TEMPERATURES

air T, T_{10}

potential temperature $\theta \approx T + [0.01\ Z\ (m)]^{\circ}$

virtual $T_v = T + 0.61\ q(T + 273^{\circ})$, T_{v10}, θ_v

temperature of dry air having same density

bulk water T_s

HUMIDITY

specific humidity q, q_{10} (kg/kg)

saturation humidity over fresh water $q_s(T_s)$; over seawater $0.98\ q_s$

DENSITY

air $\rho(T,q,P) = \rho(T_v,\rho)$ (kg/m^3)

water vapour $\rho_v = \rho q$

TURBULENT FLUXES IN SURFACE LAYER (Constant with height Z)

wind stress $\tau = -\rho\langle u'w'\rangle = \rho u_*^2$

sensible heat $H = \rho c_p \langle t'w'\rangle = -\rho c_p u_* t_*$

evaporation $E = \rho\langle q'w'\rangle = -\rho u_* q_*$

vertical wind fluctuation w'

specific heat of air c_p

wave phase velocity C_p

BULK FORMULAS

wind stress $\tau = \rho C_{10} U_{10}^2 = \rho C_{10N} U_{10N}^2$

U_{10N} is "Neutral" wind at Z=10 from neutral log wind profile, same u_*.

sensible heat flux $H = \rho c_p C_H U_{10}(T_s - \theta_{10})$

evaporation $E = \rho C_E U_{10}(0.98q_s - q_{10})$

Given C_{10}, C_H and C_E, fluxes estimated from bulk parameters.

PROFILE FORMULAS

$U(Z) = (u_*/\kappa)[\ln(Z/z_o) - \Psi_m(Z/L)]$

$\theta(Z) - T_s = (t_*/\kappa)[\ln(Z/z_t) - \Psi_t(Z/L)]$

$q(Z) - 0.98q_s = (q_*/\kappa)[\ln(Z/z_q) - \Psi_q(Z/L)]$

von Karman $\kappa = 0.4$

EFFECTIVE surface roughness z_o, z_t, z_q

Stability $Z/L = \kappa g z(t_*/T + 0.61q_*)/u_*^2$, L is Monin-Obukhov length

ROUGNNESSES UNIQUELY SPECIFY COEFFICIENTS or vice-versa: from profile formulae

$C_{10N} = (u_*/U_{10N})^2 = [\kappa/\ln(Z/z_o)]^2$

$C_{HN} = \kappa^2/[\ln(Z/z_o)\ \ln(Z/z_t)]$

$C_{EN} = \kappa^2/[\ln(Z/z_o)\ \ln(Z/z_q)]$

Wind and Windstress

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and
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Fields of wind stress at an ocean basin scale are almost exclusively determined from wind fields by aid of parameterisations. We therefore discuss wind and wind stress together, since any bias which besets the derived wind field will affect the flux fields too. Before we discuss fields it seems appropriate to point to some difficulties, that often have affected the use of wind data.

Wind is not wind

For a long time the term "surface wind" has been used in common by modelers and experimentalists without even noticing the difference in meaning. In a numerical model surface wind is the wind at a gridpoint at the lowest computational level (not necessarily equal to 10 m), that corresponds to a vector averaged wind and is influenced by the physics of the model. In observations, the wind speed is a local variable, averaged over 10 minutes, and nominally taken at 10 or 25 m height. Though we use the same word, we talk about different physics. A similar remark holds for satellite retrievals. It is important that we understand and account for the difference in physics (Gulev, 1994), otherwise modelling will not gain from improved experimental results.

In marine climate data sets winds are given as speeds, coded in knots. It is worth noting that most of these were originally obtained as Beaufort estimates of wind strength. In order to convert Beaufort forces into velocities, a so-called Beaufort equivalent scale is used. There are two basic problems with an equivalent scale: First, the Beaufort strength is estimated from characteristics produced by the action of the wind. The signatures e.g. of sea state represent different physics than a 10 minute average of a local wind at 10 m height. Second, the scale may have slipped. Originally the Beaufort scale was defined by the wind force exerted at a typical sailship. Later, the scale was re-defined in terms of visible properties of the sea surface.

Typically an Beaufort equivalent scale is obtained by comparing Beaufort estimates from passing ships with wind speed measurements on suitable islands or ocean weather ships. Due to natural variability and error variances and different statistical treatment, such derivations give slightly different results. Most climate data are converted by the official WMO scale code 1100. For research, WMO permits the use of a "scientific" scale. Together with Bunker's data we used a scale by Kaufeld (1981), similar to the WMO scientific scale. It showed that this scale is not suitable with the COADS (Consolidated Ocean Atmosphere Data Set). We therefore derived a new scale (figure 1) from measurements on ocean weather ships and observations by passing ships, carefully considering the respective error variances (Lindau,1995a). The scale is similar to WMO code 1100 with lower wind speeds at higher Beaufort forces. We believe that this is the appropriate scale for use with COADS. The new scale optimally translates Beaufort forces into wind speeds, hence the error variances of the Beaufort estimates need to be considered when deriving statistics from these data.

Peterson and Hasse (1987) demonstrated that the Beaufort scale has shifted from the time of sailships until today. A time dependent correction is possible by reference to sea level air pressure gradients (Lindau et al., 1995b). The technique merely implies that the relationship between surface wind and geostrophic wind has not changed with time on average over a large number of data. The results show that the wind climate at the North Atlantic Ocean has not changed noticeably during the last century (Figure 2) when the Beaufort scale is referred to a common base period. It is evident that in work with climatological wind data one should use a correction to eliminate spurious trends. This is important as sailships travelled areas other than modern shipping lanes and a mix of uncorrected data may result in spurious gradients, especially in operations like div or curl. A drift of input data could also induce spurious trends in the results of reanalysis projects.

Coastal effects: The different roughness of land and sea provides for an acceleration of the wind near the shore, both for onshore and offshore winds. A study of the surface wind over the Baltic Sea has been used to empirically quantify this effect; preliminary results are given in Table 1. The transition from higher speeds over sea to lower speeds over land is mainly concentrated in the first 10 km to the shore. The spatial variation of wind speed is important to consider for studies of wave generation near shore.

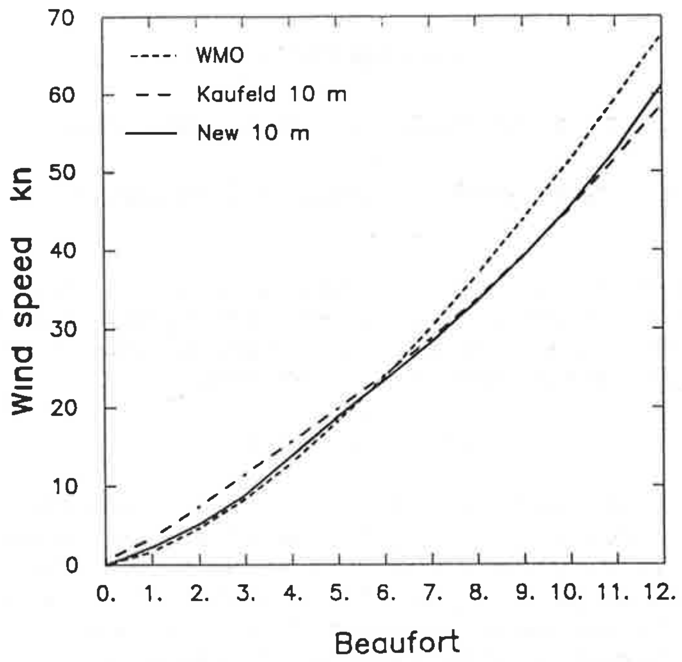

Figure 1 Comparison of Beaufort equivalent scales, from Lindau (1995a).

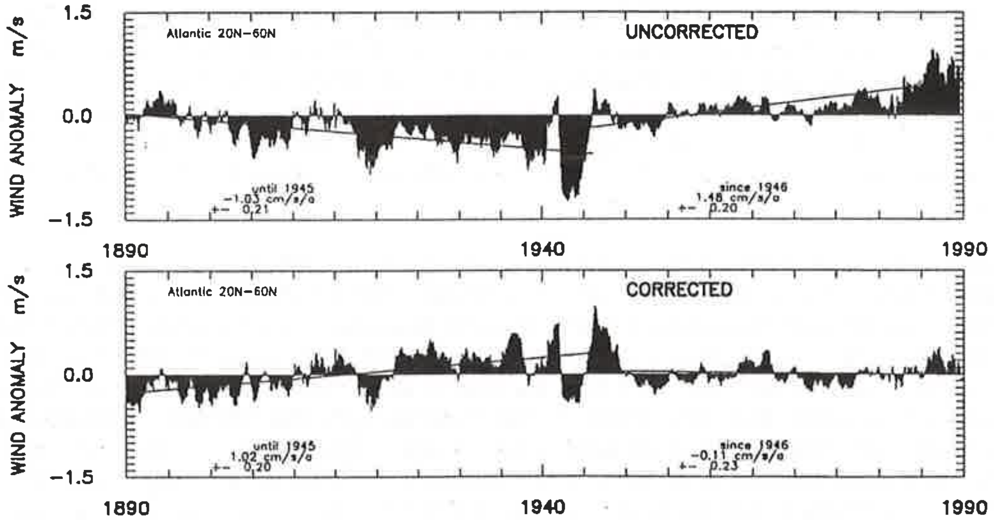

Figure 2 Variation of mean wind speed at the North Atlantic Ocean, upper panel uncorrected, lower panel corrected for drift of Beaufort equivalent scale, from Lindau (1995b).

offshore distance	onshore distance					
	1-5 km	5-10 km	10-20 km	20-30 km	30-50 km	>50 km
1-5 km	0.52	0.56	0.59	0.62	0.65	0.66
5-10 km	0.57	0.59	0.62	0.66	0.69	0.70
10-20 km	0.61	0.63	0.66	0.70	0.74	0.75
20-30 km	0.65	0.68	0.72	0.76	0.79	0.80
30-50 km	0.69	0.73	0.76	0.82	0.85	0.86
>50 km	0.70	0.74	0.78	0.84	0.86	0.88

Table 1: Ratio $\gamma = U_{beo}/U_{geo}$ of observed surface to analysed geostrophic windspeed for classes of onshore and offshore distances to the coast.

Parameterizations of wind stress

Wind stress for use in numerical models typically is determined from surface layer variables by parameterisations. In the present contribution we assume that the necessary informations are available either from surface layer observations or other data with similar characteristics.The most commonly used tool is the socalled bulk aerodynamic formula, where the drag coefficient. c_D represents the influence of turbulence on the relation between wind and stress and thus depends mainly on stability and seastate and/or wind speed. Hence c_D itself needs to be parameterized. c_D is commonly referred to a height of 10 m above mean sea level. The wave influence decays exponentially with height z as -kz, where k is wavenumber of the dominant waves. The influence of stability increases linearly with height. Thus it is reasonable to treat the two influences as if they were independent. This assumption may seem less plausible under conditions of high sea state/wind speed, but then the stability is near neutral anyway.

The drag coefficient is meant to parameterize the total air sea momentum flux, while nearer to the sea surface the flux is thought to be partitioned into a wave induced and a turbulent contribution. The influence of stability in the surface layer can well be treated by the Monin-Obukhov similarity theory, that is not reproduced here, see e.g. Smith (1988), Isemer and Hasse (1985), or Gulev(1995).

As an alternate to the drag coefficient, the roughness length z0 is often used. It is defined through the logarithmic windprofile, applicable in the constant stress layer under neutral conditions only.The roughness length is related to the neutral drag coefficient through the logarithmic profile. Considering that in the presence of waves the stress is not constant with height in the surface layer, the roughness length has no real meaning for use over sea. Hence it should not be used in dimensional arguments. This to my judgement also precludes the use of the so-called Charnock relation (Charnock, 1981)

The dependence of the momentum transfer or the drag coefficient on sea state has been recognized early and has remained a matter of research until today. Smith (1980) found different dependences of drag coefficient with wind speed for limited and unlimited fetch. During HEXOS (Smith et al.,1992) a large set of data with wind speeds up to 20 m/s has been obtained. The general result that the momentum transfer to the sea (expressed by a drag coefficient) is less effective with increasing wave age (expressed by the ratio of windspeed to wave phase velocity) as depicted in Figure 3.

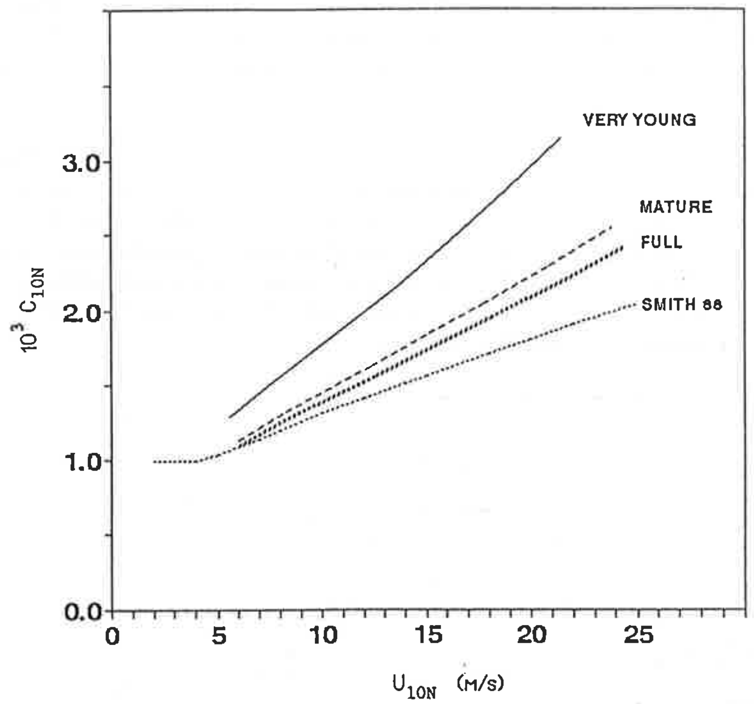

Figure 3 Influence of sea state on momentum transfer, from Smith et al. (1992)

It is difficult, however, to give a reliable numerical formulation for the process. The problem is seen in the fact that a fit between roughness and wave age best would be made in terms of nondimensional variables. Unfortunately, from a theoretical point of view, friction velocity is preferred for normalisation, that itself is difficult to measure and hence beset with comparatively large error variances. Hence in a relation that contains friction velocity on both sides, a spurious variation is induced. Unfortunately Figure 3 is calculated from such a fit instead from the data directly. From our understanding of the physics of the process, we can say that the figure shows the pattern of the wave influence, but the magnitude likely is overstated. The matter is even more complicated by the ubiquitous coexistance of wind sea and swells from different directions. Wind driven seas exchange momentum with the air above, while the momentum exchange of swell is negligible. In an experimental situation it is difficult to differentiate between wind seas and swell. This likely is a reason too for the limited success of scaling with wave height.

The dilemma that we face here is somewhat alleviated by the correlation between wind speed and waves. The conventional increase of drag coefficient with wind speed in my opinion can be seen as a first order parametrisation of increasing non-equilibrium of wind and sea at higher wind speeds, resulting in lower wave age and higher roughness. The formulations e.g of Smith (1980)

$$10^3\ c_{DN} = 0.61 + 0.063\ U10N\ /\ ms^{-1}$$

and Large and Pond (1981)

$$10^3\ c_{DN} = \begin{cases} 1.14 & \text{for } U10 < 10\ \text{m/s} \\ 0.49 + 0.065\ U10\ /\ ms^{-1} & \text{for } 10 < U10 < 26\ \text{m/s} \end{cases}$$

in effect are rather similar to other open ocean results from direct and profile measurements and are recommended. These could be called local scale drag coefficients. For use as parameterisations in numerical models, the subgrid scale variations need to be considered explicitely.

Available fields of air sea interactions

J. Oberhuber has dealt with global fields in a preceeding section with a necessarily coarse resolution. The density of ship observations at the Atlantic Ocean makes a higher resolution possible (Bunker,1976; Isemer and Hasse,1985,1987; Lindau,1996) In the Bunker Climate Atlas of the North Atlantic Ocean, wind and stress fields on a 1x1° grid are given between the equator and 65°N as monthly and annual fields together with derived fields, e.g. curl of stress. We have kept Bunker's treatment of stability and windspeed dependence, but adjusted the mean level of c_{DN} to better correspond to flux determinations (eddy correlation and profiles) from the open ocean.

The larger number of data available now in COADS suggests a re-evaluation of the climate of the Atlantic andan extensionof the study to the South Atlantic too. A climate atlas of the Atlantic Ocean has been derived (LINDAU,1996), based on COADS and using todays best available parameterisations for the heat balance at the sea surface. It is interesting to see that in the long term mean the heat loss of the ocean to the atmosphere at the Atlantic north of 35°S is balanced by advection in agreement with independent oceanographic evidence (Figure 4) without use of any constraint.This result implicitely supports the use of the new Beaufort equivalence scale.

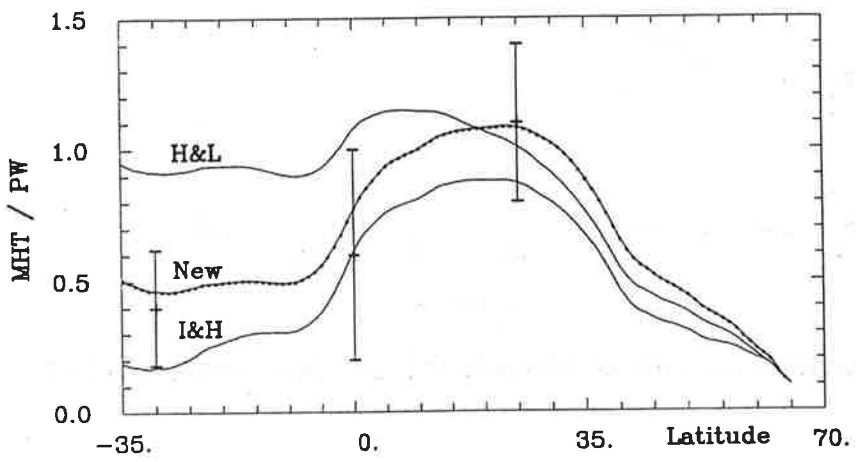

Figure 4 Meridional heat transport calculated from meteorological air sea fluxes compared with oceanographic data as a test of consistency of parameterisations, from Lindau (1996).

The fields of observations and air seas fluxes from the Bunker data set of the North Atlantic Ocean (Isemer and Hasse 1985, 1987) are available on mass storage devices from IfM Kiel, NCAR, WHOI. The COADS climate atlas of the Atlantic Ocean (Lindau 1996) is in preparation for the publisher, fields are already available for exchange at request.

We have analysed surface wind fields using wind and pressure observations to obtain a high resolution. Fields are available for the North Atlantic Ocean for 12.00 GMT from 4/1982 through 3/1985 and running since 10/1991, also for the Baltic Sea at 0, 6, 12, and 18 GMT from 1/1992 through 12/1994.

Research Needs

It has been recognised by researchers that the "inertial dissipation technique" of stress determination at sea is not as simple (Wucknitz 1979) as thought in the beginning. At the suggestion of Dr. S.D.Smith IAMAS/ICDM has recommendedan international intercomparison of such techniques. This is important because the "dissipation" technique is typically used to determine sea truth for algorithms of stress estimation from satellite remote sensing.

The observed variability of the roughness of the sea surface in natural wave fields can still be only insufficiently be parameterized. This is seen as the most important factor for improvement of the description of air sea momentum flux. A renewed effort to improve the knowledge of energy input and dissipation in a sea state is necessary in order to improve these terms in the otherwise already very successful wave modelling.

There have been reports of disagreement of directions of stress and wind in coastal areas.
Smith (1980) reported an extensive series of stress measurements made at a mast 10 km offshore SE of Halifax in deep water. It would be interesting to compare wave model data with these stress observations.

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OBSERVATIONS OF SURFACE FORCING FROM THE SUBDUCTION EXPERIMENT AND TOGA COARE: A COMPARISON WITH GLOBAL MODEL PRODUCTS AND CLIMATOLOGY

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1. INTRODUCTION

Reliable estimates of the exchange of heat, moisture, and momentum between the atmosphere and ocean are of great interest to atmospheric scientists and oceanographers alike. Such estimates are needed to verify the local "representativeness" of the gridded surface forcing fields offered by global atmospheric model and climatological data sets. Modelers from both communities rely on these gridded forcing fields to provide the proper boundary conditions for their models. Unfortunately, high quality, long-term, *in-situ* flux estimates are rare over marine areas. Fortunately, the surface mooring data collected during the Subduction Experiment and the Tropical Ocean Global Atmosphere (TOGA) Coupled Ocean Atmosphere Response Experiment (COARE) meet these requirements, making these two data sets ideally suited for validation purposes.

2. AN OVERVIEW OF THE DATA SETS

In an effort to more fully understand oceanic subduction, the process by which mixed layer water is injected into the main thermocline, the Subduction Experiment was undertaken in the eastern subtropical North Atlantic. A large-scale five mooring array was maintained in this area between June, 1991 and June, 1993 (Brink et al., 1995). The moorings were located at 33°N 34°W, 33°N 22°W, 18°N 34°W, 18°N 22°W, and 25.5°N 29°W and are referred to by their relative positions (NW, NE, SW, SE, and C) within the array (Figure 1a). Although several of the moorings did not remain on station continuously for the duration of the experiment, the overall scope and quality of the Subduction mooring data are exceptionally good.

Many of the Subduction moorings carried two independent meteorological instrument systems: a Vector Averaging Wind Recorder (VAWR) and an Improved METeorological recorder (IMET). Both systems measured barometric pressure, wind speed and direction, air temperature, sea temperature, relative humidity, and incoming shortwave and longwave radiation. The accuracy of the measurements was assured through extensive intercomparisons and pre and post-deployment instrument calibrations.

A principal aim of TOGA COARE was to obtain accurate observations of the near surface meteorology and air/sea fluxes in the western equatorial Pacific warm pool region (Webster and Lukas, 1992). One of the primary platforms used in this endeavor was a surface mooring located in the middle of the warm pool at 1° 45'S and 156°E (Figure 1b). Dynamic, thermodynamic, and radiometric data were collected by VAWR and IMET systems aboard this mooring between October 21, 1992 and March 4, 1993. The veracity of the COARE data was substantiated by complementary data from ships, aircraft, satellites, and buoys operating within the Intensive Flux Array (IFA).

Model data are from the global numerical weather prediction analysis/forecast systems of the European Centre for Medium-Range Weather Forecasts (ECMWF) and the National Meteorological Center (NMC). The models' surface analyses represent snapshots of the surface meteorology at the analysis times of 0, 6, 12, and 18UTC. The model fluxes are archived as accumulated values over an initial 6hr forecast from which 6hr averages were subsequently computed. The Subduction data are compared to operational products generated by the T106/L19 ECMWF model prior to Sept. 17, 1991 and the T213/L31 ECMWF model at all times thereafter (ECMWF Technical Attachment, 1994). The Subduction data are also compared to products generated by the operational version of the T126/L28 NMC model, although the NMC archive did have occasional gaps of time in which data were unavailable (Kanamitsu et al., 1991). The COARE data, on the other hand, are compared to initialized analyses and forecast fields from special runs of the operational ECMWF model (circa Fall 1994), but with a lower resolution of T106.