AIR-SEA INTERACTION IN THE ECMWF MODEL

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Abstract

The basics of the parametrization of atmosphere-ocean exchange are described in the context of the ECMWF model and the importance of an accurate representation of air-sea interaction is demonstrated with help of a few examples of model sensitivity. Furthermore, the ocean surface cool skin and warm layer effects are described. It turns out that the cool skin and warm layer effects have only a marginal impact on the model climate, but may be relevant to the use of satellite data.

1 Introduction

Transfer of momentum, heat and moisture in the lowest meters of the atmosphere is often referred to as air-sea interaction and a realistic representation of turbulent transport in this layer is an essential aspect of state of the art numerical models of the atmosphere. The scheme for air-sea transfer provides the surface boundary condition for wind temperature and moisture and controls the way the sea surface temperature (SST) is communicated to the atmosphere. Parametrization of air-sea transfer also controls the surface fluxes of momentum, heat and moisture, which are not only important for atmospheric models, but also for the ocean models in coupled integrations. The reason that models are sensitive to the parametrization of air-sea interaction is that the gradients of wind temperature and moisture are very steep in a shallow layer near the ocean surface and therefore the parametrized fluxes depend vey much on the precise formulation of the surface layer scheme. It does not mean that surface fluxes depend on the surface layer parametrization only. Surface fluxes are also determined by the boundary layer structure. If we think of the boundary layer as a reservoir (e.g. a mixed layer) of momentum, heat and moisture, then it becomes clear that all processes affecting the contents of the reservoir have impact on the surface fluxes. If for instance the boundary layer is ventilated from the top through dry entrainment or through moist convection, then the boundary layer becomes dryer and the surface evaporation over the ocean increases (Tiedtke et al., 1988). Moist convection and processes at the top of the boundary layer are not considered in this paper. For a discussion of boundary layer processes in the ECMWF model we refer to Beljaars and Viterbo (1998).

In this paper different aspects of air-sea interaction will be reviewed with emphasis on heat and moisture fluxes. This will be done by describing the ECMWF scheme and by showing the impact on the model of several changes that were made in the past. Some of these changes had a big impact on the model climate and the model's capability to respond to SST anomalies. Other changes had negligible or negative impact in spite of being supported by observations. Such a model response can be due to compensation of errors by other model parametrizations.

Recent observational programs have emphasized the role of ocean skin temperature variations (e.g. Fairall et al., 1996b), the impact of which is also discussed in this paper.

2 The surface layer formulation

The standard way of expressing (kinematic) surface fluxes of momentum (u_*^2; square of the friction velocity), sensible heat ($\overline{w'\theta'_o}$) and latent heat ($\overline{w'q'_o}$) into wind, temperature and moisture differences over the surface layer is with help of transfer coefficients (e.g. Brutsaert, 1982; Stull, 1988):

$$u_*^2 = C_m \, |\vec{V}_1|^2 \,, \tag{1}$$
$$\overline{w'\theta'}_o = C_h \, |\vec{V}_1| \, (\theta_s - \theta_1) \,, \tag{2}$$
$$\overline{w'q'}_o = C_q \, |\vec{V}_1| \, (q_s - q_1) \,, \tag{3}$$

where C_m is the transfer coefficient for momentum (drag coefficient), C_h is the transfer coefficient for heat, C_q is the transfer coefficient for moisture, $|\vec{V}_1|$ is the absolute horizontal wind speed at the lowest model level, θ_1 and q_1 are potential temperature and specific humidity at the lowest model level, and θ_s and q_s are potential temperature and specific humidity at the surface. In accordance with Monin Obukhov similarity theory, the transfer coefficients can be written in terms of profile functions containing a logarithmic part, with roughness lengths as surface characteristics, and a stability function describing the effect of stability as a function of the Obukhov length:

$$C_m = \frac{k^2}{[ln(z_1/z_{om}) - \Psi_m(z_1/L)]^2} \,, \tag{4}$$

$$C_h = \frac{k^2}{[ln(z_1/z_{om}) - \Psi_m(z_1/L)] \, [ln(z_1/z_{oh}) - \Psi_h(z_1/L)]} \,, \tag{5}$$

$$C_q = \frac{k^2}{[ln(z_1/z_{om}) - \Psi_m(z_1/L)] \, [ln(z_1/z_{oq}) - \Psi_h(z_1/L)]} \,, \tag{6}$$

where L is the Obukhov length ($=-u_*^3/(\overline{w'\theta'_v} \, k \, g/\theta_{vo})$, θ_v represents virtual potential temperature; see Stull, 1988; Garratt, 1992), k is the VonKarman constant, g the constant of gravitation, z_{om}, z_{oh}, z_{oq} are the roughness lengths for momentum, heat and moisture, z_1 is the height of the lowest model level (about 30 m in the ECMWF model), and $\Psi_{m,h}$ are the stability functions for momentum and heat/moisture. In the ECMWF model a displacement height z_{om} is added to z_1, but over the ocean this effect is negligible because $z_1 >> z_{om}$ and therefore the displacement height is not included in the formulae here. The transfer coefficients can also be written as

$$C_m = C_{mn} F_m(Ri_b, z_1/z_{om}, z_1/z_{oh}) \,, \tag{7}$$
$$C_h = C_{hn} F_h(Ri_b, z_1/z_{om}, z_1/z_{oh}) \,, \tag{8}$$
$$C_q = C_{qn} F_h(Ri_b, z_1/z_{om}, z_1/z_{oh}) \,, \tag{9}$$

where the neutral transfer coefficients are defined by the logarithmic part of the profile functions

$$C_{mn} = \frac{k^2}{[ln(z_1/z_{om})]^2} \,, \tag{10}$$

$$C_{hn} = \frac{k^2}{[ln(z_1/z_{om})] \, [ln(z_1/z_{oh})]} \,, \tag{11}$$

$$C_{qn} = \frac{k^2}{[ln(z_1/z_{om})] \, [ln(z_1/z_{oq})]} \,, \tag{12}$$

and F_m, F_h and F_q are stability functions dependent on the bulk Richardson number $Ri_b = (g/\theta_v) z_1 (\theta_{vs} - \theta_{v1})/|\vec{V}_1|^2$, and the ratios z_1/z_{om} and z_1/z_{oh}. It can easily be shown that Ri_b is related to z_1/L, so the functions F_m and F_h can be derived from the Monin Obukhov stability functions for which the main body of empirical information exists (see Högström, 1988 for a review).

The discussion of air-sea transfer is not about the validity of the approach described above but about the details of parameter and function choices. Three aspects will be discussed in the following subsections: (i) the neutral transfer coefficients or roughness lengths, (ii) the stability functions, and (iii) the definition of absolute horizontal wind at the lowest model level.

2.1 Neutral transfer coefficients

Neutral transfer coefficients and surface roughness lengths are compatible concepts as is clear from equations (10)-(12). However, to convert from one to the other, a reference height z_1 is needed. In observational studies, z_1 is the measuring height, in models it is the height of the lowest model level. To have a parametrization that is independent of the vertical discretization it is better to prescribe the roughness lengths rather than neutral transfer coefficients. A substantial part of oceanographic literature on air-sea interaction reports neutral transfer coefficients for a reference height of 10 m (e.g. Large and Pond, 1982), but this information can easily be translated into roughness lengths. The sea surface roughness lengths depend in principle on wave parameters, but for this aspect we refer to the paper by Janssen in these seminar proceedings. For use in numerical weather prediction and climate models, a parametrization of roughness lengths as a function of friction velocity with help of the Charnock relation ($z_{om} = z_{oh} = z_{oh} = \alpha u_*^2/g$) is often sufficient. Smith (1988) reviews oceanographic data and reports a range of observed α values of 0.01 to 0.02 and recommends 0.011 for the open ocean. ECMWF uses a value of 0.018 for the Charnock parameter but the difference is only marginally significant given the uncertainty in the data.

The ECMWF model that was operational until 1990 had equal roughness lengths for momentum heat and moisture and used the Charnock relation for the ocean ($z_{om} = z_{oh} = z_{oh} = 0.018\, u_*^2/g$), resulting in an increase of transfer coefficients with wind speed, which is realistic for momentum, but rather unrealistic for heat and moisture (DeCosmo, 1991). This was changed in 1993 when smooth surface scaling was added to the Charnock relation for momentum and smooth surface scaling only is used for heat and moisture (see Beljaars, 1995a):

$$z_{om} = 0.11\nu/u_* + 0.018 u_*^2/g\,, \tag{13}$$

$$z_{oh} = 0.40\nu/u_*\,, \tag{14}$$

$$z_{oq} = 0.62\nu/u_*\,, \tag{15}$$

where ν is the kinematic viscosity of air ($1.5\,10^{-5}\,m^2/s$). The increase of roughness lengths at low winds is well accepted now and also supported by ocean data (Bradley et al., 1991). The smooth surface scaling for heat and moisture together with the formulation for momentum, provide fairly constant transfer coefficients at high winds. The numerical value for C_{qn} is about $1.2\,10^{-3}$ above 5 m/s for a reference height of 10 m which is in close agreement with the conclusions from a review by Smith (1989) and data from the HEXOS experiment (Katsaros et al., 1987; Smith et al., 1990; DeCosmo, 1991).

2.2 Stability functions

Stability functions in the framework of Monin Obukhov similarity theory are well established, although their precise form is a subject of ongoing research (see Högström, 1988). The current

ECMWF formulation uses Monin Obukhov similarity functions dependent on z/L and solves the relation between z_1/L and Ri_b iteratively giving maximum flexibility for the specification of stability functions and surface roughness lengths[1]. The Paulson (1970) functions are used for unstable situations

$$\Psi_m = 2\,ln\left(\frac{x+1}{2}\right) + ln\left(\frac{x^2+1}{2}\right) - 2\arctan(x) + \frac{\pi}{2}\,, \tag{16}$$

$$\Psi_h = 2\,ln\left(\frac{x^2+1}{2}\right)\,,\ with\ \ x = (1-16z/L)^{1/4}\,. \tag{17}$$

For stable situations the following functions are used (Holtslag and De Bruin, 1988; Beljaars and Holtslag, 1991)

$$-\Psi_m = -\Psi_h = a\frac{z}{L} + b\left(\frac{z}{L}-\frac{c}{d}\right)\exp^{-d\,z/L} + \frac{bc}{d}\,, \tag{18}$$

with $a = 0.7$, $b = 0.75$, $c = 5$ and $d = 0.35$.

Literature on stability effects over the ocean is very limited and most studies use land-based observations over the ocean (e.g. Janssen and Komen, 1985; Smith, 1988). The general believe is that the land based stability functions should be applicable, because the conditions for Monin Obukhov similarity are generally better satisfied over the ocean than over land (i.e. $z_1/z_o >> 1$, homogeneous surface conditions). It is often argued that the stability corrections are small but the corrections depend very much on wind speed, which is illustrated in Fig. 1. Stability effects are small for strong winds, but can be substantial in unstable cases at low winds and very large in stable situations. However, stable surface layers are not a widespread feature of the marine boundary layer. Stable boundary layers over the ocean are limited to situations with warm air advection e.g. in a warm front or in coastal areas with warm land air advecting over cold water.

2.3 Horizontal wind near the surface

Horizontal wind near the surface appears linearly in equation (1)-(3) and is therefore an important part of the transfer law. In strong winds the air-sea difference of e.g. humidity tends to be smaller than at light winds (cold air outflow is an exception because advection compensates for the intense air-ocean exchange). In this way the wind modulates the way by which the SST field is transferred to the atmosphere. In principle there is no parametrization involved since wind at the lowest model level can be used i.e. $|\vec{V}_1| = (U_1^2 + V_1^2)^{1/2}$. However, at low winds, the air-ocean coupling becomes vey weak and the free convection limit is entirely determined by the stability corrections.

In an earlier ECMWF formulation, a free convection limit was present through stability functions $\Psi_{m,h}$ that scale as $1/|\vec{V}_1|$ for low winds (i.e. very unstable; see e.g. Louis, 1979). However, this formulation gave very weak air-sea transfer at low winds. The problem was that the bulk transfer law is the integral of the gradient function from the surface displacement height z_{om} to z_1 assuming $-z/L >> 1$ over the entire range. It turns out that there is a physical mechanism that maintains some surface wind in free convection, namely the large convective eddies with dimensions of the order of the boundary layer depth. These eddies

[1] A number of recent papers discuss alternatives to the iterative conversion from bulk Richardson numbers to Obukhov lengths e.g. Buyn, 1990; Launiainen, 1995; Mascart et al., 1995; Uno et al., 1995; Lo, 1996; Van Den Hurk and Holtslag, 1997.

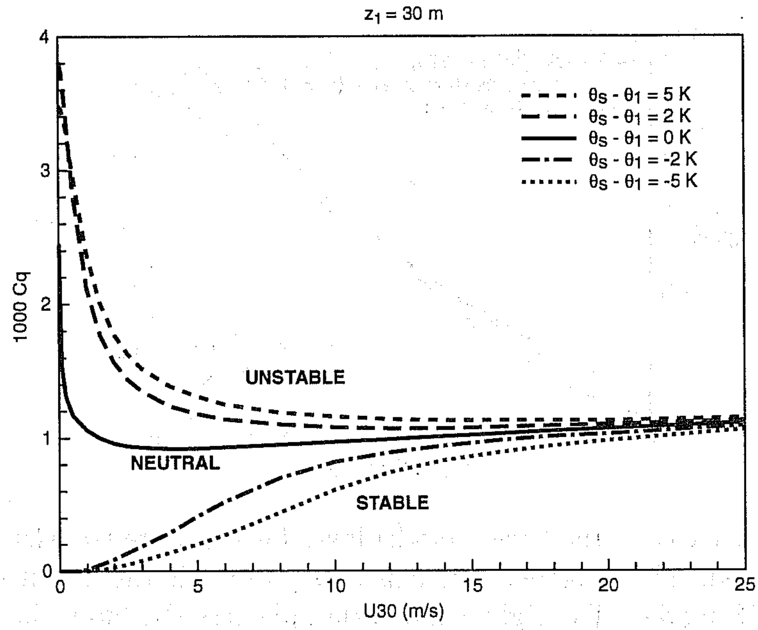


Figure 1: Ocean moisture transfer coefficients for the ECMWF surface layer with a reference height of 30 m as a function of wind speed and different potential temperature differences between the surface and the lowest model level at 30 m height.

are often called inactive because they contribute to horizontal velocity fluctuations, but have negligable vertical motion associated with them and therefore do not contribute to the vertical transport. However, these eddies maintain a finite horizontal wind often referred to as the gustiness contribution.

A natural way of handling the gustiness is by adding a free convection velocity scale to the resolved horizontal velocity (Liu et al., 1979; Schumann, 1988; Godfrey and Beljaars, 1991; Beljaars, 1995a, Fairall et al., 1996a):

$$|\vec{V}_1| = \ (U_1^2 + V_1^2 + \beta w_*^2)^{1/2} \ , \quad w_* = \ (z_i\, g/\theta_v\, \overline{w'\theta'_{vo}})^{1/3} \, , \qquad (19)$$

with z_i for boundary layer height. Observational and model studies suggest β values ranging from 0.8 to 1.25 (Beljaars, 1995a; Fairall et al., 1996a; Jabouille et al., 1996; Mahrt, 1996). Operationally, ECMWF uses $\beta = 1$. The choice of the boundary layer height is not very critical so a fixed value of $1000\,m$ is used in the ECMWF model. The dry convective contribution to gustiness is not necessarily the only contribution; any mechanism leading to subgrid variability in the surface wind should be added. Jabouille et al. (1996) propose a gustiness parameter related to precipitation from deep convection and Mahrt (1996) analyses observational material on meso-scale variability dependent on the size of the grid square. An alternative simple representation of free convection is proposed by Stull (1993), but different coefficients are needed for land and ocean.

2.4 Impact in the ECMWF model

The surface layer formulation as described above was an improvement with respect to the old formulation as documented by Louis et al. (1982). The main difference was in the moisture

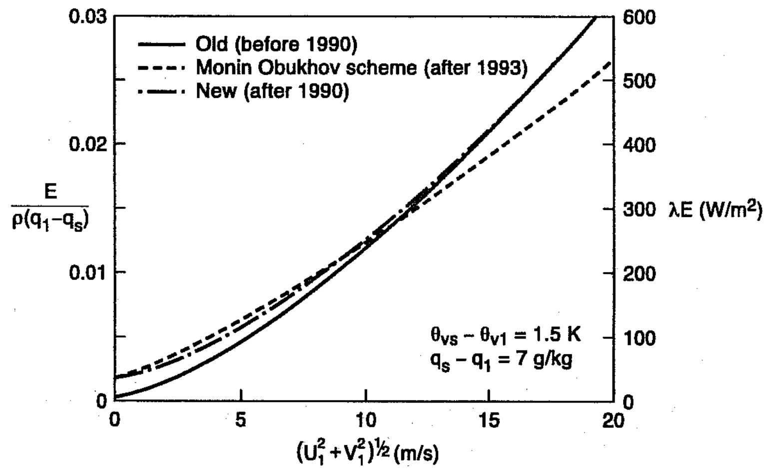


Figure 2: The conductivity of the lowest model layer for moisture transfer (left hand scale) as function of wind speed. The reference height is 30 m, which is the height of the lowest model level in the ECMWF model. The right hand scale indicates the latent heat flux for a typical temperature difference of 1.5 K and a specific humidity difference of 7 g/kg. The solid line represents the old scheme (as used by Louis et al., 1982), the dashed line represents the current scheme (operational after 1993) and the dash-dot line represents the empirical implementation of 1990 which limits the impact to low wind speeds only.

transfer at low as well as high wind speeds. The advantages of the new surface layer formulation over the ocean are: (i) it has a better free convection limit for evaporation which is highly relevant for the tropical circulation (Miller et al., 1992) and (ii) the ocean transfer coefficients for heat and moisture agree better with recent observations (Bradley et al., 1991; DeCosmo, 1991).

The impact of the improved free convection limit was substantial when it was implemented in 1990. The reason is that the tropical circulation is very sensitive to the SST's in the so-called warm pool of the Western Pacific where wind speeds tend to be low. Because winds are weak, the moisture jump over the surface layer is large and depends strongly on the specification of air-sea transfer coefficients in the surface layer. To illustrate the magnitude of the change, an example is given in Fig. 2. For a typical air-sea difference of 1.5 K and 7 g/kg, the latent heat flux is increased from 5 to 40 W/m^2 at zero wind speed. Initially (in 1990) the improved free convection limit was not implemented as described above, but in a more empirical way restricting the change to winds below 10 m/s (see Fig. 2). The impact was most noticeable in the tropical precipitation and the tropical wind errors. The old model had a dry zone over the warm pool in the Western Pacific and a tendency to produce a double Inter Tropical Convergence Zone (ITCZ). The improved scheme increased the precipitation over the Western Pacific and virtually eliminated the split ITCZ (Fig. 3). The increased latent heat release in the concentrated ITCZ enhanced the strength of the Hadley circulation and lead to reduced Easterly errors in the upper troposphere as shown in Fig. 4 (Miller et al. 1992). The impact on the tropical circulation can be interpreted in terms of Gill's linear model as a response to the latent heat release in the Western Pacific (Gill, 1982).

In the high wind speed regime, the change in the surface layer formulation was not directly beneficial to the ECMWF model as shown by Beljaars (1995b). The reduced air-sea interaction

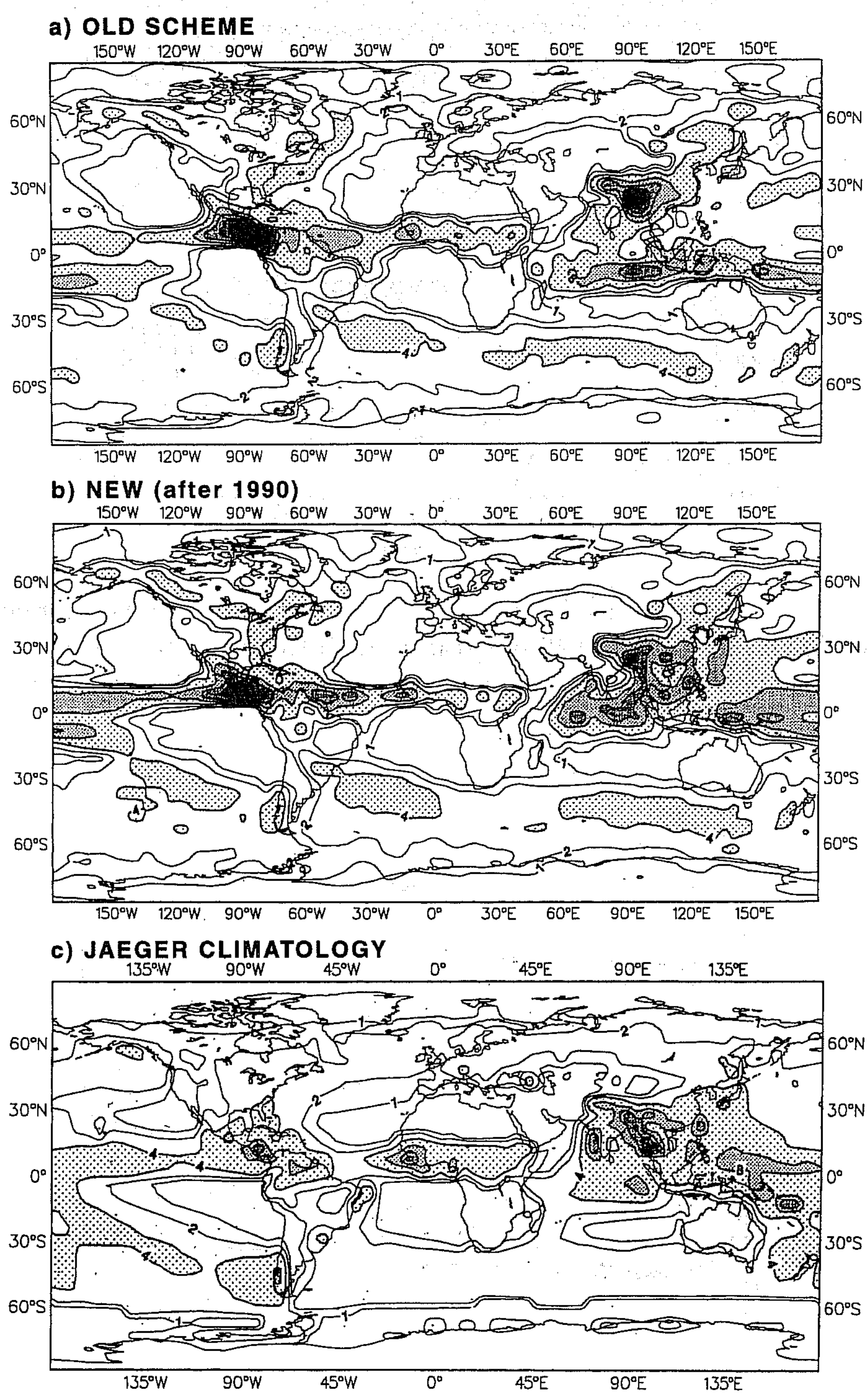


Figure 3: Rainfall averaged over 90 days of a T42, June/July/August model integration with the old scheme, and the new scheme (as introduced in 1990) in comparison with the climatology by Jaeger.

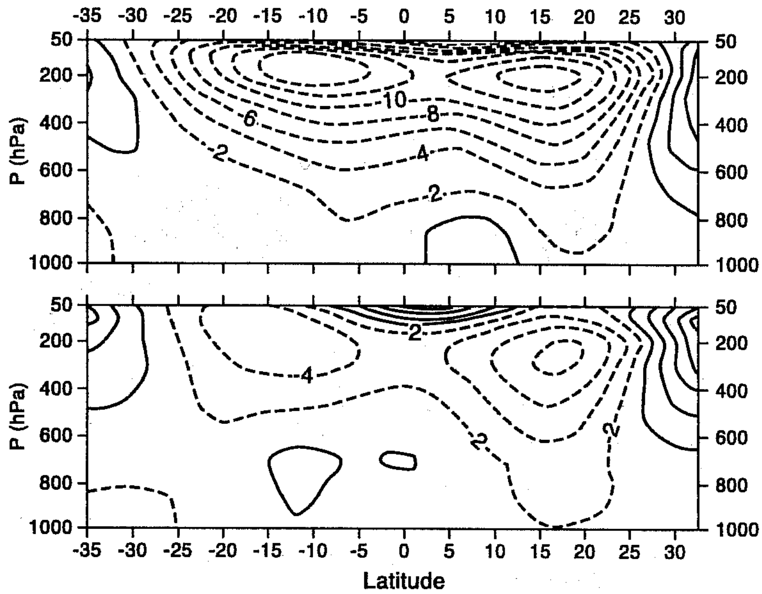


Figure 4: Zonal mean wind error (difference between model and analysis) averaged over 90 days for a December-January-February T42 integration with the old scheme (upper panel) and the new scheme with enhanced air-sea coupling at low winds as introduced in 1990 (lower panel).

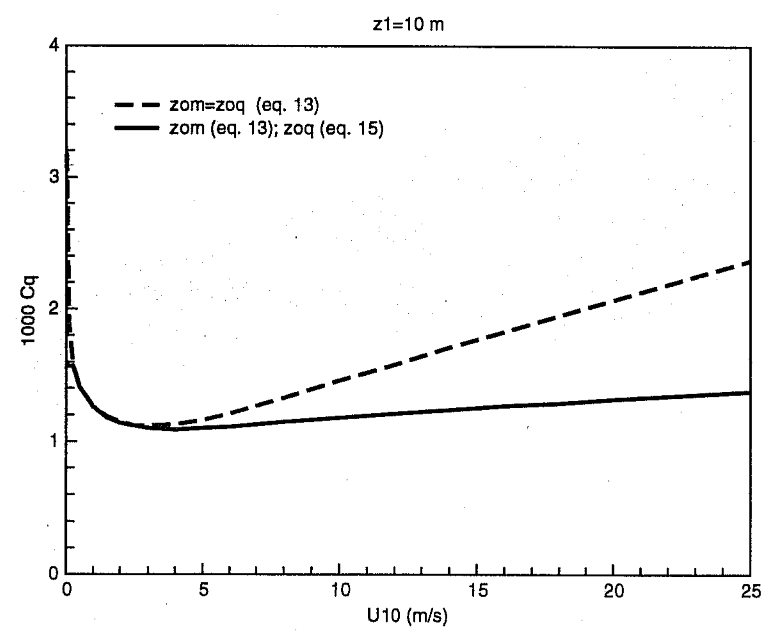


Figure 5: The neutral transfer coefficient for moisture as a function of wind speed. The solid line corresponds to z_{om} and z_{oq} equations according to (13) and (15); the dashed line has been computed from $z_{oq} = z_{om}$ according to (13).

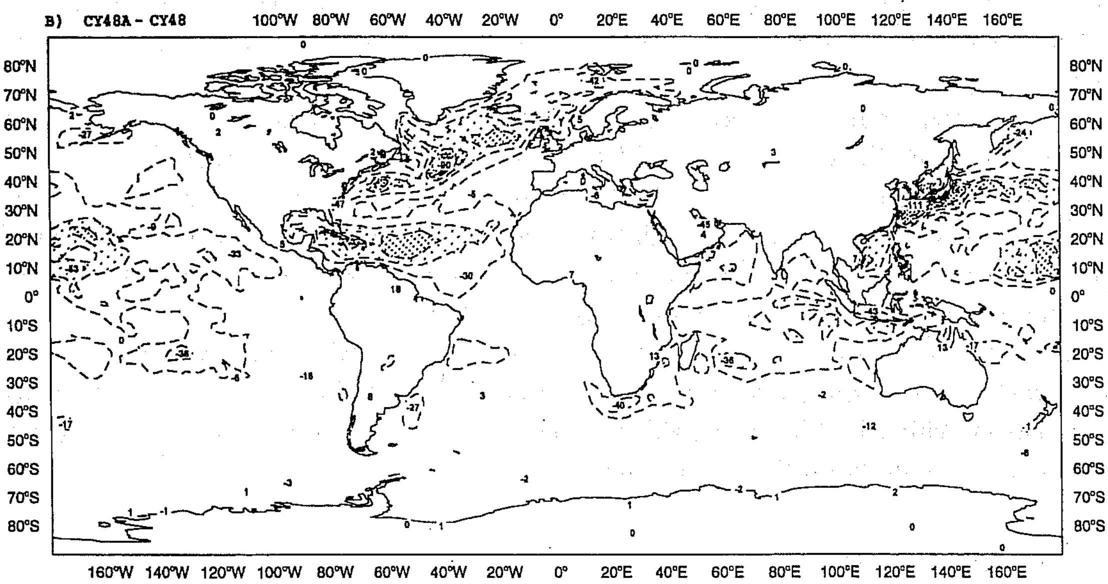


Figure 6: Difference in latent heat flux from the ocean between forecasts with the two moisture transfer formulations in figure 5. The 24 hour accumulated values are averaged over 21 forecasts initialized from their own analysis. The dashed contours indicate that the higher transfer coefficients lead to higher latent heat fluxes (contour interval: 20 W/m^2).

at high winds, as the result of much smaller sea surface roughness for heat and moisture than for momentum (introduced in 1993; see Fig. 5), was clearly in better agreement with observations (DeCosmo, 1991), but was detrimental in winter over the North Atlantic during strong cold air outflow (see Fig. 6 for the effect on latent heat flux). This was due to compensating errors; the mass flux scheme for convection did not extract enough moisture from the boundary layer, leading to an underestimation of evaporation which was compensated for by large transfer coefficients. A change in the closure of the convection scheme corrected this problem. This is a clear reminder that boundary layer turbulence is not an isolated process in a large scale model, but that it interacts with other processes as convection and cloud processes. It also implies that different schemes may perform differently in different models.

In conclusion, it can be stated that the surface layer formulation based on Monin Obukhov similarity functions is well established now in NWP. It works very well over homogeneous surfaces, provided that the boundary layer scale eddy motion is accounted for in the near surface wind and that proper surface roughness lengths are prescribed in particular over the ocean. The model sensitivity to air-sea transfer is large because the gradients near the surface are large.

3 Boundary conditions for T and q

In the operational ECMWF model, the sea surface temperature (SST) is specified from an analysis provided by NCEP (Reynolds, 1988) and kept constant during the 10 day forecast. This analysis is a blend of satellite retrievals and in situ observations from ships. The idea is to have a detailed horizontal distribution from satellite and to anchor this temperature fields to the rather sparse ship observations. It means that that the analyzed SST fields are calibrated as if they are ship observations and therefore they represent bulk SST fields i.e. measured a few meters deep.

It is well known that the ocean skin temperature is not always the same as the bulk SST. A very shallow layer (less then 1 mm thick) is cooler because of the turbulent and long wave radiative heat loss to the atmosphere which has to be compensated by the inefficient molecular transport in the water skin. Solar radiation has only a small effect on the cool skin because the solar absorption in such a thin layer is small. However, at low winds, solar radiation can create a so-called warm layer with a depth of a few meters.

Ocean skin effects are often ignored in atmospheric models. In the following subsections we will investigate the impact of simple parametrizations for the cool skin and the warm layer. Another aspect which is seldom considered in atmospheric models is the effect of salinity on the saturation specific humidity at the surface.

3.1 The cool skin

The cool ocean skin is the result of heat loss to the atmosphere which is balanced by thermal conduction in the quasi-laminar sublayer near the water surface (see Saunders, 1967 for a description). Observations, relying on radiative methods, show temperature differences across the cool skin of up to about 1 K (see e.g. Schlüssel et al., 1990; Soloviev and Vershinsky, 1982; Grassl, 1976; Coppin et al, 1991). Scaling arguments for the skin layer lead to the following expression for the temperature difference over the skin layer (cf. Fairall et al. 1996b)

$$\Delta T_c = \frac{Q\delta}{k_w}\ , with\ Q = R_{Tn} + H + \lambda E + f_c\, R_{Sn}\ , \tag{20}$$

where R_{Tn} and R_{Sn} are the net thermal and solar radiation at the ocean surface, H and λE are the sensible and latent heat flux (downward fluxes have are positive), k_w the thermal conductivity of water and δ the thickness of the quasi-laminar sublayer. Coefficient f_c is the fraction of the solar radiation that is absorbed in layer δ, (about 10%). The expression for δ is

$$\delta = \frac{\lambda_s \nu_w}{(\rho/\rho_w)^{1/2} u_*}\ , \tag{21}$$

where ν_w is the kinematic viscosity of water, ρ the density of air, ρ_w the density of water, and u_* the friction velocity. Coefficient λ_s is known as the Saunders constant, which applies to strong wind conditions (i.e. wind shear dominated conditions). Fairall et al. (1996b) recommend a value of 6 and suggest for finite winds a correction from convectively driven turbulence of the following form

$$\lambda_s = 6\left[1 + \left(\frac{2^4 Q g \alpha_w \rho_w C_p \nu_w^3}{u_*^4 (\rho/\rho_w)^2 k_w^2}\right)^{3/4}\right]^{-1/3}\ , \tag{22}$$

where g is the acceleration of gravity, α_w is the thermal expansion coefficient (decreasing from about $3\,10^{-4}\ K^{-1}$ in the tropics to near zero in polar regions.

During the TOGA-COARE experiment, Fairall et al. (1996b) observed cool skin effects in the range of 0.1 to 0.3 K in the warm pool of the Western Pacific with an amplitude of the diurnal cycle of about 0.1 K. These cool skin temperatures were well simulated by the model described above and therefore this formulation is used for sensitivity experiments (see subsection 3.4).

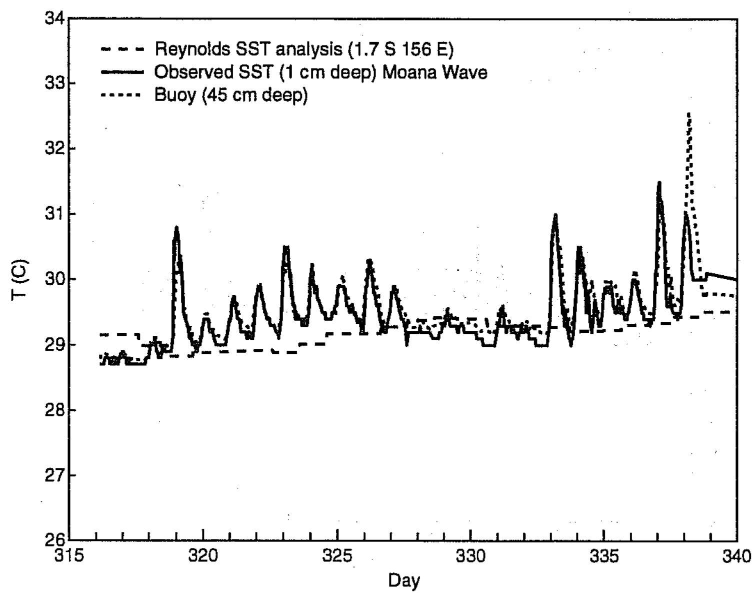


Figure 7: Time series of hourly observations during the first intensive measuring period of TOGA/COARE (from 11-11-1992 to 3-12-1992) of temperature 1 cm deep (solid), of temperature 45 cm deep (dot) and from the daily Reynolds SST analysis as used operationally by ECMWF (dash).

3.2 The warm layer

The near ocean warm layer is caused by solar absorption in the top few meters of the ocean during day time. This warm layer can develop when the wind mixing is not strong enough to prevent a stable layer to build up. The result is a diurnal cycle in the surface temperature which is commonly observed by satellite (Schlüssel et al., 1990), but not seen in routine bulk SST observations from ships. Buoys may see the diurnal cycle dependent on their measuring depth. Fig. 7 shows a time series of ocean temperature observations at a depth of 1 *cm* from R/V Moana Wave during TOGA/COARE and observations from a buoy with its sensor at a depth of 0.45 *m*. These high time resolution observations are compared with the daily SST analysis from NCEP (Reynolds, 1988) which should be interpreted as a bulk SST observation say a few meters deep. The 1 *cm* and 0.45 *m* observations show a distinct diurnal cycle with a maximum amplitude of a few degrees. On days with very low winds, the warm layer survives the night and a residual warming remains until the next day. It is obvious that a diagnostic model is not sufficient to model the warm layer and that a prognostic equation is needed. The warm layer is typically a few meters deep (Webster et al., 1996; Fairall et al., 1996b).

Diagnostic and prognostic schemes have been developed to simulate the warm layer effect by e.g. Price et al. (1986), Webster et al. (1996), and Fairall et al. (1996b). Diagnostic relations may be adequate for the amplitude of the diurnal cycle in climatological applications, but are less suitable for forecast models because also the history of the warming plays a role.

To illustrate the effects we use the empirical model by Webster et al. (1996) in a diagnostic and prognostic way. On the basis of TOGA/COARE observations, Webster et al. (1996) propose the following amplitude of the diurnal cycle for the temperature difference over the

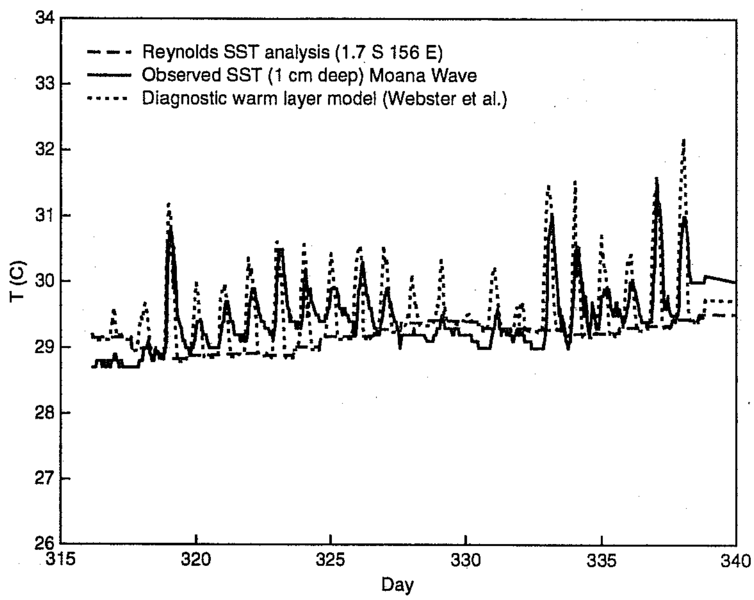


Figure 8: Time series of 1 cm deep observed (solid) and modelled (dot) SST's during TOGA/COARE (from 11-11-1992 to 3-12-1992). The Webster et al. (1996) diagnostic model is used with wind and radiation input from observations. The modelled warm layer effect is added to the Reynolds deep SST analysis (dash).

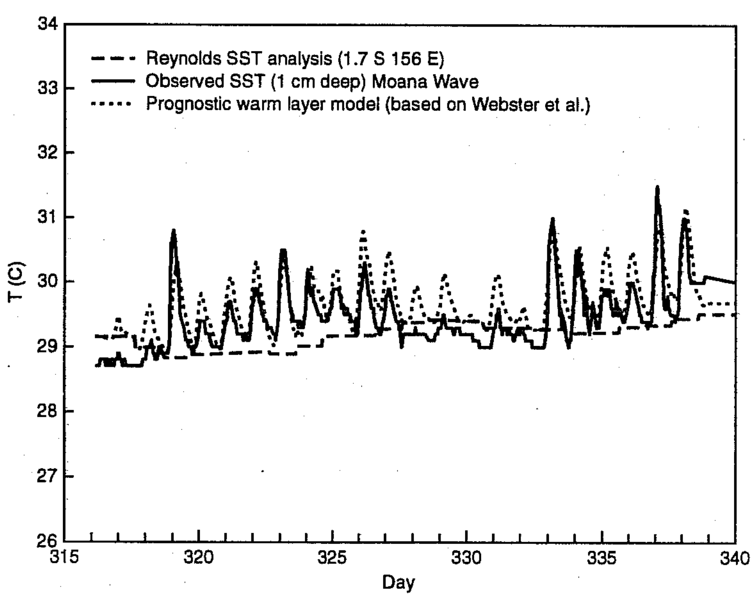


Figure 9: Time series of 1 cm deep observed (solid) and modelled (dot) SST's during TOGA/COARE (from 11-11-1992 to 3-12-1992). The prognostic warm layer model is used based on the empirical Webster et al. relations with wind and radiation input from observations. The modelled warm layer effect is added to the Reynolds deep SST analysis (dash).

warm layer

$$\Delta T_w = f + e\,|\vec{V}| + c\,ln|\vec{V}| + \{a + d\,ln|\vec{V}|\}S^p_{Sd} + b\,Pr\,, \tag{23}$$

where $|\vec{V}|$ is wind speed (in m/s), R^p_{Sd} is the peak downward solar radiation (in W/m^2), and Pr is precipitation rate (in $mm/hour$). The empirical constants are $a = 0.00228$, $b = 0.013$, $c = 0.06$, $d = -0.000275$, $e = -0.095$, and $f = 0.112$ for $|\vec{V}| < 2\,m/s$ and $a = 0.00262$, $b = 0.0223$, $c = -0.725$, $d = -0.00107$, $e = 0.139$, and $f = -0.036$ for $|\vec{V}| > 2\,m/s$. [2]

When the effect of precipitation is ignored, equation (23) can be interpreted as the solution of a prognostic equation of the form

$$\rho_w C_w \frac{d\Delta T_w}{dt} = \frac{R_{Sd}}{\Delta z} + \frac{f + e\,|\vec{V}| + c\,ln|\vec{V}| - \Delta T_w}{\Delta z\{a + d\,ln|\vec{V}|\}}\,, \tag{24}$$

where ΔZ is an additional empirical parameter representing the depth of the warm layer. The first term is the heating by solar radiation which warms the layer of depth ΔZ, and the second term represents the wind mixing which reduces the temperature of the warm layer. The equilibrium ΔT_w is determined by $e\,|\vec{V}| + c\,ln|\vec{V}|$, which is the cool skin part of the Webster et al. formulation. We set c and e to zero to isolate the effect of the solar heating only.

Figs. 8 and 9 show the results of simulations with the diagnostic (equation (23)) and prognostic (equation (24)) model respectively for the TOGA/COARE observations from R/V Moana Wave. The observed solar radiation and wind have been used as input to the model with a depth scale Δz of 2 m in equation (24). It is clear that the diagnostic model can simulate the diurnal cycle which is introduced by the solar radiation, but that a prognostic model is needed to keep the residual warm layer in low wind conditions.

3.3 Salinity effect on q_s

Most models use the saturation specific humidity at SST as boundary condition for humidity at the ocean surface. However, salinity reduces the saturation value and a reasonable approximation for a salinity of 34 parts per thousand is (Sverdrup et al., 1942)

$$q_s = 0.98\,q_{sat}(T_s) \tag{25}$$

The 2% difference due to salinity may look a small effect, but it should be seen as a fraction of the air-sea specific humidity difference, which is typically 15% in relative humidity. So a 2% change in saturation value at the surface is equivalent to a change of 2/0.15=13% in air-sea transfer (see Zeng et al. 1998 for an intercomparison of schemes).

3.4 Impact of boundary conditions for T and q

To test the impact of the cool skin, the warm layer and the salinity effect on q_s, 120-day long integrations have been performed at T63 resolution. The model runs start at 1-11-1992 with sea surface temperatures evolving in time according to the Reynolds analysis. The cool skin effect ΔT_c and warm layer effect ΔT_w are added to the bulk SST provided by the analysis before using them as a boundary condition for the atmospheric model. Four model versions are used: (i) Control model with the bulk SST's as boundary condition, (ii) Cool skin model

[2] These constants were taken from a draft version of the Webster et al. paper and are slightly different from the ones in the final paper.

as described in section 3.1, (iii) Warm layer model as described in section 3.2, and (iv) Model with salinity effect on q_s as in section 3.3.

Fig. 10 shows the mean impact of the different effects on the surface SST. The cool skin parametrization cools the sea surface mainly dependent on latitude; the polar areas show little temperature drop, whereas the tropical areas show a cooling of about 0.5 K (see Fig. 10a). The prognostic warm layer model produces a small mean warming in small areas in the tropics (see Fig. 10b). The maximum of about 0.5 K is in the warm pool of the western Pacific related to the low wind regime. The RMS of the warm layer effect, as shown in Fig. 10c, has also a maximum in the Western Pacific (between 0.4 and 0.6 K); a large area south of the equator has an RMS between 0.2 and 0.4 K due to the diurnal temperature cycle related to solar heating. That a mean effect is seen in the western Pacific is because the day time warming is not completely destroyed by wind mixing during the night and therefore a residual warm layer remains as long as the wind speed is low.

The salinity effect has no impact on the SST, but in order to illustrate the effect, an SST change can be defined that gives the same q_s change at the ocean surface. The equivalent SST change is distributed uniformly and varies from about .27 K at 0^oC to about .35 K at 30^oC, so the salinity effect looks like a general decrease of SST and does not introduce any localized anomaly.

Inspection of precipitation charts (not shown) averaged over Dec/Jan/Feb show very little impact from the three parametrizations described above. The impact on zonal wind in the tropics is also small (see differences for zonal mean wind in Fig 11), although a 3 month average over one experiment may not be sufficient to draw firm conclusions. The tropical upper troposphere shows a negative wind difference due to the cool skin, a positive wind difference due to the warm layer and a negative difference due to the salinity effect. For all three parametrizations the impact is no more than 2 m/s, which is smaller than what has been seen in 1990 with the introduction of the improved low wind speed air-sea tranfer parametrization (cf. Fig. 4). Although the model climate shows little sensitivity, the ocean temperature effects are not small particularly the diurnal cycle. This may be of importance for the use of satellite data over the ocean which needs to be investigated with help of data assimilation experiments.

4 Conclusions

Experience with the ECMWF model has shown that the parametrization of air-sea transfer is an important and sensitive aspect of the model. The reason is that the wind, temperature and moisture differences over the surface layer are large and therefore small changes in air-sea transfer lead to differences in the way the atmosphere feels the sea surface boundary condition. Changes in the transfer coefficients for moisture at low as well as high wind speeds had considerable impact on the ECMWF model climate, on its response to SST anomalies and on model performance (Palmer et al., 1992; Beljaars, 1995b).

Studies based on the TOGA/COARE experiment have shown that the current ECMWF air-sea transfer scheme is in reasonable agreement with observations (Zeng et al. 1998; Chang and Grossman, 1998). However, the surface boundary conditions for temperature and moisture cannot simply be taken from the bulk SST's as provided by the Reynolds analysis. The ocean has a very thin cool skin due to turbulent and radiative cooling which balances molecular heat transport in a quasi-laminar layer and during day time a warm layer can form as the result of solar heating. These two effects introduce mean temperature departures from the bulk SST of

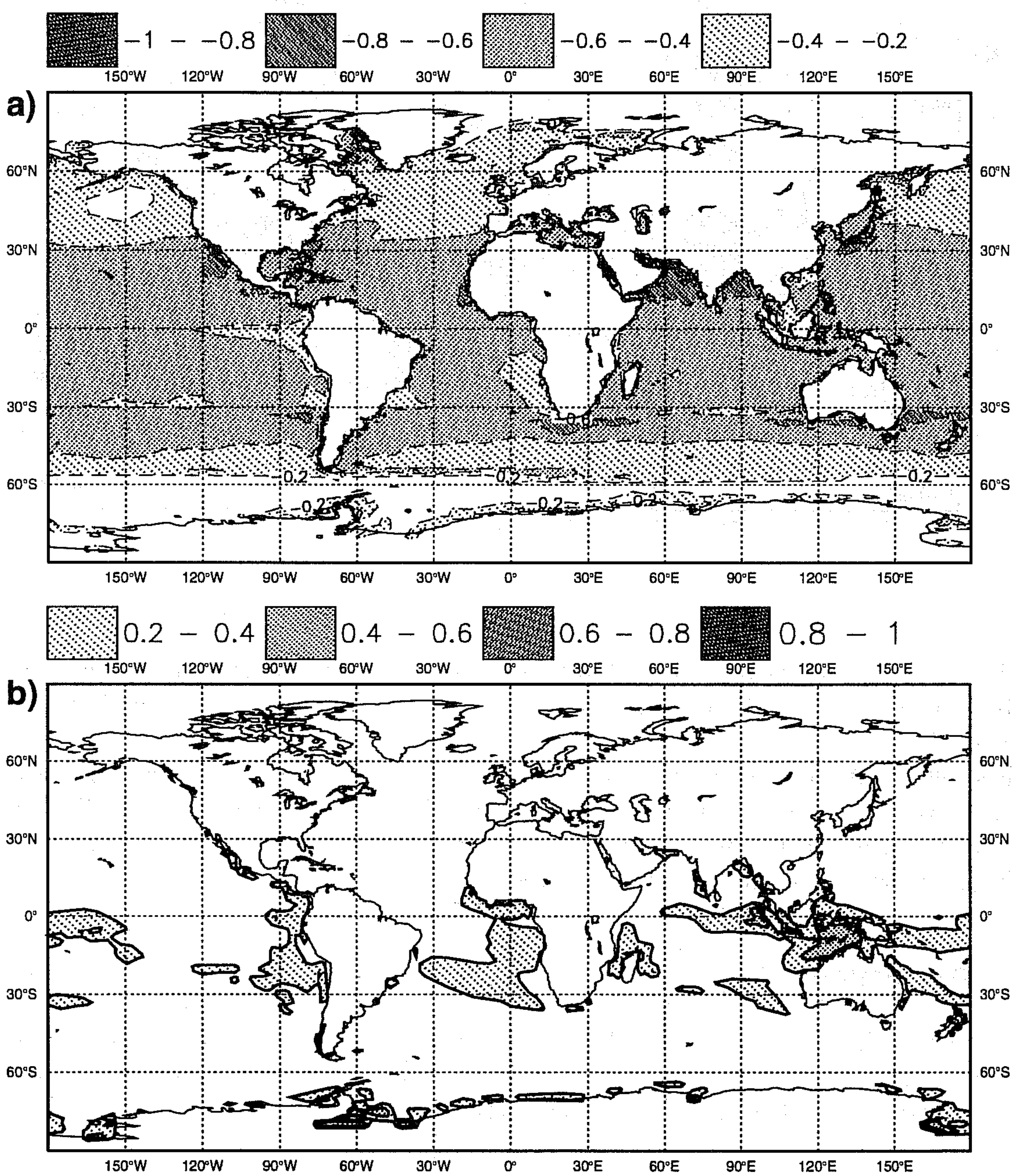
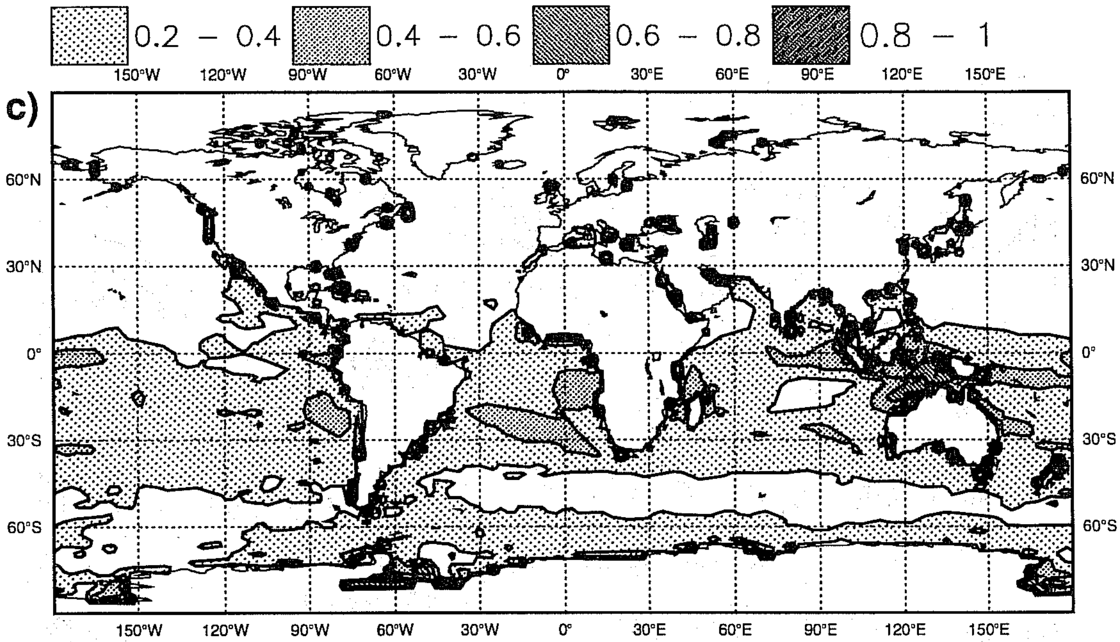


Figure 10: Temperature effects over the ocean due to different parametrizations in a 120 day integration for Dec-Jan-Feb. (a) mean cooling due to cool skin parametrization, (b) mean warming due to the warm layer parametrization, and (c) RMS of the warm layer effect.



up half a degree with a peak to peak diurnal cycle of up to a few degrees (Webster et al., 1996; Fairall et al., 1996b). Furthermore salinity reduces the saturation specific humidity at the sea surface by 2%.

These effects are non-negligable and should be incorporated in atmospheric models in spite of the relatively small impact on the model climate. The importance of a more realistic radiative surface temperature including a diurnal cycle must be relevant for satellite retrievals that rely heavily on knowledge of the sea surface temperature.

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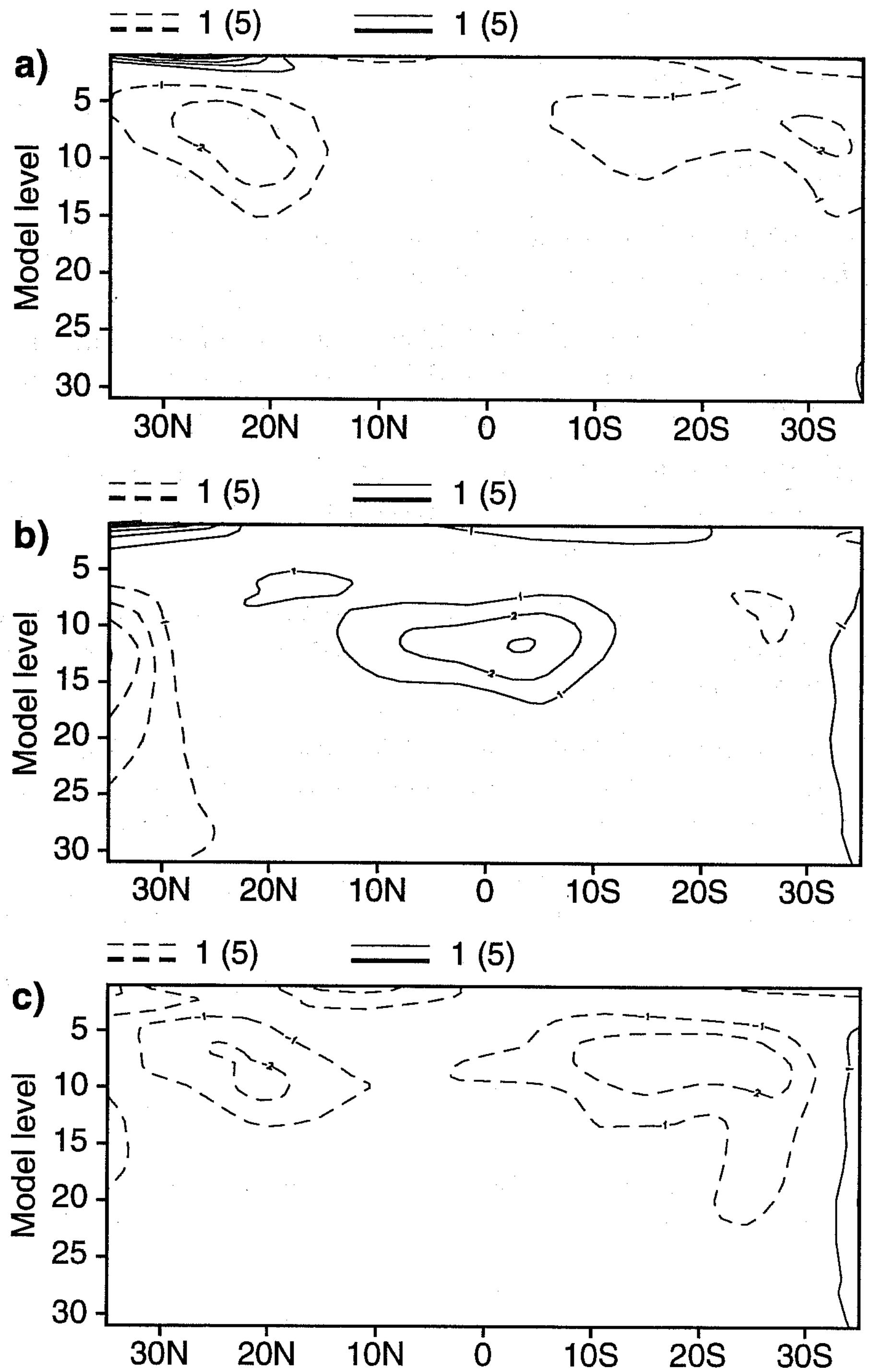

Figure 11: Impact on the zonal wind of different parametrizations in a 120 day integration for Dec-Jan-Feb. (a) effect of cool skin parametrization, (b) effect of the warm layer parametrization, (c) effect of using $q = 0.98\, q_{sat}$ instead of $q = q_{sat}$.

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