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Influence of oceanic whitecaps on the global radiation budget

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Abstract. Oceanic whitecaps may exert a cooling influence on the planet by increasing surface albedo. The direct, globally averaged radiative forcing due to whitecaps lies in the range 0-0.14 Wm^{-2} with a probable value of 0.03 Wm^{-2}. Though small, this global value is not negligible compared with the forcing due to some greenhouse gases and anthropogenic aerosols since pre-industrial times. The relative importance of whitecaps may be greater on regional and seasonal scales, with radiative forcing values reaching 0.7 Wm^{-2} in the Indian Ocean during summer. Whitecap effects on surface albedo should be taken into account explicitly in the numerical modeling and analysis of climate change.

Introduction

Two decades ago *Gordon and Jacobs* [1977] argued convincingly that whitecaps might affect the albedo of the ocean-atmosphere system and, therefore, the global radiation balance. Using radiation-transfer calculations they demonstrated that for totally reflecting whitecaps a wind speed increase from 6 to 14 ms^{-1} would double the local planetary albedo. The theoretical study followed observations by *Maul and Gordon* [1975] of increased LANDSAT-1 Thematic Mapper radiance, which they explained by the presence of whitecaps on the surface. This increased reflection constitutes a direct climate forcing, although different in nature from the forcing due to greenhouse gas and aerosol emissions which have increased steadily since pre-industrial times. One expects also, in any climate change scenario, that whitecap coverage, and thus the reflectivity of the surface, would "respond" to dynamical effects of the atmosphere.

Despite the theoretical arguments and suggestive observational evidence, whitecaps have been largely ignored --and not parameterized explicitly-- in climate models. This is attributable to the fact that whitecaps are less reflective than previously thought, with a reflectance generally not exceeding 55% in the visible and decreasing substantially with wavelength in the near infrared (*Whitlock et al.*, 1982; *Stabeno and Monahan*, 1986; *Frouin et al.*, 1996]. *Koepke* [1984] showed that the effective reflectance of whitecaps is only about 22% in the visible, due to changes in optical properties with age, and nearly independent of wind speed. He stated that "the optical influence of oceanic whitecaps can be assumed to be much less important than was formerly supposed." Earlier, *Payne* [1972] reported that whitecap effects on surface albedo are "not noticeable at wind speeds up to 30 kt." Consequently, the albedo of the ocean surface (and its parameterization) has not been perceived as a significant issue by climate modelers.

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In climate models, such as the National Center for Atmospheric Research community climate model [*Kiehl et al.*, 1998], the albedo of the ocean surface is usually parameterized using major sources of observed values such as *Payne* [1972] or *Kontratyev* [1969]. The data cover a broad range of sky and sea conditions, but they are not complete. *Payne*'s data, for example, were collected at a single location. Various fits to the data have been proposed, attempting to capture the effects of key variables, namely sun zenith angle and ratio of diffuse and direct incident sunlight. The formulas, however, do not separate the processes of backscattering by the water body, Fresnel reflection, and wave breaking, as done in *Koepke* [1984]; they only model average effects. This is not entirely satisfactory, even though all the processes may be implicitly taken into account, because the individual effects vary differently with sun zenith angle and light distribution. Hansen et al. (1997) included whitecaps in their climate model, but they did not attempt to isolate their influence.

Radiative Forcing

To quantify the effect of oceanic whitecaps on the global radiation budget, we adopt the approach used by *Charlson et al.* [1992] for sulfate aerosols, and *Penner et al.* [1992] for biomass-burning aerosols. At any location over the oceans the change in outgoing radiative flux due to whitecaps is

$$\Delta F = F_0\mu_0(1-A_c)T^u{}_aT^d{}_a\Delta R_s \qquad (1)$$

where F_0 is the extraterrestrial broad-band solar irradiance, μ_0 is the cosine of the sun zenith angle, A_c is the fraction of the surface covered by clouds, $T^u{}_a$ and $T^d{}_a$ are the clear-sky atmospheric transmissivities for up-welling and down-welling flux, respectively, and R_s is the albedo of the ocean surface. As a first approximation, it is assumed that the forcing occurs only in cloud-free regions. Atmosphere-surface interactions are neglected.

An estimate of ΔR_s can be obtained by modeling R_s as a sum of contributions due to diffuse reflection by whitecaps and the rest, i.e. Fresnel reflection by the whitecap-free surface and diffuse reflection by the water body. If f_{wc} denotes the fraction of the surface covered by whitecaps, R_s is expressed as [*Gordon and Jacobs*, 1977]

$$R_s = f_{wc}R_{wc} + (1-f_{wc})R_w \qquad (2)$$

where R_{wc} is the albedo of whitecaps and R_w the albedo of the whitecap-free surface/water system. The possibility of whitecaps interacting with the adjacent surface is ignored, and it is assumed that up-welling light below the whitecaps does not reach the surface. Thus the change in surface albedo due to whitecaps is

$$\Delta R_s = f_{wc}(R_{wc} - R_w) \qquad (3)$$

This equation, which assumes that whitecaps are opaque to under light, probably underestimates ΔR_s. Using *Koepke*'s formulation for R_w, $\Delta R_s = f_{wc}[R_{wc} - (R_g + R_{wc}R_u)] > f_{wc}[R_{wc} - (R_g + R_u)]$, where R_g and R_u are respectively the surface reflection and water back-scattering components of R_w. The difference between $(R_g + R_u)$ and $(R_g + R_{wc}R_u)$, negligible at large sun zenith angles ($R_g >> R_u$), is typically 0.01 at small sun zenith angles.

The fractional coverage of whitecaps, f_{wc}, is governed by surface wind speed, but also depends on fetch, water temperature, and thermal stability of the atmosphere. Several statistical studies provide f_{wc} as a function of wind speed, wind speed and air-sea temperature difference, or wind speed and spectral peak frequency of breaking waves. A parameterization of f_{wc} as a function of only wind speed has a large uncertainty, but is convenient. We use the optimal power-law formula obtained *by Monahan and O'Muircheartaigh* [1980]:

$$f_{wc} \approx 2.95\ 10^{-6}\ W^{3.52} \pm 6.\ 75\ 10^{-3} \tag{4}$$

where W is wind speed at 10 m elevation in meters per second. The relative uncertainty on f_{wc} is about 70% for a wind speed of 10 ms^{-1}.

Equation (4) was obtained from statistical fit analyses of warm-water whitecap data sets. Using additional data sets collected in waters of intermediate and cold temperatures and taking into account sea-air temperature difference, ΔT, *Monahan and O'Muircheartaigh* [1986] proposed a new formula: $f_{wc} = 1.95\ 10^{-5}\ W^{2.55} \exp(0.0861 \Delta T)$. For the average wind speed of 9 ms^{-1} used later in the study and a ΔT of 1° C (in general ΔT is positive on monthly and annual time scales), the f_{wc} value predicted by the new formula is 6% below the value obtained using Equation (4).

Following *Koepke* [1984] the spectral albedo of whitecaps is expressed as the product of an effective albedo, R_{eff}, and a spectral factor, $f(\lambda)$. For the effective albedo we use *Koepke*'s results for combined patches and streaks (account for the thinning of whitecaps with time), i.e., $R_{eff} = 0.22 \pm 0.11$. For the spectral factor we use a statistical fit based on available experimental data [*Frouin et al.*, 1996; *Fougnie*, 1998], i.e., $f(\lambda) = \exp[-p(\lambda - 0.6)^q]$ for _ >0.6µm and $f(\lambda) = 1$ for $\lambda < 0.6$ µm, with $p = 1.75 \pm 0.48$ and $q = 0.99 \pm 0.05$. This gives 0.51 for f at 1 µm, compared with 0.9 according to *Whitlock et al.* [1982]. The broad band albedo, R_{wc}, is deduced by multiplying the spectral albedo by the extra-terrestrial spectral solar flux, $F_0(\lambda)$, and integrating over the entire solar spectrum. That is,

$$R_{wc} = R_{eff} \int F_0(\lambda) f(\lambda) d\lambda / \int F_0(\lambda) d\lambda \approx 0.16 \pm 0.09 \tag{5}$$

For the albedo of the whitecap-free ocean surface we use the model *of Briegleb and Ramanathan* [1982], obtained by fitting *Kondratyev* [1969]'s and *Payne* [1972]'s data. Surface albedo is calculated as the sum of two weighted components, one for direct incident solar flux (depends on the sun zenith angle) and the other for diffuse incident solar flux (a constant). Direct radiation generally dominates in clear-sky conditions, where the effect of whitecaps is expected to occur. Diffuse radiation becomes important only at high sun zenith angles, but these angles contribute very little to the globally averaged albedo. Furthermore, *Payne* [1972]'s clear-sky measurements were used to parameterize the direct component. Therefore the diffuse component is neglected, leading to

$$R_w \approx 0.05/(1.1\ \mu_0^{1.4} + 0.15) \tag{6}$$

Note that *Taylor et al.* [1996] obtained $R_w = 0.037/(1.1\ \mu_0^{1.4} + 0.15)$ in cloudless skies from aircraft observations over several years at a variety of locations. Using their formula instead of Equation (6) would give smaller R_w values, for example 0.03 instead of 0.04 for a sun at zenith, and therefore larger ΔR_s estimates.

Using Equations (3) to (6) the globally averaged perturbation in reflected solar flux due to the presence of whitecaps can now be evaluated:

$$\begin{aligned} \langle \Delta F \rangle &\approx 0.5aF_0(1 - \langle A_c \rangle) \int T^u{}_a T^d{}_a \Delta R_s \mu_0 d\mu_0 \\ &\approx 1.32\mathrm{x}10^{-3}(1 - \langle A_c \rangle) \langle W' \rangle^{3.52} \int T^u{}_a T^d{}_a (R_{wc} - R_w) \mu_0 d\mu_0 \end{aligned} \tag{7}$$

where a is the fraction of ice-free (annually averaged) oceanic surface ($a = 0.65$), $\langle\rangle$ denotes globally averaged values over the oceans (simple area average), $F_0 = 1372$ Wm^{-2}, and the integral limits are 0 and 1. To account for non-linearity in the relation between f_{wc} and W, a globally averaged effective wind speed, $\langle W' \rangle$, is introduced, defined such as $\langle W' \rangle^{3.52} = \langle W^{3.52} \rangle$. Correlation between the various variables, in particular A_c and W, are neglected, as well as implicit whitecap effects in the expression of R_w.

To estimate the integral on the right-hand side of Equation (7) $T^u{}_a$ and $T^d{}_a$ are computed as a function of μ_0 using the SUNRAY radiation-transfer model [*Fouquart and Bonnel*, 1980]. The model is run with vertical profiles of temperature, ozone, and moisture from five atmospheric types (tropical, mid-latitude summer and winter, sub-arctic summer and winter). Background aerosols are included, and their radiative properties and vertical mass loading are those of the maritime (MAR-I) model of the World Climate Research Programme. The aerosol single scattering albedo, asymmetry parameter, and optical thickness are respectively 0.99, 0.75, and 0.09 at the wavelength of 0.55 µm. The $T^u{}_a$ and $T^d{}_a$ values obtained for each atmospheric type are weight-averaged using the respective fractions of tropical, mid-latitude, and sub-arctic oceans. They are also weighted by the fractions of summer and winter months during the year, assumed to be equal to 0.5 in both mid-latitude and sub-arctic regions. This gives

$$\begin{aligned} \int T^u{}_a T^d{}_a (R_{wc} - R_w)\mu_0 d\mu_0 &= \overline{T^u{}_a T^d{}_a} \int (R_{wc} - R_w)\mu_0 d\mu_0 \\ &\approx 0.65 \int [0.16 - 0.05/(1.1\mu_0^{1.4} + 0.15)]\mu_0 d\mu_0 \approx 2.7\mathrm{x}10^{-2} \end{aligned} \tag{8}$$

where the average product of $T^u{}_a$ and $T^d{}_a$ is made explicit so that its value, 0.65, can be compared with values used in other studies, e.g., 0.58 in *Charlson et al.* [1992] and 0.66 in *Iacobellis et al.* [1999].

The globally averaged fractional cloud coverage, $\langle A_c \rangle$, is obtained from 7 years (1983-1989) of monthly International Satellite Cloud Climatology (ISCCP) C2 data [*ISCCP*, 1992]. Only oceanic regions are considered in the average. This yields $\langle A_c \rangle \approx 0.59$. The effective wind speed, $\langle W' \rangle$, is obtained by averaging 10 years (1980-1989) of monthly oceanic wind data from European Centre Medium Range Forecast (ECMWF) analysis, which gives $\langle W' \rangle = 9.0$ ms^{-1}. Inter-annual variability is small over the 10-year period, with $\langle W' \rangle$ values ranging from 8.8 to 9.2 ms^{-1}.

Using these estimates of $\langle A_c \rangle$ and $\langle W' \rangle$, the value of 2.7x10^{-2} for the integral on the right-hand side of Equation (7), and taking into account uncertainties on R_{wc} and f_{wc} we find for $\langle \Delta F \rangle$ a probable value of 0.03 Wm^{-2} and a possible range of 0 to 0.14 Wm^{-2}. A similar probable $\langle \Delta F \rangle$ value would have been obtained by using *Koepke* [1984]'s model for R_s, *Monahan and*

O'Muircheartaigh [1986]'s formula for f_{wc}, and *Taylor et al.* [1996]'s formula for R_w, because of compensating effects.

Discussion

Some amount of whitecap forcing does occur in cloudy conditions because clouds are not completely opaque to incident sunlight. In the case of anthropogenic aerosols, the ratio of forcing in cloudy and clear-sky conditions was found to be significant, with a value of 0.25 for sulfate-type [*Boucher and Anderson*, 1992] and of 0.39 for carbonaceous-type [*Iacobellis et al.*, 1999]. This is equivalent to reducing fractional cloud coverage, i.e., multiplying $<A_c>$ by a factor of 0.75 and 0.61, respectively, in models assuming that the forcing occurs only in cloud-free regions. However aerosols, unlike whitecaps, may be present within and above the cloud layer. *Boucher and Anderson* [1992] suggested that the aerosols above the cloud layer are mostly responsible for the effect in cloudy regions. *Iacobellis et al.* [1999] showed that for carbonaceous aerosols the higher effect in cloudy conditions is due primarily to the latitudinal and seasonal dependence of the aerosols.

To examine this closer, the whitecap forcing in the presence of clouds is estimated by changing $<A_c>$ in Equation (7) into $<A_c'>$ as follows

$$<A_c'> = (1 - \overline{T^u_c T^d_c}) <A_c> \tag{9}$$

with

$$\overline{T^u_c T^d_c} = \int T^u_c T^d_c T^u_a T^d_a (R_{wc} - R_w) \mu_0 d\mu_0 \Big/ \int T^u_a T^d_a (R_{wc} - R_w) \mu_0 d\mu_0 \tag{10}$$

where T^u_c and T^d_c are cloud transmissivities for up-welling and down-welling flux, respectively, and the integration limits are 0 and 1. These functions are computed as a function of μ_0 by running the SUNRAY model with average values of cloud optical thickness in the model's two spectral intervals. The average value in the short-wavelength interval is obtained by area averaging (oceans only) seven years of ISCCP C2 monthly cloud optical thickness retrievals, made at approximately 0.6 µm. The average value in the long-wavelength interval is obtained using equations derived in Stephens *et al.* [1984]. This gives 6.5 and 7.5 for the cloud optical thickness in the model's short- and long-wavelength intervals, yielding $\overline{T^u_c T^d_c} \approx 0.34$ and $<A_c'> \approx 0.66<A_c>$. Since $<A_c> \approx 0.59$, the whitecap forcing in cloudy regions may increase our $<\Delta F>$ estimate by about 50%. This value is conservative, because $\overline{T^u_c T^d_c}$ is underestimated by simply averaging ISCCP cloud optical thickness retrievals, due to the non-linear relationship between cloud optical thickness and transmissivity. In regions of high winds where whitecaps occur most, however, the actual cloud layer may be thicker optically than the cloud layer used in the SUNRAY model, mitigating to some extent the estimated increase.

On monthly or longer time scales wind speed is correlated positively to cloudiness, although in up-welling regions, where the lower atmosphere is stabilized, strong and sustained winds may exist under clear skies. Since whitecap forcing is relatively small in the presence of clouds, $<\Delta F>$ might be overestimated by neglecting correlations between the controlling variables. Three-dimensional simulations of climate forcing by sulfate aerosols [*Boucher and Anderson*, 1992; *Kiehl and Briegleb*, 1993; *Taylor and Penner*, 1994; *Haywood et al.*, *1997*] revealed values about twice lower compared with box-model estimates. The discrepancies, however, might be due to differences not only in the input data sets, but also in the radiation-transfer models. *Iacobellis et al.* [1999], on the other hand, found that the climate forcing by carbonaceous aerosols was reduced by only 13% when taking into account the spatial and temporal distribution of the controlling variables. For whitecaps, it is important to note that their fractional coverage increases steeply with wind speed (see Equation 4). In the quasi-stationary low-pressure regions and the southern oceans, where winds are strong, fractional cloud coverage may be higher than in the tropics by a few tenths (i.e. 0.7-0.8 instead of 0.5), but the whitecap fractional coverage may be higher by a factor of 5 to 10. Thus the effect of increased wind speed would largely compensate the effect of increased cloudiness and could even be dominant, leading to an augmentation, not a reduction, of the $<\Delta F>$ estimate.

Our estimate of the global, annual-mean radiative forcing by oceanic whitecaps is small, yet not negligible compared with the direct forcing by some greenhouse gases and even anthropogenic aerosols. In magnitude, the probable value of 0.03 Wm^{-2} is about 20% of the value for nitrous oxide, 30 to 50% of the values for halo-carbons, and 8%, 15%, and 30% of the values for sulfate, biomass-burning, and soot aerosols, respectively [*IPCC*, 1996]. But it is only a few percent of the value for carbon dioxide, the major contributor to the direct greenhouse effect.

In contrast to well-mixed greenhouse gases whitecaps are more localized, and their forcing may be much larger on regional and seasonal scales, with definite effects on climate. An estimate

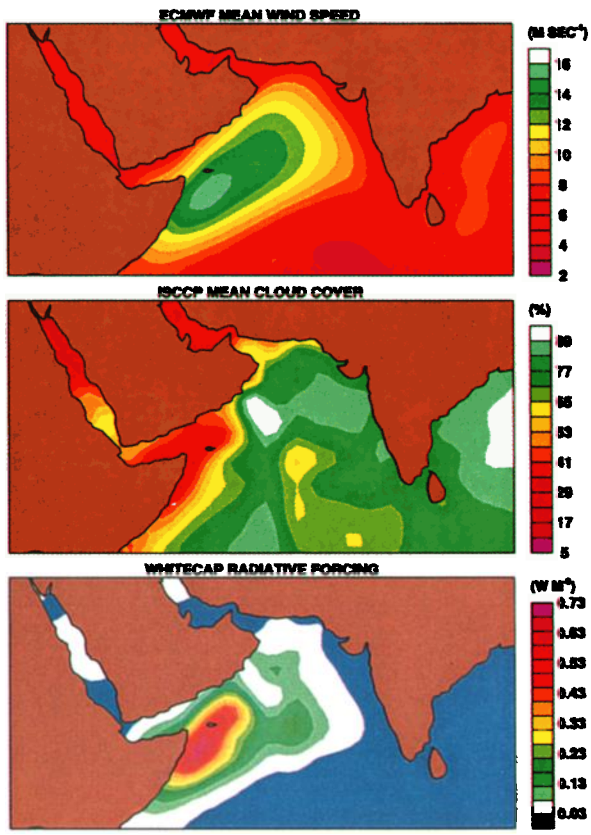

Figure 1. July mean surface wind speed (top), cloud coverage (middle), and whitecap radiative forcing (bottom) in the Northern Indian Ocean. Values of ΔF reaching 0.7 W m^{-2} occur in the Western Arabian Sea due to strong winds and relatively low cloud amounts.

of the monthly mean whitecap forcing on regional scales can be obtained by combining Equations (1) and (3)-(6) and integrating over time:

$$\Delta F \approx 2.95\ 10^{-6}\ F_0 [1 - A_c(x,y,m)]\ W(x,y,m)^{3.52}$$

$$\int \mu_0 T^u{}_a T^d{}_a(x,y,m,\mu_0)[0.16 - 0.05/(1.1\mu_0{}^{1.4} + 0.15)]dt \qquad (11)$$

where x, y, and m refer to latitude, longitude, and month, respectively. Equation (11) is integrated over one day at the mid-point of each month and μ_0 is calculated as a function of latitude, longitude, day of year, and time of day using astronomical formulas. We assume that the values obtained at the mid-point of the month are representative of the entire month. The results of this calculation indicate that whitecap radiative forcing is particularly strong in the Arabian Sea region during the Northern Hemisphere summer. Figure 1 shows the surface wind speed, cloud coverage, and whitecap forcing for mean July conditions. Values of ΔF of up to 0.7 Wm^{-2} occur and are due to a combination of high surface wind speeds and relatively low cloud amounts. Figure 1 also illustrates how high cloud amounts significantly reduce the calculated whitecap radiative forcing. It is important to note that Equation (11) assumes no solar radiative transmission through clouds, thus the values displayed in Figure 1 should be considered conservative.

The above analysis does not imply that the present climate is changing due to whitecaps. Our examination of ECMWF winds did not reveal any significant trend in $<W'>$. By modifying the radiation balance, however, whitecaps may affect the response of the climate system to changes in greenhouse gases and other active atmospheric constituents. Many competing effects and feed-backs may be involved, and are difficult to untangle. It is through the use of coupled climate models and an explicit parameterization of the whitecap contribution to surface albedo that the impact of whitecaps on climate, potentially significant as shown here, can be further quantified and assessed.

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