

Influence of phytoplankton on the global radiation budget

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Received 28 February 2001; revised 12 December 2001; accepted 13 February 2002; published 5 October 2002.

[1] Oceanic phytoplankton may exert a warming influence on the planet by decreasing surface albedo. Compared with the case of pure seawater, the globally and annually averaged outgoing radiative flux is decreased by a probable value of 0.25 Wm^{-2} . This value corresponds to about 20% of the combined radiative forcing by greenhouse gases and anthropogenic aerosols since preindustrial times, including indirect effects. The relative importance of phytoplankton is greater on regional and seasonal scales, with forcing values reaching -1.5 Wm^{-2} in coastal zones and high-latitude regions during summer. The annual amplitude of radiative forcing by phytoplankton is large in subpolar regions, owing to the conjugate action of cloud amount and biomass level. Spatial and temporal variability of the forcing is affected by phytoplankton type, some reflective species increasing the outgoing radiative flux. The effects of space- and time-varying phytoplankton on surface albedo should be taken into account explicitly in the numerical modeling of climate change.

INDEX TERMS: 3359 Meteorology and Atmospheric Dynamics: Radiative processes; 3360 Meteorology and Atmospheric Dynamics: Remote sensing; 4552 Oceanography: Physical: Ocean optics; 4855 Oceanography: Biological and Chemical: Plankton; *KEYWORDS:* Phytoplankton, albedo, radiative budget, climate, ocean color

Citation: Frouin, R. J., and S. F. Iacobellis, Influence of phytoplankton on the global radiation budget, *J. Geophys. Res.*, 107(D19), 4377, doi:10.1029/2001JD000562, 2002.

1. Introduction

[2] In general circulation climate models, even modern-day models such as the National Center for Atmospheric Research community climate model [Kiehl *et al.*, 1998], the albedo of the ocean surface is crudely parameterized. Formulas attempt to capture the effects of key variables, namely sun zenith angle and ratio of diffuse and direct sunlight, but generally do not separate the processes of backscattering by the water body, Fresnel reflection, and wave breaking. The treatment is not satisfactory, not only because the individual processes vary differently with sun geometry and light distribution, but also because unaccounted factors may play a significant role, in particular whitecaps and water composition. Hansen *et al.* [1997] included the contribution of whitecaps and sub-surface scattering in their climate model, but did not incorporate ocean color variability. The sub-surface scattering was for oceanic waters with low phytoplankton pigment concentration such as Sargasso Sea, i.e., 0.05 mgm^{-3} (Howard Gordon, personal communication, 2001).

[3] Depending on the amount of phytoplanktonic algae, detrital particles, yellow substances, sediments, etc., the apparent optical properties of oceanic waters may vary substantially [e.g., Morel and Prieur, 1977]. Case 1 waters, which occupy approximately 98% of the world's oceans,

change color from blue to green and dark green as algae concentration increases, due to absorption by algae pigments in the blue. Case 2 waters, encountered in coastal regions, are generally more reflective in the visible. They are discolored when the sediment load is high (scattering effects are dominant), but may appear brownish in the presence of yellow substances. Due to diverse possibilities, the diffuse reflectance of the ocean just below the surface, averaged over the whole visible spectrum, may change typically from 0.005 to 0.1 [Morel, 1980], i.e., by a factor of 20. Variability remains large above the surface, even though diffuse reflectance is reduced by a factor of about 0.53 due to internal reflection [Austin, 1974].

[4] Coccolithophore blooms, in particular, produce "white waters" readily detectable from space by means of remote sensing [e.g., Holligan *et al.*, 1983]. They can extend over $100,000 \text{ km}^2$ areas [Brown and Yoder, 1994] and contain large numbers of cells and detached coccoliths [Holligan *et al.*, 1993]. Their diffuse reflectance just below the surface may be as high as 0.22 and 0.38 at the wavelengths of 410 and 520 nm, respectively [Balch *et al.*, 1991]. Tyrell *et al.* [1999] indicated that, for typical conditions, increasing calcite (CaCO_3 -carbon) concentration from 0 to 300 mgm^{-3} would increase above-surface reflectance from 0.069 to 0.12 in the photo-synthetically active range. Assuming that blooms cover only 0.3% of the planet's surface and occur during one month each year, as suggested by Coastal Zone Color Scanner (CZCS) analysis [Brown and Yoder, 1994], they estimated that coccolith light

scattering should increase the global, annually averaged albedo by a negligible value of $3 \cdot 10^{-6}$. However, when taking into account sub-blooms or background concentrations, they obtained a maximum effect of $1.3 \cdot 10^{-3}$, which they indicated is equivalent to a radiative forcing of about 0.22 Wm^{-2} .

[5] In the following, we focus on changes in surface reflection due to phytoplankton. As mentioned above, and as recognized by *Holligan* [1992], the changes are potentially important; they may affect significantly the response of the climate system to increasing concentrations of greenhouse gases and anthropogenic aerosols. In any climate change scenario, the amount and distribution of phytoplanktonic algae in the oceans would also be sensitive, through various feedbacks, to dynamical effects of the ocean and atmosphere. Using radiation-transfer modeling and CZCS-pigment imagery, we compute the direct radiative forcing by phytoplankton and discuss its magnitude and variability, as well as consequences on climate and projections of climate change.

2. Radiative Forcing

[6] To quantify the effect of phytoplankton on the global radiation budget, we adopt the approach used by *Iacobellis et al.* [1999] for biomass-burning aerosols and *Frouin et al.* [2001] for whitecaps. At any location over the oceans the change in outgoing radiative flux is

$$\Delta F = F_0 \mu_0 [1 - (1 - T_c^u T_c^d) N] T_a^u T_a^d \Delta A_s \quad (1)$$

where F_0 is the extraterrestrial broadband solar irradiance, μ_0 is the cosine of the sun zenith angle (in air), N is the fraction of the surface covered by clouds, T_a^u and T_a^d are the clear atmosphere transmittances and T_c^u and T_c^d the cloud transmittances for up-welling and down-welling flux, respectively, and ΔA_s is the surface albedo change due to phytoplankton and their derivative, detrital products. Atmosphere-surface interactions are neglected in equation (1), but forcing in the presence of clouds, not completely opaque to incident sunlight, is taken into account by reducing N by the factor $(1 - T_c^u T_c^d)$.

[7] An estimate of ΔA_s can be obtained by modeling A_s as the sum of contributions due to diffuse reflection by the water body, diffuse reflection by whitecaps, and Fresnel reflection by the whitecap-free surface. Using *Koepke's* [1984] formulation, A_s is expressed as

$$A_s = f_{wc} A_{wc} + (1 - f_{wc}) A_f + (1 - f_{wc} A_{wc}) A_w \quad (2)$$

where f_{wc} is the fraction of the surface covered by whitecaps, A_{wc} is the effective albedo of whitecaps, A_f is the Fresnel albedo, and A_w is the diffuse albedo of the water body. The weight $(1 - f_{wc} A_{wc})$ is introduced to take into account the reduction of under light due to whitecaps. In this formulation, the possibility of whitecaps interacting with the adjacent surface is ignored. Consequently the change in surface albedo is

$$\Delta A_s = (1 - f_{wc} A_{wc}) (A_w - A_{w0}) \approx A_w - A_{w0} \quad (3)$$

where A_{w0} is the diffuse albedo of optically pure seawater. The term $f_{wc} A_{wc}$ can be neglected because f_{wc} is generally

less than 0.1, except in rare conditions of very strong winds, and A_{wc} is about 0.16 [*Frouin et al.*, 2001], giving $f_{wc} A_{wc} < 0.016 \ll 1$. All oceanic waters are treated as Case 1 waters, i.e., A_s is determined by biogenous matter content. Of course, sediments and/or yellow substances may dictate the optical properties of some coastal waters (Case 2 waters). These waters, however, are present in a small fraction of the oceans (<2%), and treating them as Case 1 waters is expected to impact negligibly the global radiative budget.

[8] Following *Morel and Antoine* [1994], A_w is defined as

$$A_w = \int R_w(0^+, \lambda) F_d(0^+, \lambda) d\lambda / \int F_d(0^+, \lambda) d\lambda \quad (4)$$

where λ is wavelength, $R_w(0^+, \lambda)$ is the diffuse spectral reflectance of the ocean just above the surface, $F_d(0^+, \lambda)$ is the downwelling spectral solar irradiance just above the surface, and the integral is over the entire solar spectrum. The spectral composition of F_d depends on absorption and scattering by atmospheric constituents (molecules, aerosols, and clouds). Since radiative forcing by phytoplankton is expected to occur mostly in clear sky conditions, F_d is computed for a clear, standard atmosphere containing maritime aerosols. In the computation, the analytical formula developed by *Tanré et al.* [1979] is used for diffuse spectral atmospheric transmittance.

[9] The diffuse reflectance $R_w(0^+, \lambda)$ can be modeled as

$$R_w(0^+, \lambda) = t R_w(0^-, \lambda) \approx T_w \gamma b_b / a \quad (5)$$

where T_w is the water-air interface transmittance, $R_w(0^-, \lambda)$ is the diffuse spectral reflectance just below the surface, γ is a factor characterizing the dependence of $R_w(0^-, \lambda)$ upon sun zenith angle, b_b is the backscattering coefficient, and a is the absorption coefficient. The variables on the right-hand side of equation (5) are wavelength-dependent, but the symbol λ is omitted for clarity.

[10] *Austin* [1974] reported T_w values varying from 0.515 to 0.537 (a small change) when wind speed increased from 0 to 16 ms^{-1} . The values were obtained for a refractive index of 1.341 and assuming that the radiance below the surface was constant with angle. Even though this assumption is not precisely correct, and T_w varies spectrally, the average value of 0.526 is used for all wavelengths. A more accurate modeling of T_w is not necessary, because the variability of (and uncertainty on) b_b and a is relatively large (see below).

[11] The angular factor γ depends not only on sun zenith angle, but also on water type and sky conditions [*Kirk*, 1984; *Gordon*, 1989; *Morel and Gentilli*, 1991]. For waters dominated by particles, for example, γ may vary from 0.35 to 0.55. *Morel and Gentilli* [1991] parameterized γ as a function of a modified sun zenith angle and of the ratio of b_{b0} , the backscattering coefficient of pure seawater, and b_b . The modified sun zenith angle is a function of the diffuse fraction of incident sunlight and, when the radiation is direct, reduces to the actual sun zenith angle. Since, as mentioned above, radiative forcing is dominant under cloudless skies, and since sunlight is mostly direct under such skies, the effect of diffuse sunlight is neglected. The

following approximate formula for direct radiation [Morel and Gentili, 1991] is used:

$$\gamma = 0.6279 - 0.2227(b_{b0}/b_b) - 0.0513(b_{b0}/b_b)^2 - 0.3119\mu_0 + 0.2465(b_{b0}/b_b)\mu_0 \quad (6)$$

[12] The optical coefficients, b_b and a , are modeled according to Morel [1988]. They are expressed in terms of a single biological variable, the phytoplankton pigment (chlorophyll-a + phaeopigment-a) concentration, C , as follows:

$$b_b(\lambda) = b_{b0}(\lambda) + b_{bp}(\lambda) = b_{b0}(\lambda) + b^0 C^{0.62} [0.002 + 0.02 \cdot (0.5 - 0.25 \log_{10} C) (550/\lambda)] \quad (7)$$

$$a(\lambda) = K(\lambda)/u = [K_0(\lambda) + X(\lambda)C^{e(\lambda)}]/u \quad (8)$$

where b_{bp} is the backscattering coefficient due to suspended particles, K is the vertical diffuse attenuation coefficient (assumed to be related linearly to a via the factor u), K_0 is the vertical diffuse attenuation coefficient for pure seawater, b^0 , X , and e are empirical coefficients, and C is expressed in mgm^{-3} and λ in nm. The second term on the right-hand side of equation (8) represents the contribution to a of material in seawater (phytoplankton, dissolved substances). The coefficient b^0 varies in the range 0.12–0.45 m^{-1} , with a mean value of 0.3 m^{-1} [Gordon *et al.*, 1988]; it is fixed at the mean value. The backscattering coefficient for pure seawater, b_{b0} , is half the scattering coefficient, b_0 . The b_0 values are taken from Morel [1974] for wavelengths between 350 and 600 nm, and extrapolated spectrally using the approximate power law $\lambda^{-4.3}$. The spectral dependence of b_{bp} is in λ^{-1} (equation 7), except when C is large (no spectral dependence); but a wider range of exponents, at least between -2 and 0 , may be encountered [Bricaud *et al.*, 1983; Stramski and Kiefer, 1991; Lee *et al.*, 1994]. The values tabulated in Morel and Antoine [1994] are used for K_0 , X , and e . Variability in X is large due to phytoplankton type, by at least $\pm 30\%$ [e.g., Bricaud *et al.*, 1995, 1998]. New, probably more accurate spectral absorption coefficients of pure seawater and, thus, K_0 have been proposed by Pope and Fry [1997]. It is sufficient, for the purpose of the study, to use the Morel and Antoine [1994] values for K_0 , all the more as X and e were obtained from measurements of K by subtracting the K_0 values. In the computation of $R_w(0^-, \lambda)$, the linear relationship between K and u is actually relaxed using an iterative procedure (see Morel [1988] for details).

[13] Using equations (4) to (8) A_w , A_{w0} , and thus ΔA_s can now be evaluated. Since $R_w(0^-, \lambda)$ is practically equal to zero above 750 nm, and since a very small amount of solar radiation ($< 0.5\%$ of total) reaches the ocean surface below 300 nm, the contribution of those wavelengths to the integrals on the right-hand side of equation (4) is neglected. Figure 1 displays ΔA_s as a function of C for $\mu_0 = 1$ (top) and $\mu_0 = 0.5$ (bottom). Due to phytoplankton type variability, the actual relationship between ΔA_s and C may lie between the upper and lower dashed curves of Figure 1. These curves were obtained by varying b^0 from 0.12 to 0.45 m^{-1} , X by $\pm 30\%$, and the spectral dependence of b_{bp}

from λ^{-2} to λ^0 . No correlation was assumed between b^0 , X , and the spectral dependence of b_{bp} , and changes in e were ignored.

[14] For the average conditions (Figure 1, solid curves), ΔA_s is generally negative and decreases with increasing C , to about -0.005 at $C = 50 \text{ mgm}^{-3}$. The effect is larger in magnitude for a sun at zenith than for a sun at 60 degrees from zenith, except when C is large ($> 10 \text{ mgm}^{-3}$). This is due to the fact that γ increases more rapidly with decreasing μ_0 in the presence of suspended particles ($b_{b0}/b_b = 1$). Positive ΔA_s values may be encountered when particulate backscattering is dominant (Figure 1, upper dashed curves), reaching 0.002 for $\mu_0 = 1$ and 0.004 for $\mu_0 = 0.5$. For coccolithophores, the scattering coefficient may be much larger than the one used in the upper dashed curves of Figure 1, resulting in larger positive ΔA_s values. Thus variability and, therefore, uncertainty in the relationship between ΔA_s and C may be large locally, but extreme conditions are unlikely to occur over a major part of the oceans. Further, variability is reduced on the regional and seasonal scales considered in the present study.

[15] The monthly averaged perturbation in reflected solar flux, $\langle \Delta F \rangle$, is estimated by combining equations (1) and (3) and integrating over time, t :

$$\langle \Delta F \rangle = F_0 \int \mu_0 (A_w - A_{w0}) [1 - (1 - T_c^u T_c^d) N] T_a^u T_a^d dt \quad (9)$$

where the integral is over one day at the midpoint of each month, $F_0 = 1372 \text{ Wm}^{-2}$, $A_w - A_{w0}$ is obtained from equations (4)–(8), and μ_0 is calculated as a function of latitude, longitude, day of year, and time of day using astronomical formulas. It is assumed in equation (9) that the $\langle \Delta F \rangle$ values obtained at the midpoint of the month are representative of the entire month. Earth-Sun distance variations during the course of the year, which affect F_0 by a few percent, are neglected.

[16] To estimate the integral on the right-hand side of equation (9), T_a^u , T_a^d , T_c^u , and T_c^d are computed as a function of μ_0 using the SUNRAY radiation-transfer model [Fouquart and Bonnel, 1980]. In this model the solar spectrum is divided into two spectral intervals, 200–680 nm and 680–4000 nm. Only the results obtained in the short-wave interval are used, since $R_w(0^+, \lambda)$ is practically equal to zero in the long-wave interval. Necessary input to SUNRAY includes the atmospheric profiles of temperature, ozone, and moisture, and cloud and aerosol variables (thickness, single scattering albedo, and asymmetry factor). Clouds are placed in the model layer containing the 650 hPa level, while the aerosols are placed in the region extending from 950 to 750 hPa (i.e., completely beneath the clouds). The vertical profiles from five standard atmospheric types (tropical, midlatitude summer and winter, sub-artic summer and winter) are used. Aerosol properties are those of the maritime model (MAR-1) 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 550 nm. Fractional cloud coverage, N , and cloud optical thickness are obtained from 7 years of monthly International Satellite Cloud Climatology (ISCCP) C2 data [ISCCP, 1992]. The ISCCP estimates of cloud optical thickness were made at approx-

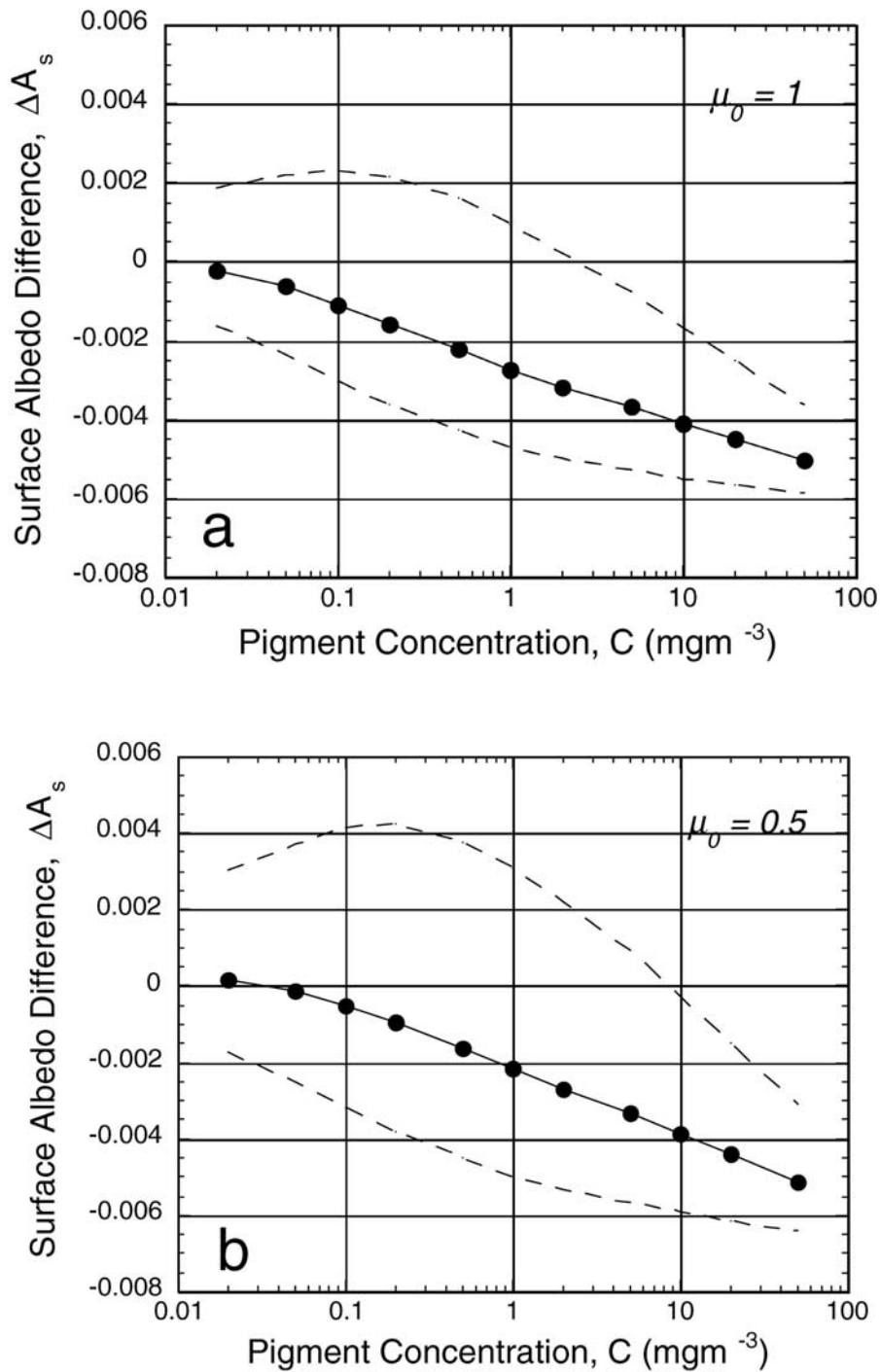


Figure 1. Change in surface albedo due to phytoplankton. (a) Sun zenith angle of 0 degree ($\mu_0 = 1$). (b) Sun zenith angle of 60 degrees ($\mu_0 = 0.5$). Solid lines correspond to average optical properties of phytoplankton. Dashed lines delineate phytoplankton-type variability.

imately 600 nm, but they are assumed to be representative of the short-wave interval. The cloud asymmetry factor is fixed at 0.865 in the short-wave interval, and the cloud single scattering albedo is parameterized as a function of cloud optical thickness [see Fouquart and Bonnel, 1980]. Surface albedo A_w is computed with monthly averages of

phytoplankton pigment concentration provided by Brookhaven National Laboratory (courtesy of Paul G. Falkowski). The data set was created from the 1978–1986 CZCS archive [Feldman *et al.*, 1989]. It is a quality-controlled data set used in primary production studies [see Behrenfeld and Falkowski, 1997]. Linear interpolation based on neighbor-

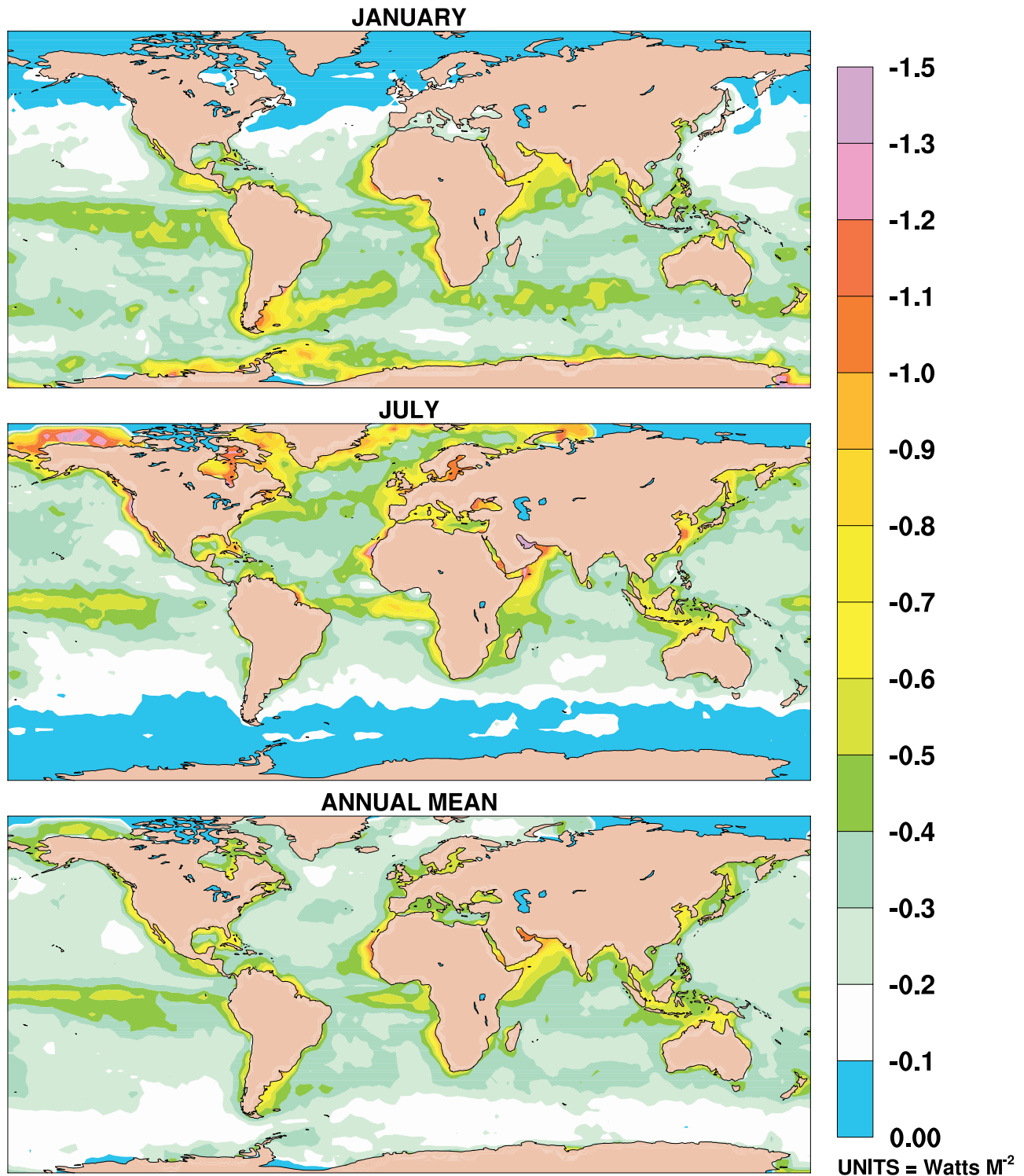


Figure 2. Mean geographic distribution of the change in outgoing solar flux from the planet due to phytoplankton for January (top), July (middle), and the year (bottom).

ing pixels was applied to fill up missing information, and the data was remapped onto the ISCCP C2 grid.

3. Results and Discussion

[17] The mean geographic distribution of radiative forcing by phytoplankton is displayed in Figure 2 for January

(top panel), July (middle panel), and the year (bottom panel). Calculations were made using equation (9) and average optical properties for the phytoplankton (i.e., the solid lines in Figure 1). The resulting forcing varies from -1.5 Wm^{-2} to practically zero, with low values in coastal regions where phytoplankton is relatively abundant. Low values are also obtained at subpolar latitudes, but only

during summer when available solar radiation and biological production are large. At these latitudes, the annual amplitude of radiative forcing reaches 1.5 Wm^{-2} due to the conjugate action of phytoplankton biomass and incident solar irradiance, both high during summer and low during winter. On average globally and annually, the outgoing radiative flux from the planet is decreased by a probable value of 0.25 Wm^{-2} with respect to pure seawater. Due to variability in the optical properties of phytoplankton (see previous section), however, the actual change may lie within a range of -0.51 to 0.20 Wm^{-2} .

[18] The probable decrease of 0.25 Wm^{-2} in reflected solar flux is not negligible compared with the direct radiative forcing by greenhouse gases and anthropogenic aerosols that may have occurred since preindustrial times. It corresponds, in relative magnitude, to approximately 15% of the value for carbon dioxide, 10% of the value for all greenhouse gases, 65, 130, and 250% of the values for sulphate, biomass-burning, and soot aerosols, respectively, and 35% of the total aerosol forcing [Houghton *et al.*, 1996]. It is also more than 8 times larger than the average effect of whitecaps, which increase surface albedo [Frouin *et al.*, 2001]. In this comparison, inter-annual variability of phytoplankton is neglected. Including direct and indirect effects of ozone and the indirect effect of aerosols (assumed to be 0.8 Wm^{-2} despite the large uncertainty), the decrease of 0.25 Wm^{-2} due to phytoplankton represents about 20% of the total forcing by greenhouse gases and anthropogenic aerosols. Until 1920, the phytoplankton forcing was relatively more important, i.e., about half the greenhouse-gas effect since 1850. If the IS92a scenario [Houghton *et al.*, 1996] materialized, however, and assuming no significant phytoplankton feedbacks (i.e., change in amount and type), in 2050 the phytoplankton forcing would represent only a few percent of the net greenhouse-gas and aerosol forcing.

[19] Assuming a climate sensitivity of 0.3 to 1.4 K per Wm^{-2} , and using the above estimate of -0.25 Wm^{-2} , the impact of phytoplankton on the global mean surface temperature of the planet is a warming of 0.07 to 0.35 K. The impact is significant in view of the 0.6 K increase due to greenhouse-gas and aerosol emissions predicted by climate models during the period 1850–1990 [e.g., Mitchell *et al.*, 1995, 1997] and inferred from temperature records during the same period [Parker *et al.*, 1994; Mann *et al.*, 1998]. Current climate models that neglect phytoplankton forcing would predict a higher global mean surface temperature than actually observed if they included phytoplankton forcing. Parameterizations of surface albedo, however, take into account some phytoplankton contribution, either implicitly when based on measurements [e.g., Briegleb and Ramanathan, 1982; Taylor *et al.*, 1996] or explicitly [Hansen *et al.*, 1997], and the resultant extra warming may not be as large as expected. Figure 3 displays the residual phytoplankton forcing for January (top), July (middle), and the year (bottom) with respect to an ocean with a constant pigment concentration of 0.05 mgm^{-3} (case of Hansen *et al.* [1997]). Compared with Figure 2, forcing values now range between -1.2 and 0.2 Wm^{-2} , and regional and seasonal variability remains large. The globally and annually average value is substantially reduced in magnitude, i.e., -0.1 Wm^{-2} instead of -0.25 Wm^{-2} , yet not negligible. Model experiments with space- and time-varying

phytoplankton are in order to quantify and assess the resulting pattern of climate change.

[20] The warming influence of phytoplankton, 0.07 to 0.35 K, would also be effective at the end of the 19th Century and beginning of the 20th Century. For this preindustrial period, and the previous 1000 years, large-scale temperature variations are well explained by solar, volcanic, and greenhouse-gas forcing [Crowley, 2000]. Residuals are similar to the internal variability of climate models, i.e., $\pm 0.1 \text{ K}$, and there is no significant bias. Since climate models that ignore phytoplankton variability reproduce faithfully preindustrial temperatures, missing processes might be at work to counteract or reduce the effect of natural and anthropogenic forcing. It is perhaps more plausible that some of the inventoried processes are not taken into account properly (e.g., because of uncertainties). Seasonal flux adjustments could be applied in the presence of phytoplankton forcing, to reconstruct closely the unperturbed present-day climate in control runs. In principle the new flux adjustments should be smaller than those without phytoplankton forcing. Larger adjustments would point to model deficiency and the need for a more complete and realistic representation of the climate system. The gross patterns of change due to greenhouse-gas and aerosols reported in previous studies [see Houghton *et al.*, 1996] would probably remain unchanged with flux adjustments, but potentially large modifications and unforeseen feedbacks might be revealed on regional and seasonal scales. In any case, making flux adjustments or not, explicit parameterization of the effect of space- and time-varying phytoplankton on surface albedo appears necessary to improve projections of future climate change.

[21] The above results do not imply that the climate of our planet is changing due to phytoplankton. Analysis of existing pigment concentration data sets, such as the National Oceanographic Data Center (NODC) in-situ archive (starting in 1957) and satellite imagery of CZCS (1978–1986) and Sea-Viewing Wide Field-of-view Sensor (SeaWiFS) (starting in 1997), so far has not provided conclusive evidence for a global trend in phytoplankton pigment concentration. Yet significant changes have been documented for some regions. Venrick *et al.* [1987], for example, reported a nearly doubled chlorophyll-a concentration in the North Pacific during the period 1968–1985.

[22] Biological production certainly can be modified by global warming, either via direct effects of temperature and sunlight on growth rates and metabolic processes or via changes in ocean circulation, in particular the rate of upwelling. Bakun [1990] proposed that greenhouse warming, by intensifying along shore wind stress, would increase coastal upwelling and, therefore, primary production. Sarmiento *et al.* [1998], on the other hand, argued that the general warming trend, by increasing stratification, would reduce the flux of nutrients to the upper ocean and, therefore, the effectiveness of the biological pump. They also suggested that increased precipitation, by decreasing the aeolian transport of nutrients (iron), would expand high-nutrient/low-chlorophyll regions.

[23] Due to the non-linearity in the relationship between phytoplankton abundance and surface albedo, however, a change in pigment concentration from current levels, even

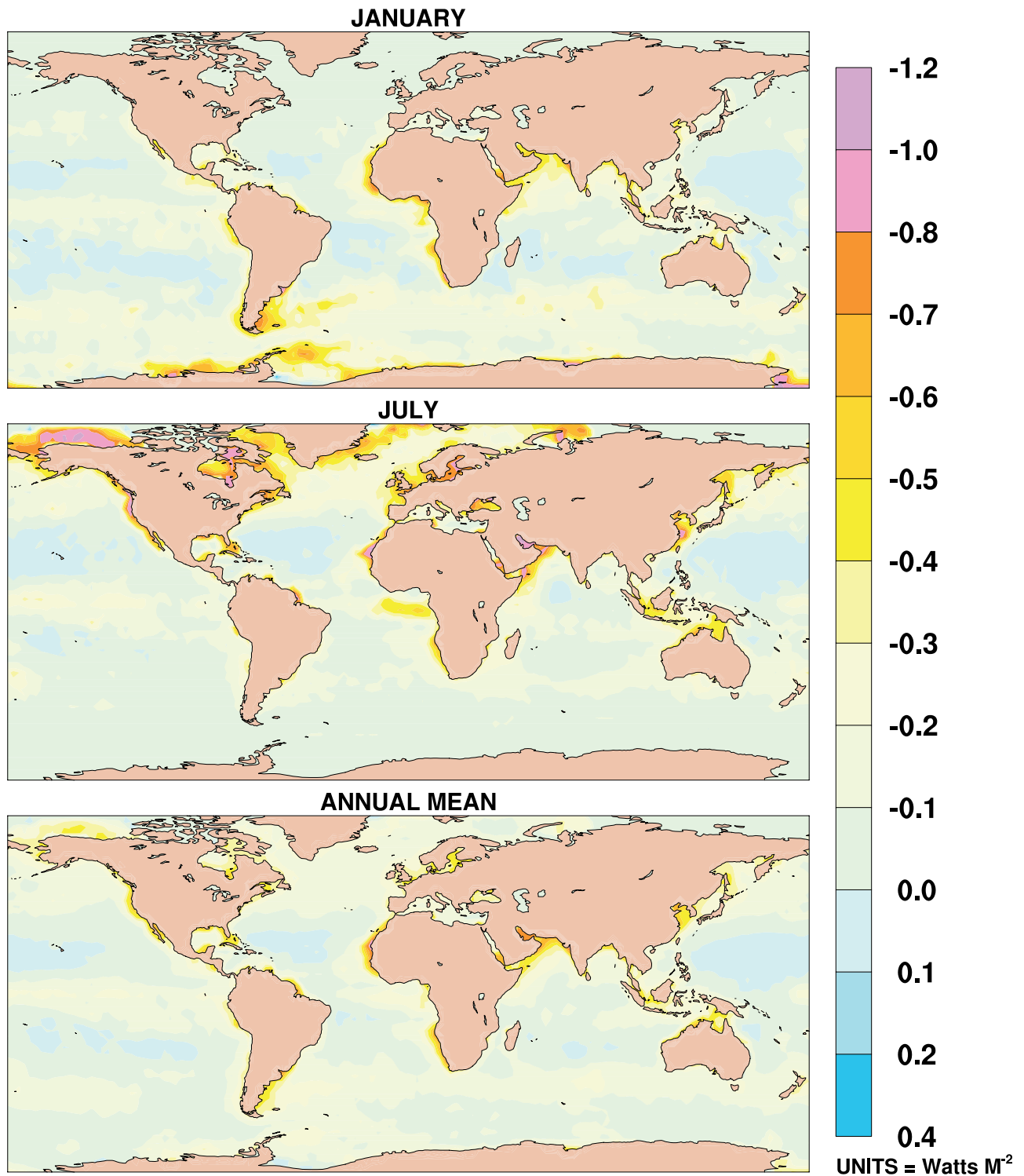


Figure 3. Same as Figure 2, but with respect to an ocean with a surface pigment concentration of 0.05 mgm^{-3} everywhere instead of an ocean containing only pure seawater.

large, would result in a small change in surface albedo and, therefore, radiative forcing. Figure 4 shows the sensitivity of radiative forcing to pigment concentration, assuming the same distribution in space and time. The scale factor represents the change in pigment concentration with respect to 1978–1986 CZCS climatology. Doubling pigment concentration everywhere (scale factor of 2) would only

decrease the outgoing flux by 0.07 Wm^{-2} (30% of the mean value). Adding only 10% of the present phytoplankton biomass to an ocean containing pure seawater would produce the same effect. Figure 4 suggests that the response of radiative forcing to a given change in pigment concentration would be larger in oligotrophic waters. Small effects are expected in productive, coastal waters.

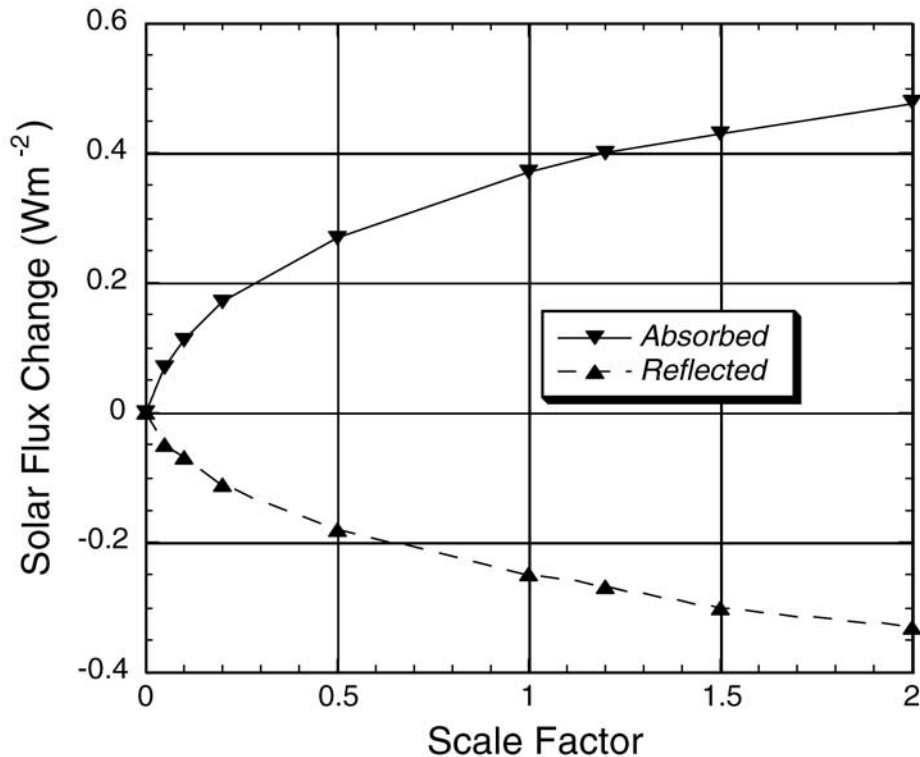


Figure 4. Annual mean change in solar flux at the top of the atmosphere (solid line) and absorbed at the surface (dashed line) due to change in the phytoplankton pigment concentration. The scale factor is normalized to 1979–1986 CZCS pigment concentration. The absorbed flux at the surface is computed using equation (1) with $T_a^u = T_c^u = 1$.

[24] It has been argued that a strengthening of the biological pump might reduce atmospheric carbon dioxide concentration and consequently global warming [McElroy, 1983; Martin, 1990; see also Falkowski and Raven, 1997]. Martin [1990] contemplated the “ultimate enrichment experiment”, in which large quantities of iron would be artificially injected in the Southern Ocean. Siegenthaler and Sarmiento [1993] pointed out that iron fertilization would not stop the carbon dioxide increase; it would only reduce the growth rate. Iron fertilization might also increase the release to the atmosphere of greenhouse gases with large warming potential such as nitrous oxide, which would counteract the cooling effect of reduced carbon dioxide [Fuhrman and Capone, 1991]. If coastal upwelling were increased as a consequence of global warming [Bakun, 1990], surface waters would be enriched not only in nutrients, but also in dissolved carbon, and this carbon might be sufficient for organic production (i.e., no need for carbon from the atmosphere).

[25] In any event, an increase in phytoplankton abundance associated with enhanced upwelling or biological pumping of atmospheric carbon dioxide, natural or artificial, would increase absorption of solar radiation and, therefore, heating of the upper ocean. The increased heating, and more importantly the trapping of more solar radiation near the surface [Frouin *et al.*, 2000], would contribute to reducing mixed layer depth, increasing sea surface temperature, and warming the lower troposphere. This negative climate feed-

back, due to ocean color, would exacerbate the effect mentioned by Falkowski *et al.* [1998] of increased heat storage and stratification and of reduced nutrients that would result from a warmer atmosphere.

[26] The present study made use of 1979–1986 CZCS imagery. This data set covers two El Niño episodes, including the exceptional 1982–1983 event, and therefore might not represent the long-term average conditions. Pigment concentrations could be biased low, at least in the central Pacific, although values higher than average are generally observed in this region during non-El Niño (i.e., La Niña) years [e.g., Chavez *et al.*, 1999]. Furthermore, the CZCS did not have optimum spectral bands for atmospheric correction, affecting the accuracy of marine reflectance retrievals. Other, more accurate and global pigment concentration data sets exist or are being produced, from SeaWiFS (starting in 1997), the ocean Color and Temperature Sensor (OCTS) and the Polarization and Directionality of the Earth Reflectance (POLDER) instrument (1996–1997), and the MODerate-resolution Imaging spectrometer (MODIS) (starting in 2000). They are currently being merged with historical in situ data [Gregg and Conkright, 2001] and the resulting climatology, when available, will allow a more accurate evaluation of the phytoplankton forcing.

[27] As indicated previously, reflective phytoplankton species may increase surface albedo, introducing uncertainties in the global forcing estimate. According to Tyrell *et al.*

[1999], the outgoing flux from the planet would be increased by as much as 0.22 Wm^{-2} due to coccolithophores, which would practically offset the effect of -0.25 Wm^{-2} computed for "average" phytoplankton. Tyrell *et al.* [1999], however, made drastic assumptions about sub-bloom concentrations, such as all the calcite in surface waters is in the form of coccoliths and the concentration of calcite across Atlantic oligotrophic gyres is the same for all the oceans. Their maximum radiative forcing estimate, therefore, is uncertain and probably overestimated. After all, sub-blooms are not detected in the satellite imagery. The best approach to the problem, and the object of future study, is to work directly with satellite-derived marine reflectance and compare this reflectance with that of pure seawater. This approach, which does not require invoking bio-optics or knowing phytoplankton type, will yield more precise estimates of the effect on planetary albedo of all variable ocean absorbers and scatterers, not just phytoplankton.

[28] **Acknowledgments.** This work was supported by the National Aeronautics and Space Administration, the Department of Energy, and by the California Space Institute.

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