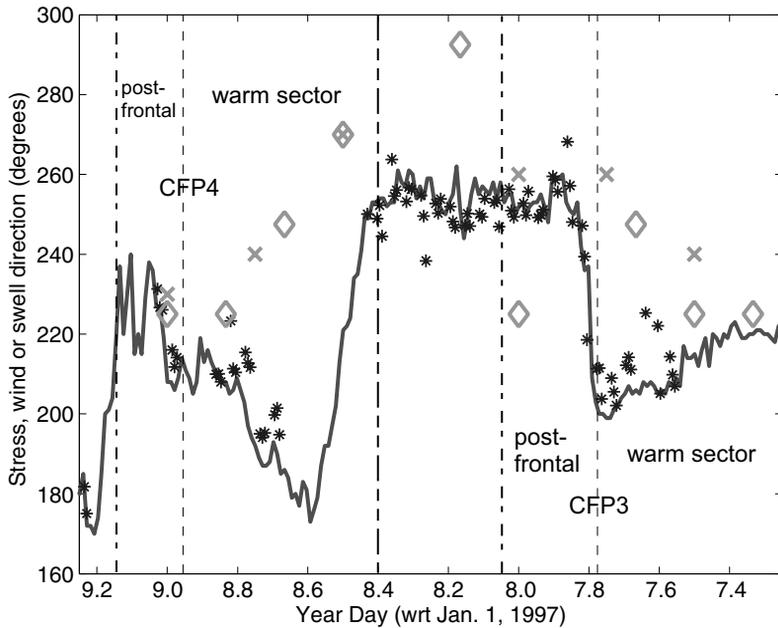


Figure 16. Time series from the RV *Knorr* during FASTEX of wind direction (line), stress direction (\*), and manual observations of the swell direction from two different ship logs (diamond and x) for the period Julian Day 7.3–9.25. The two vertical dashed lines show the cold-frontal passages for cases 3 and 4 (labelled CFP3, CFP4), the long-dashed and double-short dashed line shows the onset of the warm sector for case 4, and the two long-dashed/short dashed lines the two ends of post-frontal events. See text for details.

differences may be due to the effects of swell (Geernaert *et al.* 1993; Grachev *et al.* 2003) or baroclinicity (Geernaert 1996). The composite of the difference between the stress and wind directions (Fig. 15(d)) shows that in the central and western portion of the warm sector, the stress direction is often greater than the wind direction by 5–12° (that is, the stress direction is to the right of the wind direction); while in the post-frontal regime the directional difference is of the same magnitude but with opposite sign.

If swell is influencing the stress, then the stress direction should be between the swell and wind directions (Grachev *et al.* 2003). Visual observations of the swell direction by the ship's crew on the bridge of the RV *Knorr* show that the warm-sector stress direction is frequently between the swell direction and the wind direction. Figure 16 shows examples of two cases where the difference between the wind and swell directions occurs mainly in the warm sectors (note that the time axis runs from right to left in this figure). Hence, since the stress vector is influenced by the swell direction, the difference between the wind and stress directions occurs only in the warm sectors for these two cases. The post-frontal regime differences shown in Fig. 15(d) are also produced by other cases. Apparently, in some cases the pre-frontal swell direction can be influenced by the post-frontal swell direction, probably through different phase velocities of the swell and the cold front. Therefore, the swell direction appears to change before the wind direction near 7.5–7.8 in Fig. 16. These results suggest that in the vicinity of fronts the stress vector may not be an accurate indicator of the wind direction, either in the warm sector, post-frontal regime, or both.

ID calculations of the surface fluxes (Fig. 17) also show the same qualitative trends that were noted for the covariance fluxes (Fig. 15). However, the ID stresses are slightly

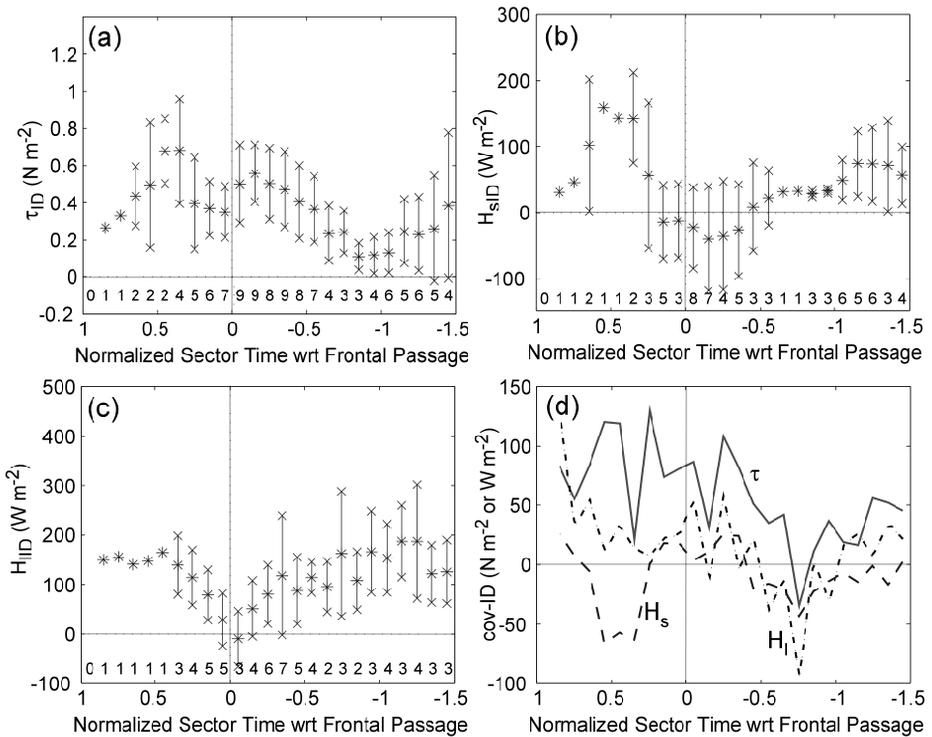


Figure 17. Same as Fig. 15 in (a) to (c), but for fluxes calculated from the inertial dissipation (ID) technique. In (d), the differences between the covariance and ID values of stress (solid),  $H_s$  (dashed) and  $H_l$  (dot-dash) are shown. Note that the stress differences have been multiplied by 500 to scale properly on the plot.

smaller than the covariance stresses, especially in the higher wind speed regime in the vicinity of the fronts. Furthermore, the ID sensible-heat fluxes are lower near the fronts and higher in other areas. The differences in  $H_l$  are not quite as systematic, though there is still a tendency for the ID  $H_l$  to be weaker in the vicinity of the front and stronger in the eastern half of the warm sector. These differences may reflect either flow distortion or ship-motion problems for the covariance technique at high wind speeds, or the inapplicability of the assumed inertial subrange characteristics at these wind speeds and/or environmental conditions (e.g. see the discussion in subsection 3(b) of this study, or in the appendix of Fairall *et al.* 2003).

The bulk fluxes (Figs. 18(a) to (c)) calculated from the measured basic parameters and the bulk relationships of Fairall *et al.* (1996, 2003) show the same general characteristics as discussed for the covariance and ID fluxes. However, the differences plot (Fig. 18(d)) shows that the bulk stresses are substantially (up to  $0.25 \text{ N m}^{-2}$ ) smaller than the covariance estimates, particularly in the post-frontal regime. Large and Pond (1981) attributed such a discrepancy in stress to the rougher seas after a front due to the change in wind direction, a phenomenon not included in parametrization schemes. Note that the maximum in  $\tau_{sc} - \tau_b$  does occur during the period of greatest wind direction change (compare with Fig. 13(d)). The secondary peak in stress error in the warm sector of about  $0.1 \text{ N m}^{-2}$ , suggests that the parametrization scheme also does not increase the stress sufficiently as the wind speed increases. The bulk  $H_s$  are within  $20 \text{ W m}^{-2}$  of the covariance estimates. The bulk  $H_l$  appear to be  $40\text{--}60 \text{ W m}^{-2}$  larger in most of the warm sector and  $20\text{--}40 \text{ W m}^{-2}$  smaller in portions of the post-frontal regime.

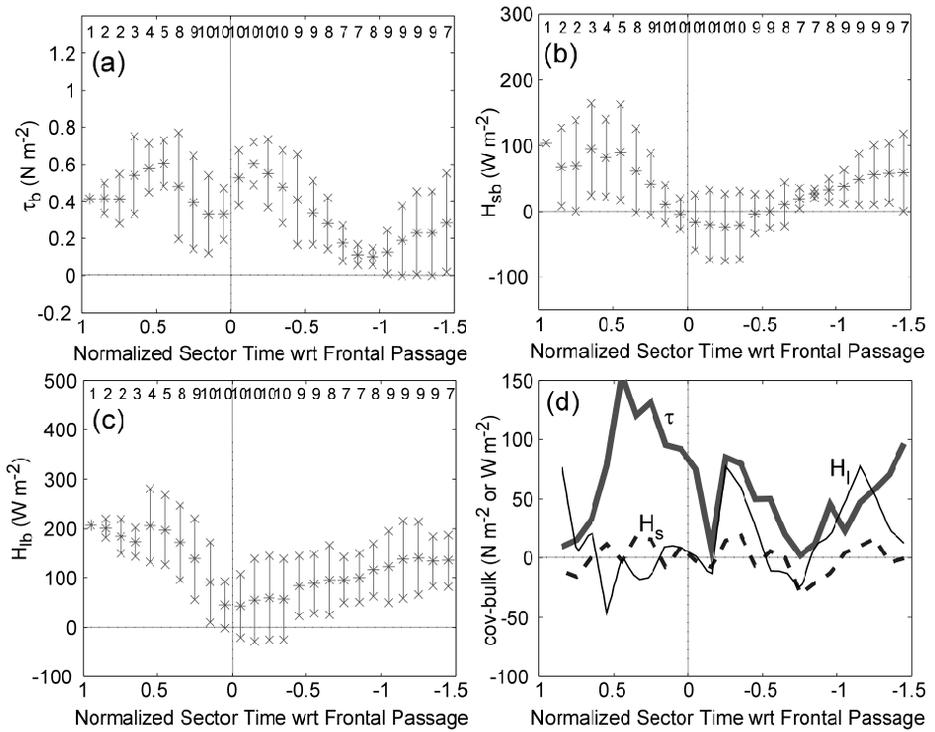


Figure 18. Composites from the RV *Knorr* during FASTEX of: (a) stress, (b) sensible-heat flux, and (c) latent-heat flux calculated from the bulk formulas of Fairall *et al.* (2003); the vertical bars show  $\pm$  one standard deviation. In (d), the differences between the covariance and bulk values of stress (heavy solid),  $H_s$  (dashed) and  $H_l$  (thin solid) are shown. Note that the stress differences have been multiplied by 500 to scale properly on the plot. See text for details.

The reasons for the differences between the covariance, ID, and the bulk fluxes may include: flow-distortion effects on turbulent eddies, errors due to the assumptions inherent in the inertial dissipation and bulk techniques, and real effects of the wind and wave conditions in the vicinity of the fronts.

## 7. DISCUSSION

While previous investigations have examined the contribution of synoptic variability to variations in surface fluxes for individual storms, to our knowledge this is the first attempt at compositing the fluxes from multiple storms. Such compositing allows us to make broader conclusions regarding the relationships between the surface fluxes and the synoptic environment than would be possible from only a single case-study. In studies of maritime storm systems, the synoptically forced variations of the near-surface atmospheric parameters are frequently considered to be damped compared to their land-based counterparts, primarily through the effects of surface fluxes. The variations in the composites for the surface-layer atmospheric parameters (Fig. 13) permit us to quantify these synoptically forced variations for the North Atlantic Ocean, showing that changes of 4–5 degC in temperature, 3.5–4  $\text{g kg}^{-1}$  in absolute moisture, and 8–12  $\text{m s}^{-1}$  in near-surface wind speed are typical variations associated with atmospheric cold fronts over the open ocean. Similar compositing studies can and should be carried out using a

longer database of buoy and ship data, in order to obtain variations of greater statistical significance than can be obtained with this 5-week dataset.

This compositing method also allows us to obtain insights into the synoptically modulated physical processes producing the observed surface fluxes. The variations of the surface sensible- and latent-heat fluxes indicate that they are primarily determined by the synoptically driven, low-level thermal and moisture advection. While relatively dry and cool air exists outside the warm-sector region of the cyclones, the strong southerly flow within the warm sector, reaching a maximum near the surface cold front, advects warmer and moister air northward. This advected air is typically warmer than the local SST, and hence produces a small downward sensible-heat flux. Upward sensible-heat flux may contribute to warming of the air in the slightly cooler eastern portion of the warm sector (e.g. Fig. 15(b)), which can then contribute to the warm-air advection as the confluent southerly flow brings this air closer to the front further north. Generally, the specific humidity of the warm-sector air, even closest to the front, is not as large as the saturated specific humidity given by the local SST, leading to a positive latent-heat flux in the warm sector of  $30\text{--}140\text{ W m}^{-2}$  which contributes significant moisture to the strong southerly flow. Both the sensible- and latent-heat fluxes are significantly larger in the post-frontal region where the advected air is cooler and drier, but they may be dynamically less important in this region.

The strong southerly winds near the cold front in the warm sector, and the even stronger west-north-westerly winds in the post-frontal regime, lead to peaks in the surface stress (Figs. 13(c) and 15(a)). The significant wave heights reach a maximum just before the CFP and remain high throughout the post-frontal period, while the waves prior to the warm sector and near the CFP have the longest periods (Figs. 14(a) and (b)). The wave conditions are likely to contribute to the variations in stress since, for instance, the bulk estimate of the stress, which does not explicitly account for variations in wave height or period, significantly underestimates the stress in the middle of the warm sector (when waves are growing and the peak frequency of waves with intermediate periods is reached—Fig. 14(c)) and in the first portion of the post-frontal region (when the wind direction is changing—Fig. 18(d)).

Systematic differences between the wind direction and stress direction occur, and change sign from the warm sector to the post-frontal regions (Fig. 15(d)). One possible explanation is that stress effects from swells are present (Geernaert *et al.* 1993; Rieder *et al.* 1994; Grachev *et al.* 2003) and that swell orientations change much less or more slowly across a cold front than does wind direction, as shown in Fig. 16. Physically, this may result either because the swells move faster than the cold front, so that post-frontal swell orientations are found ahead of the cold front, or because a cold front moves fast enough to not influence the swells over a long enough time period to change their orientation to that of the winds. Geernaert (1996) has also suggested that the thermal-wind effects can reorient the turbulent eddies, so the stress direction differs from the wind direction when significant thermal gradients in the low-level along-wind direction are present. Since strong along-wind thermal gradients are generally present within the warm sector and in the post-frontal regime, and the thermal wind will generally change less rapidly than the surface wind across a front accounting for the change in sign of the stress-wind directional differences, this mechanism is also plausible from our dataset. However, application of Geernaert's Eq. (12) produces angular differences between the stress and wind direction that are three to four times larger than observed, while the limited swell directional data (such as in Fig. 16) show that the stress direction is between the swell and wind directions, supporting the first hypothesis. Though these FASTEX observations show the presence of these stress-wind directional differences,

and show that they do occur in frontal regions of large thermal gradients and significant changes in wind direction, our data do not show conclusive evidence supporting one mechanism or the other; additional studies directly linking the swell to the stress are needed for this. Furthermore, the presence of these directional differences implies that satellite-based scatterometer wind directions, which rely on the surface stress field, will be in error, and will underestimate the surface directional wind shift across the front and thus the derivative fields such as convergence and vorticity. However, if such directional biases are seen in the scatterometer comparisons with other wind measurements, their magnitude is such that they may be ascribed to uncertainties in the observations.

The composited FASTEX data also suggests the possible influence of the surface fluxes on the maritime synoptic evolution in this region, which was an important objective of FASTEX. In an adjoint modelling study, Langland *et al.* (1995) show that the development of maritime cyclones is sensitive to the surface heat fluxes in the warm-sector regions. Other studies of the influence of surface sensible- and latent-heat fluxes on the development of maritime extratropical cyclones have shown that when heat fluxes are negative within the warm sector ahead of a cold front and positive in the post-frontal region, such a configuration does not promote development of the surface low (e.g. Haltiner 1967; Kuo *et al.* 1991) and may even be slightly detrimental to its development (e.g. Reed and Simmons 1991). However, when the surface heat fluxes are positive in the warm-sector region, especially during the earlier portions of the cyclone development, they contribute significantly to the cyclogenesis through decreased stability, increased moisture content and subsequent latent heating (e.g. Kuo *et al.* 1991; Zhang *et al.* 1999; Gyakum and Danielson 2000). Our study shows that the composite sensible-heat fluxes in the warm-sector region are near zero or slightly negative in these ten FASTEX storms, but that the composite latent-heat flux is positive. Hence, the composite sum of the surface-heat fluxes in the warm sector is positive, and we can conclude that the warm-sector surface heat fluxes, dominated by the latent-heat flux, should contribute to the development of the associated cyclones.

The observed variation of the surface stress also has implications for the dynamical feedback to the cyclone development. Low-level maxima of dry potential vorticity ( $PV_d$ ) along the cold front have been hypothesized to lead to the development of frontal waves and frontal cyclones (Joly and Thorpe 1990). These low-level  $PV_d$  maxima are believed to be produced by diabatic processes, especially latent heating in the main updraught near the surface cold front (e.g. Persson 1995; Stoelinga 1996). However,  $PV_d$  can also be modified by surface diabatic and frictional processes. Examination of the conservation equation for  $PV_d$ , shows that the observed increase in the surface sensible-heat flux toward the east in the warm sector would be likely to contribute to an increase in  $PV_d$ , while the observed decrease in stress toward the east in the warm sector would be likely to contribute to a decrease. More quantitative analysis of the vertical gradients in addition to the horizontal gradients is necessary to determine which term would dominate, and is beyond the scope of this paper.

## 8. CONCLUSIONS

Surface-layer and flux data were collected during FASTEX from the RV *Knorr* in high wind speed conditions in the North Atlantic during December 1996 and January 1997. This dataset includes the surface momentum, sensible-heat and moisture fluxes calculated using three different methods, SSTs, and wave characteristics, which were all used in this study. It is the only field experiment where ship-based covariance flux measurements in an open-ocean storm environment have been successfully obtained.

These measurements contribute significantly to the limited set of surface flux measurements available from strong wind environments, with 88, 77, and 44 hours of high-quality covariance momentum, sensible-heat, and latent-heat flux measurements, respectively, obtained for 10 m high neutral winds in the 15–21 m s<sup>-1</sup> range. The coincident measurement of wave heights through a microwave Doppler radar mounted in the bow of the ship adds additional value to this dataset, which also includes remote-sensor and sounding data to be used in future studies.

The 10 m neutral drag coefficients determined from the covariance technique are generally larger than those determined from the ID technique. It is unclear whether this difference indicates the presence of flow-distortion problems in the covariance data, or a failure of the assumptions inherent to the ID technique at these higher winds. The ID drag coefficients from the RV *Knorr* during FASTEX are consistent with the limited ID data from other measurements in open-ocean storm environments. No other datasets of covariance measurements in such an environment are available for comparison. This dataset also suggests that the 10 m neutral transfer coefficient for sensible-heat flux increases slightly at the higher wind speeds. The linear best-fit lines for all of the transfer coefficients are given by (3.2a) to (3.2c).

In order to illustrate the relationship between the surface layer and the synoptic atmospheric environment, composites of atmospheric surface-layer measurements and ocean-surface characteristics were computed along ship paths through ten storms for which the RV *Knorr* passed through the open-wave warm sector and the cold front. These composites, summarized in Fig. 19, show the following:

(i) The moistening and warming (Fig. 19(a)) associated with synoptic-scale advective patterns and surface-layer fluxes lead to minima in the sensible- and latent-heat fluxes just before the frontal passage (Fig. 19(b)), despite the strong surface winds at this time. Though the warm-sector sensible-heat flux minimum is slightly negative, the sum of the two heat fluxes is positive, suggesting a positive impact on the synoptic development of these systems.

(ii) The momentum flux is a maximum just before the frontal passage during the peak wind speed associated with the warm-sector LLJ (Figs. 19(a) and (b)). A second stress maximum of comparable magnitude occurs in the middle of the post-frontal regime. The patterns of heat and momentum fluxes should affect the surface potential-vorticity generation, and have dynamical implications for the stability of the frontal zone for frontal-wave development.

(iii) Wave heights increase steadily from the eastern half of the warm sector to the frontal passage, remaining high through most of the post-frontal regime before decreasing (Fig. 19(b)).

(iv) Differences between covariance and ID stresses are largest during the times bracketing the cold front when the wave heights and covariance stresses are large. Differences between covariance and bulk stresses are greatest in the pre-frontal LLJ, when the frequency of waves with intermediate periods of 6–9 s reaches its maximum, and in the post-frontal regime where wind direction veers.

(v) The stress direction is consistently 5–12° to the right of the wind direction in the western half of the warm sector, and 2–15° degrees to the left of the wind direction in the post-frontal regime, supporting previous observations (Fig. 19(b)). The data suggest that these differences are due to influences of swell orientations, but cannot exclude effects from thermal advection on the turbulent-eddy orientation. Their presence implies possible errors in satellite-based scatterometer measurements of surface wind fields.

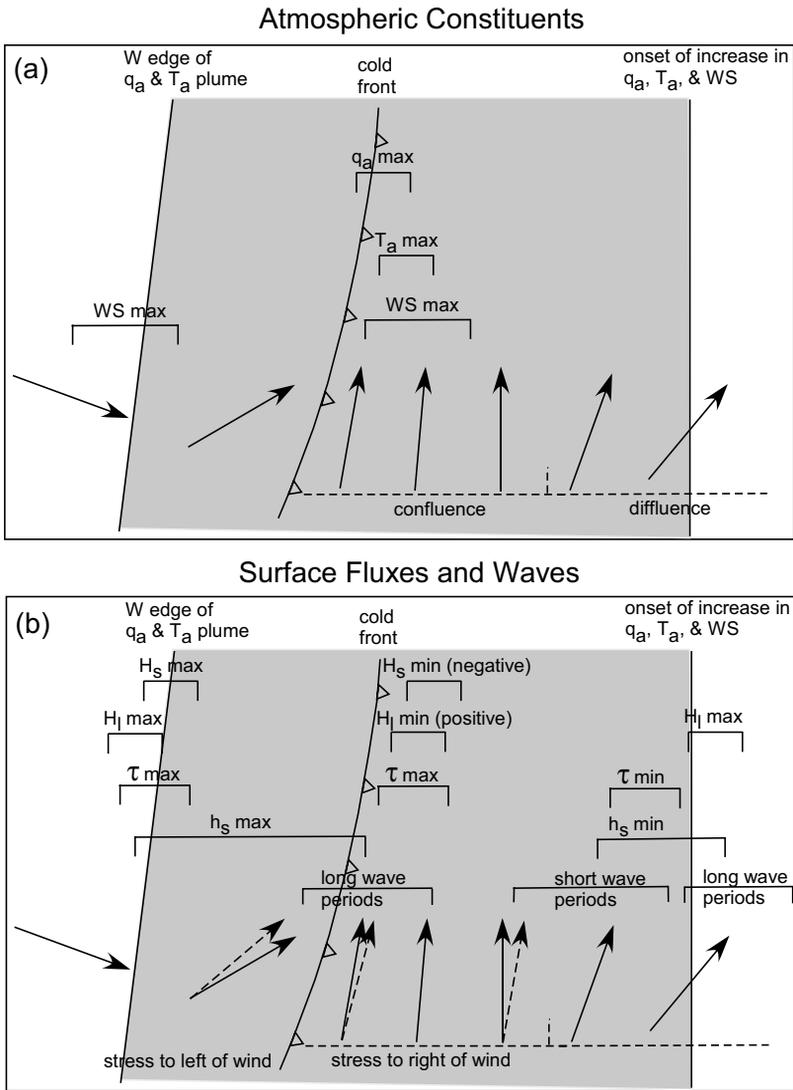


Figure 19. Schematic diagram summarizing composite variations from the RV *Knorr* during FASTEX (see text) of: (a) the atmospheric constituents, and (b) the surface fluxes and ocean waves relative to the warm sector, cold front, and post-frontal regions.  $T_a$ ,  $q_a$ ,  $WS$ ,  $H_s$ ,  $H_l$ ,  $\tau$ , and  $h_s$  represent atmospheric temperature, atmospheric water vapour, wind speed, sensible-heat flux, latent-heat flux, stress and significant wave height, respectively. The solid arrows show wind direction and the dashed arrows in (b) show stress direction. The region representing the atmospheric water vapour plume is shaded.

The FASTEX dataset has already been used for validation of the performance of surface flux parametrization schemes (Brunke *et al.* 2003; Fairall *et al.* 2003). Planned studies will utilize synoptic compositing of the atmospheric boundary-layer measurements to show how the synoptic environment modulates the boundary-layer structure and processes. Additional studies are required to examine the perplexing problem of how the surface waves influence the surface roughness in conditions of strong winds and swell, and how best to parametrize this effect for surface flux computations. Finally, the FASTEX dataset from the RV *Knorr* is useful as validation data for simulations

of the FASTEX storm environments, with a special focus on the impact of the surface fluxes and ocean characteristics on storm structure and development. The results of this current study give background and context to future studies.

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