

Determination of surface heating by convective cloud systems in the central equatorial Pacific from surface and satellite measurements

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Abstract. The heating of the ocean surface by longwave radiation from convective clouds has been estimated using measurements from the Central Equatorial Pacific Experiment (CEPEX). The ratio of the surface longwave cloud forcing to the cloud radiative forcing on the total atmospheric column is parameterized by the f factor. The f factor is a measure of the partitioning of the cloud radiative effect between the surface and the troposphere. Estimates of the f factor have been obtained by combining simultaneous observations from ship, aircraft, and satellite instruments. The cloud forcing near the ocean surface is determined from radiometers on board the National Oceanic and Atmospheric Administration P-3 aircraft and the R/V *John Vickers*. The longwave cloud forcing at the top of the atmosphere has been estimated from data obtained from the Japanese Geostationary Meteorological Satellite GMS 4. A new method for estimating longwave fluxes from satellite narrowband radiances is described. The method is based upon calibrating the satellite radiances against narrowband and broadband infrared measurements from the high-altitude NASA ER-2 aircraft. The average value of f derived from the surface and satellite observations of convective clouds is 0.15 ± 0.02 . The area-mean top-of-atmosphere longwave forcing by convective clouds in the region 10°S – 10°N , 160°E – 160°W is 40 W/m^2 during CEPEX. These results indicate that the surface longwave forcing by convective clouds was approximately 5 W/m^2 in the central equatorial Pacific and that this forcing is the smallest radiative component of the surface energy budget.

1. Introduction

The tropical clouds generated by deep convection have significant radiative effects on the ocean-atmosphere energy budget. The magnitude of the effect results from the extensive areal coverage of tropical cirrus and the large longwave cloud forcing by cirrus near the tropopause [Ackerman *et al.*, 1988]. Because the upper tropical troposphere is much colder than the sea surface, optically thick convective clouds can reduce the outgoing thermal radiation by 100 W/m^2 relative to clear-sky conditions. The partitioning of the column longwave heating by these clouds between the ocean and the atmosphere is determined by the infrared optical opacity and temperature profile of the troposphere. Radiative transfer calculations predict that most of the radiative heating associated with convec-

tive cloud systems is confined to the middle and upper troposphere [Slingo and Slingo, 1988].

Experimental evidence that the downward longwave flux at the tropical ocean surface does not vary significantly in the presence of clouds dates back at least to the GARP Atlantic Tropical Experiment (GATE) [Davies and Uboegbulam, 1979]. More recent evidence indicating that the surface longwave fluxes are relatively constant has been obtained from Coupled Ocean-Atmosphere Response Experiment (COARE) pilot cruises [Young *et al.*, 1992] and from the COARE Intensive Observing Period [Weller and Anderson, 1996]. In contrast to these earlier studies, the present analysis examines the relationship between the surface longwave fluxes and the top-of-atmosphere (TOA) longwave cloud radiative forcing in the manner suggested by Harshvardhan *et al.* [1990]. Global analyses of this relationship have been performed using satellite measurements of TOA fluxes and model-based retrievals of the surface fluxes [Wu *et al.*, 1992]. We focus on the central tropical Pacific and use direct observations of the surface fluxes and estimates of the TOA fluxes derived using aircraft and satellite data.

The fraction of the longwave cloud forcing $C_c(0)$ in a column of the ocean-atmosphere system that contributes to the heating of that column may be characterized by the f factor [Ramanathan and Collins, 1991]. This definition of f is difficult to quantify experimentally because it involves measurement of the vertical diabatic heating profile. However, if the ocean heat transports are neglected and the atmospheric transports are very efficient, then f is also the fraction of the cloud longwave forcing that reaches the ocean surface. This quantity may be

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readily determined from surface and satellite radiometric data. The f factor is defined in terms of the cloud forcing at the surface ($C_l(p_s)$) and the top of the atmosphere ($C_l(0)$) by

$$f = \frac{C_l(p_s)}{C_l(0)}, \quad (1)$$

where

$$C_l(p_s) = F^-(p_s) - F_c^-(p_s), \quad (2)$$

$$C_l(0) = F_c^+(0) - F^+(0). \quad (3)$$

Here F_c^+ and F_c^- are the upwelling and downwelling clear-sky flux, and $F^\pm$ is the mean flux averaged over cloudy and clear conditions. Since the tropical atmosphere is optically thick in the infrared, the f factor for convective cloud systems is estimated to range between 0 and 0.2 [Ramanathan and Collins, 1991]. For example, the radiative calculations by Slingo and Slingo [1988] imply that $f = 5/191$ for a high cloud above 10 km. The upper bound on f is derived from model calculations of the amount of radiation from the base of cirrus clouds that reaches the ocean surface.

Consider an atmosphere with a single layer of cloud with fractional cloud area A , broadband emissivity ε , and cloud-base and cloud-top pressures p_b and p_t , respectively. The corresponding cloud-base and cloud-top temperatures are T_b and T_t . The sea surface temperature is T_s , and the ocean is assumed to have unit emissivity. In terms of an emissivity/absorptivity formulation of radiative transfer [Ramanathan and Downey, 1986], the fluxes are approximately

$$F_c^-(p_s) = B(0)\varepsilon_c(0, p_s) + \int_0^{p_s} \alpha_c(p_s, p') dB(p') \quad (4)$$

$$F^-(p_s) = F_c^-(p_s)(1 - A\varepsilon) + A\varepsilon \left\{ B(p_b) + \int_{p_b}^{p_s} \alpha_c(p_s, p') dB(p') \right\} \quad (5)$$

$$F_c^+(0) = B(p_s) - \int_0^{p_s} \alpha_c(0, p') dB(p') \quad (6)$$

$$F^+(0) = F_c^+(0)(1 - A\varepsilon) + A\varepsilon \left\{ B(p_t) - \int_0^{p_t} \alpha_c(0, p') dB(p') \right\}. \quad (7)$$

Here $B(p)$ is the Stefan-Boltzmann flux σT^4 evaluated at pressure p , and $\varepsilon_c(p, p')$ and $\alpha_c(p, p')$ are the clear-sky emissivity and absorptivity between two atmospheric layers. It follows from the definition of f and from (2)–(7) that f is independent of cloud amount and cloud emissivity in this approximation. The quantities that determine f are the atmospheric emissivity and absorptivity and the cloud physical properties. The independence from cloud fraction will be used to generalize the Central Equatorial Pacific Experiment observations using cloud distributions determined from satellite.

This study presents an experimental determination of f from surface and satellite measurements of high-altitude clouds over the central Pacific. It is possible to estimate f from available satellite climatologies of the surface energy budget, but the

reliability of these retrievals is not firmly established [Ellingson, 1995]. The calculation of the surface cloud-forcing $C_l(p_s)$ in f requires accurate values for the cloud physical parameters and the vertical profiles of temperature and water vapor. The downward fluxes are sensitive to the vertical resolution of the radiative transfer calculation and the atmospheric profile, since a large fraction of the flux is contributed by the first kilometer of the atmosphere [Ellingson, 1995]. For these reasons, we have estimated f directly from flux measurements close to the tropical ocean surface. The data were collected during the Central Equatorial Pacific Experiment (CEPEX) [Ramanathan et al., 1993] from ship and aircraft instruments in the region bounded by 3°S–2°N and 165°E–155°W. These data sets have made it possible to determine the f factor in regions of intense tropical convection and high sea surface temperature (SST).

A method for estimating TOA longwave fluxes from the GMS 4 meteorological satellite has been developed as part of this study. Although global outgoing longwave radiation (OLR) fields are routinely derived from meteorological satellite imagery [e.g., Gruber et al., 1994; Zhang et al., 1995], several considerations motivated the development of a new algorithm. First, only limited broadband TOA radiation measurements were collected during CEPEX or the preceding Tropical Ocean–Global Atmosphere (TOGA) COARE field campaigns from a single satellite. The wide-field-of-view NASA Earth Radiation Budget Satellite (ERBS) radiometers were still operational, but the spatial and temporal resolutions of these data are not adequate for detailed studies of individual convective events. Many applications of the field data require accurate estimates of OLR over the central and western Pacific. Second, development of multiple independent estimates of OLR allows an assessment of the uncertainty introduced by differences in algorithms. Third, the OLR has been derived at higher spatial and temporal resolution than the global OLR products to allow more precise collocation with surface measurements. Finally, in CEPEX and TOGA COARE the NASA ER-2 high-altitude aircraft carried a suite of instruments measuring infrared radiances and fluxes in the lower stratosphere. While it would have been preferable to combine the ER-2 and surface data to estimate f , there are unfortunately very few ER-2 and surface measurements that are coincident in space and time. Instead, we have used the ER-2 data to develop an algorithm for OLR using Geostationary Meteorological Satellite (GMS) imagery. The hourly frequency, high spatial resolution, and spatial coverage of the GMS imagery resolves the issues regarding data collocation in the estimation of f .

The novel features of the method developed for this study are the calibration of the satellite against the aircraft radiometers and the use of simultaneous aircraft observations to estimate broadband fluxes from narrowband radiances. In general, the methods for estimating OLR from satellite imagery are based upon empirical relationships [e.g., Ohring et al., 1984; Minnis et al., 1991], theoretical relationships [e.g., Cheruy et al., 1991], or model calculations based upon satellite retrievals of cloud and atmospheric properties [e.g., Zhang et al., 1995]. The method described here falls into the class of empirical relationships, with the distinction that the relationship was obtained by combining satellite and aircraft observations. The advantage of this technique compared with existing methods is that it is explicitly designed for consistency with the aircraft data. A comparison of the OLR produced by the present

approach with other methods [Minnis *et al.*, 1991; Zhang *et al.*, 1995] is given in section 3.4.

The characteristics of the CEPEX data and the methods used to identify clear sky are discussed in section 2. A new method for estimating $C_f(0)$ from geosynchronous satellite measurements is described and validated in section 3. The calculations of the f factor and its characteristic values for convective cloud systems are presented in section 4. The extension of the CEPEX observations to the central Pacific and the relation of the CEPEX observing period to climatology are discussed in section 5.

2. Surface and Satellite Radiometric Observations

The surface cloud forcing $C_f(p_s)$ has been derived from instruments on the R/V *John Vickers*, operated by the University of Southern California, and a Lockheed P-3 turboprop aircraft operated by the National Oceanic and Atmospheric Administration (NOAA) [Williams, 1993]. The top-of-atmosphere forcing $C_f(0)$ has been estimated from narrow-band radiometers on board the Japanese Geostationary Meteorological Satellite GMS 4. The method for deriving longwave fluxes from the satellite data is based upon a comparison of the GMS observations with coincident high-altitude measurements from the NASA ER-2 aircraft. Each of these data sets is described below.

2.1. Radiometric Instrumentation on the R/V *John Vickers*

The downwelling mid-IR radiance field was measured by a Fourier transform infrared (FTIR) spectroradiometer on the *John Vickers*. The radiances are recorded with 1 cm^{-1} resolution over the wavenumber interval $500\text{--}2000\text{ cm}^{-1}$ using a narrow-field-of-view detector aligned with the local zenith. The integrated downwelling flux in this spectral interval is estimated using limb-darkening models derived from coincident high-resolution atmospheric soundings [Lubin *et al.*, 1995]. The worst-case uncertainty in the flux estimates from errors in absolute radiometric calibration is approximately $\pm 6\text{ W/m}^2$. Infrared absorption bands outside the spectral band pass of the FTIR are virtually saturated in the tropical atmosphere. As a result, the contribution to $C_f(p_s)$ from wavenumbers outside the FTIR band pass may be neglected. The difference between downwelling radiances for a cirrostratus cloud and a clear-sky scene observed on March 11, 1993 (UTC), is shown in Figure 1 as a function of wavenumber. Between $500\text{--}700$ and $1300\text{--}1500\text{ cm}^{-1}$, the spectral cloud forcing is comparable to the noise equivalent spectrum of the instrument. The principal contribution to $C_f(p_s)$ is from the spectral cloud forcing between 700 and 1300 cm^{-1} , which is fully resolved by the FTIR.

The measurements used in this study cover $170^\circ\text{E}\text{--}161^\circ\text{W}$ at 2°S beginning on March 10, 1993, and ending March 17, 1993 (UTC). During daylight hours the measurements have been tagged with the cloud type in the field of view of the FTIR detector using the World Meteorological Organization (WMO) cloud classification. The cloud morphology was identified by visual inspection. In order to calculate the f factor, we have used the daytime downwelling fluxes corresponding to clear and cloudy conditions. Subvisible cirrus are not included in the cloud sample by definition, and in turn some of the clear data may contain subvisible cirrus. However, the errors introduced by optically thin cirrus in the estimates of f may be neglected (section 4).

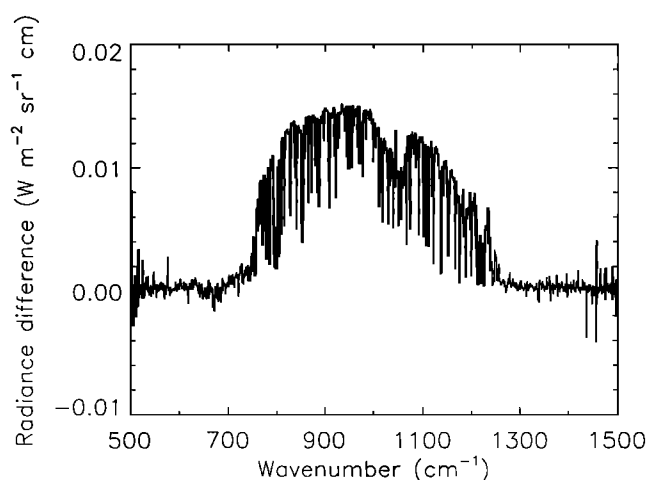


Figure 1. Difference between spectral radiance for a cirrostratus cloud and a clear-sky scene measured with the Fourier transform infrared (FTIR) instrument on the R/V *John Vickers*. The cirrostratus cloud was observed at 0435 UTC on March 11, 1993, at -2.09°S , 175.09°E , and the clear-sky scene was observed at 2130 UTC at -2.00°S , 178.20°E . At the time of the cloud observation the sea surface temperature was 302.6 K and the surface air temperature was 303.2 K .

A time series of the cloud types and meteorological conditions along the track of the *John Vickers* is shown in Figure 2a. The mean satellite brightness temperatures coincident with the observations are plotted in Figure 2b. The brightness temperatures are a measure of the narrowband thermal emission from the surface and clouds. The temperatures in Figure 2b have been averaged over a 6000 km^2 area centered on the ship. This area is equivalent to a TOGA COARE class 1 convective system and represents the median size and most frequent size of deep convective clouds observed during CEPEX [Collins *et al.*, 1996].

2.2. Radiometric Instrumentation on the NOAA P-3

A second set of fluxes were obtained within 100 m of the ocean surface using upward pointing Eppley pyrgeometers on the NOAA P-3. The hemispherically integrated downwelling flux is measured over the spectral interval $3.5\text{--}50\text{ }\mu\text{m}$. The pyrgeometers were calibrated at the start of CEPEX against a reference blackbody cavity over a temperature range of -56°C to 18°C . The nominal accuracy of the instantaneous flux measurements is approximately 15 W/m^2 . Fluxes measured by unfiltered upward and downward facing pyranometers are used for scene identification. The pyranometers were calibrated during CEPEX against a reference pyranometer supplied by NOAA. A nadir pointing narrow-field-of-view precision radiometer thermometer (PRT-5) thermistor is used to determine approximate sea surface temperatures. The spectral response of the thermistor lies in the mid-IR window between 9.5 and $11.5\text{ }\mu\text{m}$. A zenith pointing PRT-5 is used to estimate cloud-base altitude. The uncertainty of the PRT-5 measurements is approximately $\pm 0.5\text{ K}$.

The data used in this study are from four research flights in the CEPEX domain on March 15 and 25 and April 3 and 4, 1993 (UTC). These flights were selected for analysis using three criteria. First, radiometric data were collected at low altitudes over long distances; second, the aircraft flew over the region measured from the *John Vickers*; and third, the aircraft

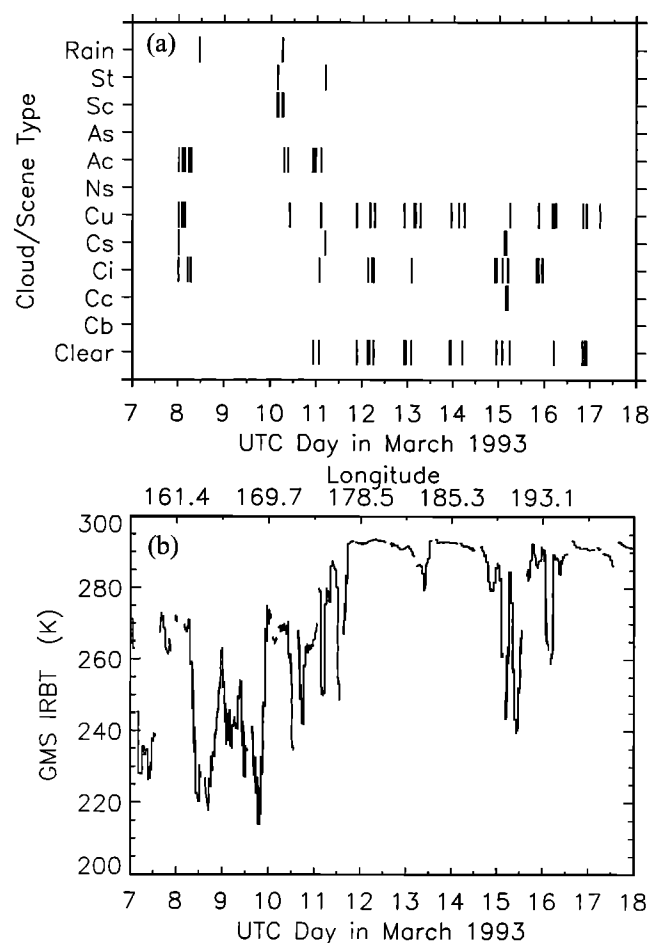


Figure 2. (a) Time series of World Meteorological Organization (WMO) cloud types observed along the zenith line of sight of the FTIR instrument at the time of FTIR observations. When two cloud types were observed simultaneously, both are plotted. (b) Time series of Geostationary Meteorological Satellite (GMS) brightness temperatures along the *John Vickers* track. The GMS brightness temperatures closest in time to each point along the track have been averaged over a 6000-km² region centered on the ship. The values along the upper abscissa are the longitudes (degrees east) of the *John Vickers* at the corresponding times on the lower abscissa.

flew beneath nonprecipitating clouds with low cloud-top temperatures. Additional measurements from research flights on March 14, 16, and 26, 1993, are combined with the data from these flights to parameterize the clear-sky surface net longwave flux in terms of the sea surface and air temperatures [Collins *et al.*, 1996]. The differences between the longwave cloud forcing at the altitude of the aircraft and the surface have been estimated from column-radiation model calculations using the Intercomparison of Radiation Codes in Climate Models (ICRCCM) tropical atmosphere. The difference between C_l at 0 and 50 m above sea level ranges from 0.3 to 2 W/m² for cloud-base pressures ranging from 170 to 950 mbar. These differences are small compared with the uncertainty in the instantaneous pyrgeometer measurements and will be neglected in the remainder of the analysis.

The f factor has been calculated using pyrgeometer data from periods of level flight less than 100 m above sea level. In a typical research flight during CEPEX these periods are pre-

ceded by excursions above the boundary layer accompanied by rapid changes in altitude. Inspection of the flux time series indicates that the pyrgeometers respond slowly to changes in ambient air temperature. The time required for the instruments to equilibrate after the vertical excursions is approximately 9 min (Figure 3). The concave curvature of the time series is caused by warming of the instrumentation and is reproduced on all four flights. Figure 3 also indicates that the equilibration time increases with the maximum altitude of the P-3 prior to the start of each low-level flight segment. Low-altitude measurements collected within 9 min of descent into the boundary layer or collected after descent from above 2000 m have been omitted from the analysis. The remaining low-altitude measurements should not be affected by the transient response of the pyrgeometers to changing air temperature.

The times, locations, and altitudes of the low-altitude observations used in this study are given in Table 1. The mean satellite brightness temperatures coincident with the observations are plotted as a function of time during each flight in Figure 4. The satellite measurements show that the low-altitude data include both clear and convectively disturbed conditions.

2.3. Radiometric Instrumentation on the NASA ER-2

The ER-2 was equipped with a Radiation Measurement System (RAMS) [Valero *et al.*, 1982] and a cloud-detection lidar [Spinhrne and Hart, 1990]. The upwelling longwave flux F^+ is measured using a hemispherical-field-of-view instrument with a spectral band pass spanning 4–40 μ m. The flux measurements are accurate to better than 1%. The upwelling radiances in the mid-IR window between 10 and 10.982 μ m from the ER-2 are used to help calibrate the GMS satellite radiometers. The upwelling radiances between 4 and 40 μ m from the

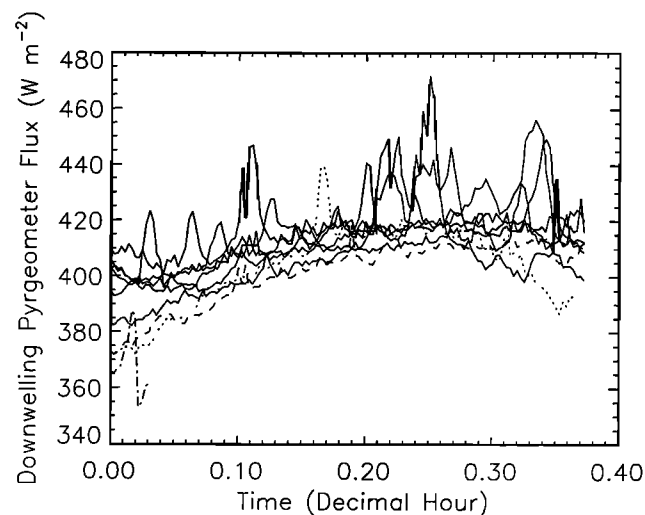


Figure 3. Downwelling flux from the upward pointing pyrgeometer on the National Oceanic and Atmospheric Administration (NOAA) P-3. The data were collected during the mission on March 25, 1993, when the P-3 was less than 100 m above the ocean surface. Time is measured relative to the starting time of each segment of the flight when the P-3 descended below 100 m. The maximum altitude $z_{\max}$ of the aircraft preceding each low-altitude flight segment is indicated by the type of line in the plot: solid lines, $z_{\max} \leq 2000$ m; dotted lines, $2000 < z_{\max} \leq 4000$ m; dashed lines, $4000 < z_{\max} \leq 6000$ m; and dash-dotted lines, $6000 < z_{\max} \leq 8000$ m.

Table 1. Dates, Starting and Ending Times, Locations, and Mean Altitudes of Low-Altitude Flight Segments Conducted With the NOAA P-3

Date	Start, UTC	Latitude	Longitude	End, UTC	Latitude	Longitude	Mean Altitude, m above sea level
March 15, 1993	1941	0°18'S	157°24'W	1945	0°36'S	157°25'W	33.1
	1959	1°32'S	157°28'W	2006	1°57'S	157°29'W	35.6
	2121	2°00'S	162°41'W	2133	2°00'S	163°28'W	34.6
	2147	2°00'S	164°26'W	2159	1°59'S	165°13'W	34.4
	2215	2°00'S	166°17'W	2223	1°59'S	166°46'W	35.4
	2235	1°59'S	167°36'W	2243	2°00'S	168°07'W	34.0
	2255	1°54'S	167°33'W	2306	2°05'S	166°55'W	67.6
March 25, 1993	1940	2°00'S	165°18'W	1959	1°59'S	166°31'W	33.5
	2014	2°00'S	167°32'W	2026	1°59'S	168°17'W	34.4
	2108	2°00'S	171°05'W	2120	2°00'S	171°51'W	34.5
	2138	2°00'S	173°02'W	2147	2°00'S	173°36'W	35.3
	2309	2°01'S	179°18'W	2320	2°00'S	179°59'E	35.0
April 3, 1993	2231	2°00'S	174°40'E	2245	1°59'S	173°50'E	55.0
	2301	1°59'S	172°50'E	2310	1°59'S	172°16'E	55.6
	2439	2°35'S	167°57'E	2440	2°39'S	168°00'E	59.4
	2454	2°14'S	167°34'E	2459	1°59'S	167°20'E	56.8
	2522	1°17'S	166°19'E	2544	1°15'S	164°56'E	55.7
April 4, 1993	2259	0°36'S	168°40'E	2307	1°00'S	168°20'E	54.8
	2325	0°32'S	167°56'E	2333	0°09'S	167°37'E	55.5
	2348	0°36'N	166°59'E	2356	1°00'N	166°39'E	56.4
	2413	0°29'N	166°14'E	2419	0°10'N	165°59'E	57.3
	2434	0°35'S	165°20'E	2442	0°59'S	165°00'E	55.7

ER-2 are used to relate mid-IR and broadband quantities (section 3.2). Data from two flights on March 20 and April 3, 1993, are not included in the development of the GMS OLR algorithm due to instrument malfunction.

Infrared images collected by the Moderate Resolution Imaging Spectroradiometer (MODIS) Airborne Simulator (MAS) on board the ER-2 [King *et al.*, 1996] are used to calibrate the GMS satellite. During CEPEX the MAS was configured to acquire high-resolution imagery in 12 spectral bands. The central wavelength of the channel used in this study is 11.002 μm , with a bandwidth of 0.448 μm . The MAS instrument completes 6.25 scans perpendicular to the ground track of the aircraft each second. The field of view across the scan is 85.92°. The pixel spatial resolution is 50 m at a distance of 20 km, the typical cruising altitude of the ER-2. MAS data from nine ER-2 flights have been used in the analysis. No MAS data are available for the ER-2 missions on March 15, 16, and 23, 1993.

The CEPEX MAS data have not been corrected for the effective emissivity of the MAS onboard blackbodies (C. Moeller and P. Hrubik, private communication, 1996, and Moeller *et al.* [1996]). The MAS data are used as an intermediate “transfer” calibration standard rather than an absolute calibration reference (section 3.1). The GMS data are calibrated by first relating GMS radiances to MAS radiances, then relating MAS radiances to RAMS measurements. Our application of MAS mid-IR data is valid in the absence of the corrections to account for blackbody emissivity because these corrections are nearly linear. Since the relationships between GMS and MAS and between MAS and RAMS brightness temperatures are also linear, the calibration of GMS against RAMS is not affected by the absence of the blackbody corrections.

2.4. Correspondence Between the Surface and Aircraft Longwave Fluxes

In order to compare longwave fluxes from the ER-2, John Vickers, and NOAA P-3, it is necessary to remove or minimize the effects of differing spectral band passes for the various

radiometers. For the purposes of computing cloud forcing, the longwave fluxes are assumed to span 3.3–500 μm with no contribution from the solar radiation between 3.3 and 5 μm . None of the radiometers deployed in CEPEX measure infrared wavelengths longer than 50 μm , and the ER-2 and P-3 instruments both measure some solar radiation in the near-infrared. Since the OLR is estimated from GMS with a technique based upon the ER-2 data, the spectral range of the GMS OLR is identical to the ER-2 measurements. The magnitude of the differences in the cloud forcing introduced by the spectral band pass of each instrument has been estimated from model calculations.

The spectral cloud forcing between 3 and 500 μm has been simulated for low, middle, and upper tropospheric clouds embedded in atmospheric profiles measured from the John Vickers [Weaver *et al.*, 1994; Kley *et al.*, 1997]. The calculations show that the differences between $C_l(p_s)$ in the band pass of the FTIR and a simultaneous broadband estimate of $C_l(p_s)$ (excluding solar radiation) are $0.63 \pm 0.45 \text{ W/m}^2$. The corresponding differences for the P-3 and ER-2 are 0.013 ± 0.012 and $0.37 \pm 0.1 \text{ W/m}^2$, respectively. These differences are much smaller than the uncertainty in the instantaneous measurements (sections 2.1, 2.2, and 2.3) and may be neglected.

The solar contribution to the longwave cloud forcing derived from ER-2 measurements has been estimated for liquid and ice-phase clouds. The contribution was computed using a 1690-spectral-band discrete-ordinate radiative transfer model [Zender *et al.*, 1997] applied to clouds embedded in the ICRCCM tropical reference atmosphere at nine solar zenith angles. The condensed water path for both types of cloud is 0.5 kg/m^2 . The solar contributions to $|C_l(0)|$ for the ER-2 RAMS instrument are 0.06 ± 0.04 and $0.08 \pm 0.03 \text{ W/m}^2$ for the liquid and ice-phase clouds, respectively. These corrections for the solar radiation are negligible compared with the uncertainty in the instantaneous RAMS measurements. At wavelengths longer than 4 μm , clouds are sufficiently absorptive that the cloud

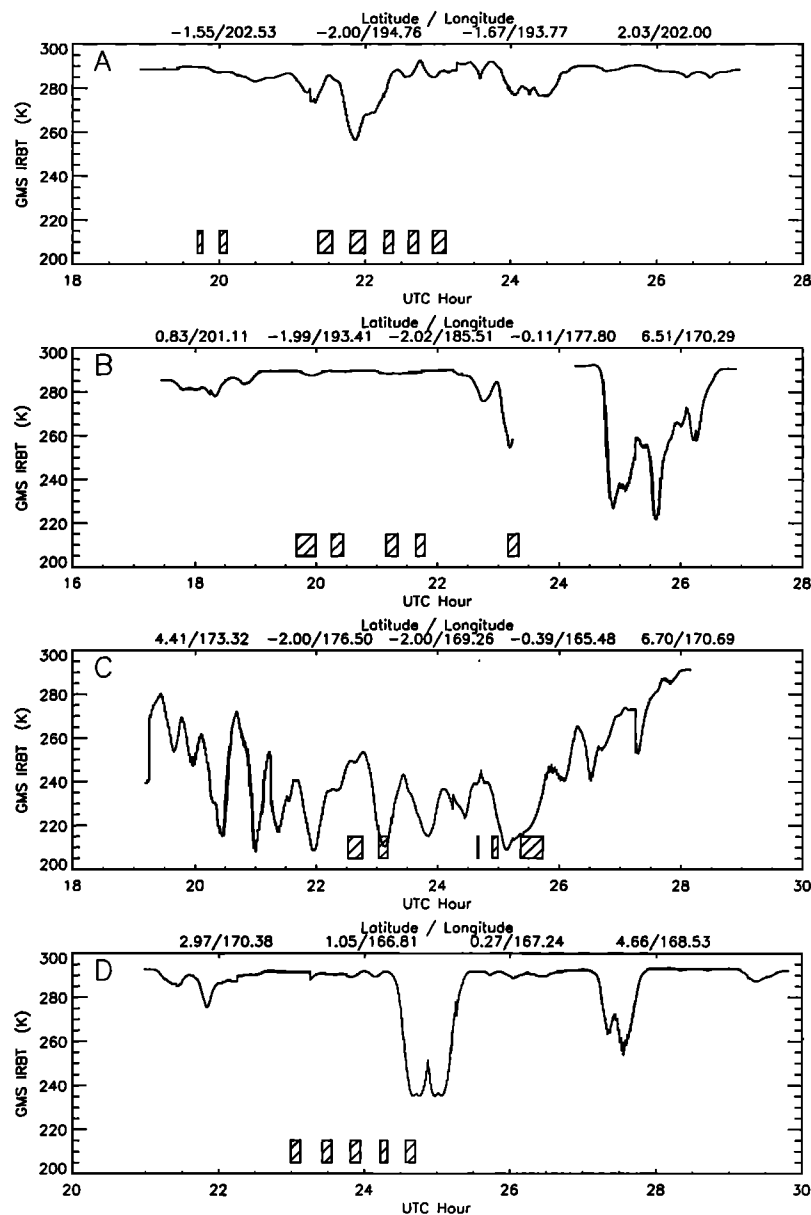


Figure 4. GMS brightness temperatures for each of the four NOAA P-3 flights with low-altitude flight segments listed in Table 1. The GMS brightness temperatures closest in time to each P-3 observation have been averaged over a 6000-km² region centered on the aircraft. The values along the upper abscissa are the latitudes (degrees north) and longitudes (degrees east) of the P-3 at the corresponding times on the lower abscissa. Boxes with hatching show duration of low-altitude flight segments in Table 1. Dates are (a) March 15, 1993 (UTC), (b) March 25, 1993, (c) April 3, 1993, and (d) April 4, 1993. Some GMS imagery is not available for March 25, 1993.

albedo cannot be distinguished from the albedo of the dark ocean background.

However, the effects of solar radiation on the P-3 data used to estimate longwave cloud forcing cannot be neglected. Clear-sky simulations with the tropical reference atmosphere show that the surface insolation in the band pass of the pyrgeometer can approach 7 W/m² for overhead Sun. This solar contribution is approximately half the 15 W/m² error in the instantaneous fluxes. Under cloudy conditions the surface insolation in the band pass of the pyrgeometers is less than 3 W/m² for cloud liquid-water paths (LWP) greater than 0.015 kg/m². Since this lower bound on LWP corresponds to an optically thin cloud,

the solar contribution will only be subtracted from clear-sky longwave measurements. The P-3 was equipped with a filtered pyranometer that measured the downwelling insolation between 0.695 and 3.8 μ m. The clear-sky simulations show that the insolation measured by the filtered pyranometer (S_f) and pyrgeometer (S_p) are highly correlated with $r = 0.99997$. To a good approximation,

$$S_p = 0.0144S_f. \quad (8)$$

The estimates of the solar contribution given by S_p are subtracted from the pyrgeometer measurements under clear-sky conditions (section 2.6).

2.5. Radiometric Instrumentation on the GMS 4 Satellite

The instrumentation on the GMS 4 satellite includes a spin-scan visible/infrared radiometer (SVISSR) that produces images of the Earth's disk centered at 0°N, 140°E. The narrow-band visible radiance field emitted by the atmosphere is measured with a nadir resolution of 1.25 km [Murayama, 1989; Rossow *et al.*, 1992]. The band pass of the instrument is 0.6–0.8 μm , and the radiance values have been normalized to the integrated solar emission in these wavelengths to give a narrowband reflectance. The absolute accuracy of the SVISSR visible data is uncertain because the instrument has not been calibrated since the launch of GMS 4. The SVISSR also measures radiances in the mid-IR window between 10 and 12 μm with a nadir resolution of 5 km. The radiances are converted to infrared brightness temperatures (IRBT) for an equivalent blackbody source. Comparison of coincident satellite and cloud lidar data shows that IRBTs greater than 290 K are frequently associated with clear-sky conditions, while IRBTs less than approximately 240 K indicate the presence of cold cloud tops in the upper troposphere. The GMS infrared radiometer is calibrated using onboard blackbody sources. The IRBTs have been corrected for a spurious diurnal cycle introduced by heating and cooling of the satellite during periods close to the solar equinox [Desormeaux *et al.*, 1993].

2.6. Detection of Clear-Sky Measurements Using Surface and Satellite Data

Clear-sky observations in the GMS data have been identified using spatial coherence analysis [Coakley and Bretherton, 1982; Coakley and Baldwin, 1984]. Since correlative information from both the visible and infrared channels is used to detect the presence of clouds, the identification of clear-sky scenes is limited to daytime imagery from the GMS satellite. The techniques of Coakley and Baldwin [1984] for constructing a cloud-free radiative climatology have been applied to each daytime GMS image during CEPEX to identify clear-sky pixels. The fractional area of clear regions detected with this method depends primarily on the size of the pixel arrays used to calculate the local variance of the satellite radiances. In order to determine the optimum size of these arrays, we have compared the spatial coherence analysis for 15 coincident advanced very high resolution radiometer (AVHRR) and GMS images. The spatial resolution of the AVHRR data is approximately 4 times the resolution of the GMS imagery. Therefore the analysis of AVHRR data can identify much smaller clouds and clear-sky regions than is possible with GMS. The highest correlation between the amounts of clear sky detected in AVHRR and GMS is obtained using 2×2 arrays of GMS pixels. The results in the remainder of this paper are based upon spatial coherence analysis with 2×2 pixel arrays.

The clear-sky fluxes $F_c^+(0)$ used to compute the TOA cloud forcing have been averaged on a 0.25° grid. Under convectively disturbed conditions, there are frequently no clear-sky satellite measurements coincident and collocated with the surface observations. In order to compute the TOA cloud forcing $C_f(0)$ corresponding to each surface measurement, the satellite clear-sky observations have been averaged for a 7-day period centered on the times of observation along the path of each surface platform. The variation in the clear-sky flux during the averaging period introduces an RMS uncertainty of less than 5 W/m^2 in the clear-sky TOA fluxes used to estimate $C_f(0)$.

Clear-sky measurements in the data from the NOAA P-3 have been identified using a modified version of the method

developed by Waliser *et al.* [1996]. In order for a P-3 data point to be classified as a clear-sky observation, two criteria must be satisfied. First, the GMS pixel closest in space and time must be flagged as cloud-free using spatial coherence analysis. Second, the change in the net solar flux relative to the previous or subsequent measurements must be less than the change in the top-of-atmosphere shortwave insolation, plus an empirically determined uncertainty. This second criterion eliminates observations that are probably associated with broken clouds [Waliser *et al.*, 1996]. The P-3 data that satisfy both conditions are identified as clear-sky observations.

3. Estimation of Longwave Flux From GMS Narrowband Radiances

The upwelling longwave flux at the tropopause has been estimated from satellite measurements of IRBT using an empirical relation between IRBT and flux. The relationship of OLR to this flux is derived from model calculations of the effects of the stratosphere. The retrieval of flux from IRBT is based upon the nearly linear relationship between upwelling mid-IR radiances R and corresponding upwelling broadband radiances I over the range of cloud-top temperatures characteristic of the troposphere [Minnis *et al.*, 1991]. The mid-IR radiances are related to IRBT by

$$R(T_c) = \int_0^\infty B_\lambda(T_c) \phi(\lambda) d\lambda, \quad (9)$$

where T_c is the IRBT, $B_\lambda(T_c)$ is the Planck function, λ is wavelength, and $\phi(\lambda)$ is the response function of the instrumental filter. The filter response function of the GMS 4 instrument is given by Rossow *et al.* [1992]. After correcting for the viewing geometry of the satellite observations, the estimates of T_c from GMS 4 have been calibrated against collocated measurements of T_c from the ER-2. The relations between IRBT, mid-IR radiance, broadband radiance, and broadband flux have been determined using simultaneous radiometric measurements of each quantity from the NASA ER-2. A schematic of the calibration procedure is shown in Figure 5. In the following discussion, T_c'' and R'' will denote the IRBT and corresponding radiance from GMS, T_c' will denote the IRBT measured by the MAS, and T_c and R will denote the IRBT and corresponding radiance measured by the RAMS instruments.

3.1. Calibration of GMS Radiances

The effects of the satellite viewing geometry generally reduce the mid-IR radiances measured by satellite instruments because the radiation travels along a slant path through the atmosphere to the satellite. This effect, known as limb darkening, generally increases with increasing satellite zenith angle θ , defined as the angle between local zenith and the vector to the satellite from the target region. The effects of limb darkening need to be minimized in order to calibrate the satellite against the nadir viewing aircraft radiometers. To first approximation the radiance $R''(0)$ along the zenith is related to the radiances $R''(\theta)$ along a slant path by

$$R''(0) = a(\theta) + b(\theta)R''(\theta). \quad (10)$$

The functions $a(\theta)$ and $b(\theta)$ have been determined from model calculations of the radiances emitted by black clouds at different levels in the atmosphere. The model is a version of

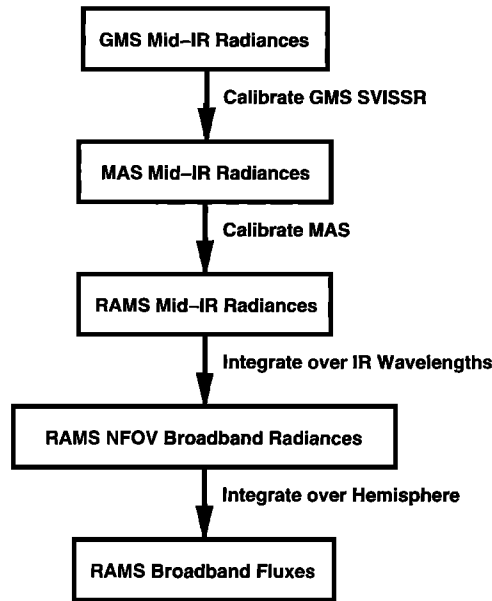


Figure 5. Steps used in the procedure to estimate broadband hemispherical fluxes from the GMS infrared radiances.

the Lowtran 7 code with 20 cm^{-1} spectral bands [Kneizys *et al.*, 1988]. The mean vertical resolution of the model layers is 20 mbar. Absorption by the water vapor continuum and present-day distributions of carbon dioxide, methane, and nitrous oxide have been included. The calculations have been performed using radiosonde profiles of temperature and humidity obtained during CEPEX to specify the background atmosphere [Weaver *et al.*, 1994]. The black clouds have been simulated by inserting surfaces of unit spectral emissivity at each altitude reported in the soundings. The TOA radiances have been determined by seven different values of θ . The limb-darkening functions $a(\theta)$ and $b(\theta)$ have been derived by least squares regression of $R''(\theta)$ and $R''(0)$. The functions may be accurately approximated using low-order polynomials:

$$a(\theta) = \sum_{k=0}^3 a_k [1 - \cos(\theta)]^k \quad (11)$$

$$b(\theta) = \sum_{k=0}^3 b_k [1 - \cos(\theta)]^k,$$

where

$$\begin{aligned} a_0 &= 0.0011 & b_0 &= 0.9997 \\ a_1 &= -1.0362 & b_1 &= 0.1868 \\ a_2 &= 0.0415 & b_2 &= -0.0562 \\ a_3 &= -2.2212 & b_3 &= 0.4867. \end{aligned} \quad (12)$$

The variation of $a(\theta)$, $b(\theta)$, and the linear least squares correlation coefficient $r(\theta)$ with θ is shown in Figure 6. As expected, the limb correction is small when the satellite is viewing close to the nadir direction and increases with larger θ . For black clouds in the middle and upper troposphere the model calculations indicate that the effects of limb darkening are negligible. The correction is not applied when $T_e'' < 260 \text{ K}$, which is approximately the value of T_e'' where $R''(0) \approx R''(\theta)$.

The calibration of the GMS IRBT against the IRBT measured by the aircraft instruments is performed in two stages (Figure 5). In the first stage the IRBT measured by the SVISSR instrument on GMS is compared with the IRBT measured by the MAS. The MAS IRBTs are averaged over the field of view of the SVISSR instrument projected onto the ocean surface (the GMS pixel “footprint”). This process minimizes the uncertainty introduced by tremendous disparity between the fields of view of the satellite instruments and the nadir pointing RAMS radiometers on the ER-2. Attempts to calibrate GMS directly against the RAMS data were plagued by large RMS uncertainties arising from geolocation errors in the satellite data and from subpixel variability detected by the RAMS instrument. The infrared channels on MAS were not sufficiently well characterized during CEPEX to serve as an absolute calibration standard for GMS, although subsequent intercomparisons between MAS and hyperspectral infrared spectrometers indicate that the MAS IRBTs are accurate to approximately 1 K for warm scenes and 5 K for cold scenes [Moeller *et al.*, 1996]. Instead, the MAS is used as an intermediate transfer calibration standard between the GMS imager and the RAMS narrow-field-of-view (NFOV) instrument.

The relation between instantaneous collocated averages of the GMS and MAS IRBT (T_e'' and T_e' , respectively) from nine flights of the ER-2 is shown in Figure 7. The least squares linear fit to this relationship is

$$T_e' = 200.0 + 1.052(T_e'' - 200.0) - 3.04. \quad (13)$$

In the second stage the nadir radiances measured by the MAS scanner and the NFOV RAMS mid-IR radiometer are intercalibrated. A scatterplot of the IRBT derived from the radiances is shown in Figure 8. Each point represents an average of the measurements over 25 s to reduce the noise in the relationship. The magnitudes of the offsets between RAMS and MAS for cold and warm scenes are consistent with Moeller *et al.* [1996]. The least squares linear fit to the scatterplot is

$$T_e = 200.0 + 0.944(T_e' - 200.0) + 5.34. \quad (14)$$

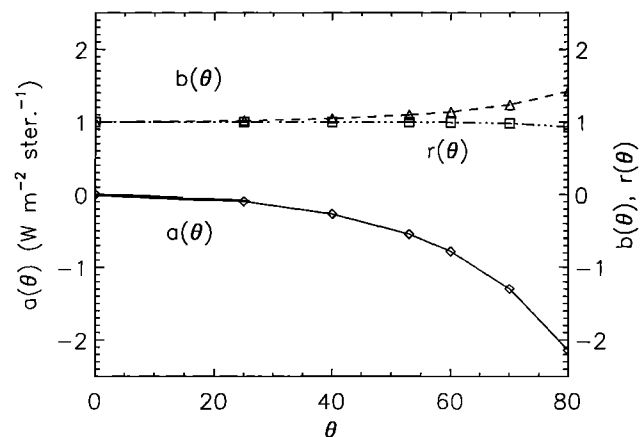


Figure 6. Limb-darkening functions $a(\theta)$ and $b(\theta)$ defined by equation (10) and the linear correlation coefficient $r(\theta)$. Here $a(\theta)$ and $b(\theta)$ are derived from the least squares regression of $R''(\theta)$ against $R''(0)$. The functions are determined at satellite viewing angles $\theta = 0^\circ, 25^\circ, 40^\circ, 53^\circ, 60^\circ$, and 70° from calculations based upon soundings launched from the John Vickers.

The combination of (13) and (14) shows that the GMS infrared radiometer underestimates the IRBT measured by the RAMS mid-IR radiometer by approximately 1.5–2.5 K. The offset in IRBT is explained in part by the greater width of the SVISSR filter response function compared with those of the MAS and RAMS instruments.

3.2. Estimation of Broadband Fluxes From Narrowband Radiances

The relation between the narrowband radiances $R(T_e)$ and coincident broadband radiances I is shown in Figure 9. The figure includes measurements from 10 ER-2 flights during CEPEX. The relation between the narrowband and broadband radiances is approximately

$$I = 23.68 + 9.108R(T_e) - 0.1251R^2(T_e). \quad (15)$$

The quadratic term changes I by less than 10% over the range of R characteristic of the tropical troposphere.

The upwelling longwave flux at the ER-2 altitude is related to the nadir longwave intensity by

$$F = \pi I. \quad (16)$$

The term πI is introduced by the integration over solid angle, and the factor I accounts for limb-darkening effects. For emission from a surface with unit emissivity and no intervening absorption between the detector and the surface, I is equal to 1. Effects of tropospheric absorption reduce I to values slightly less than 1, and I is moderately sensitive to changes in the vertical profile of atmospheric and cloud optical depth. The value of I used to retrieve fluxes at the ER-2 altitude is given by

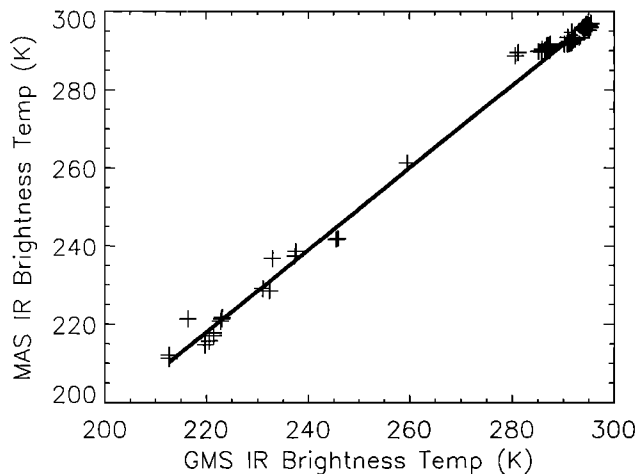


Figure 7. Relation of coincident Moderate Resolution Imaging Spectroradiometer (MODIS) Airborne Simulator (MAS) 11- μm IR brightness temperature (IRBT) and GMS IRBT for Central Equatorial Pacific Experiment (CEPEX) ER-2 flights with valid MAS data. The IRBT from both instruments has been averaged over 3×3 arrays of GMS pixels centered on the ER-2. Each symbol represents the average for a single array. The MAS pixels included in the average are selected assuming a 5-km diameter for GMS IR pixels (the nadir resolution). Points within 900 s of exact temporal coincidence between GMS and MAS have been included. The GMS IRBT has been corrected for effects of limb darkening using equations (10)–(12). The solid line is the least squares linear fit given by equation (13) with a linear correlation coefficient of 0.998.

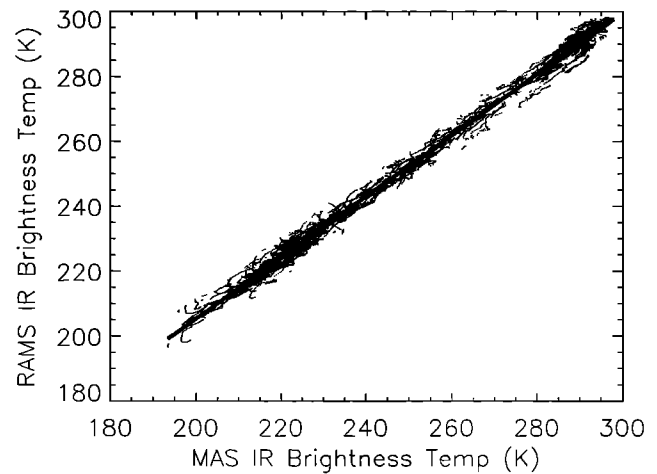


Figure 8. Relation of coincident MAS 11- μm and Radiation Measurement System (RAMS) 10.5- μm IRBT. Each point represents an average of the nadir MAS pixels and measurements from the nadir pointing RAMS narrow-field-of-view radiometer over 5 km (approximately 25 s of flight time). Data from the same flights used in Figure 7 have been included. The solid line is the least squares linear fit given by equation (14) with a linear correlation coefficient of 0.998. The RAMS data have been shifted, on average, by 8 s relative to the MAS data to maximize the correlation. The fit was computed iteratively by excluding measurements more than 3σ from the linear fit while $\sigma \geq 2\text{K}$. Two percent of the original sample has been excluded.

$I =$

$$\begin{cases} 0.847 & T_e < 285 \\ 0.847 + (0.893 - 0.847)(T_e - 285)/10 & 285 \leq T_e \leq 295 \\ 0.893 & T_e > 295. \end{cases} \quad (17)$$

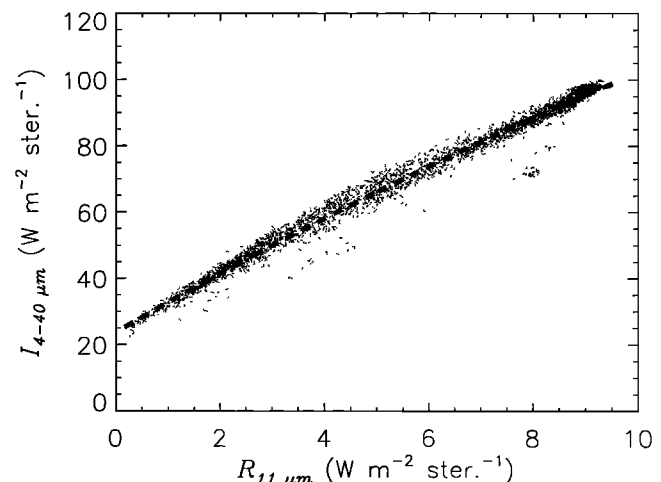


Figure 9. Longwave radiance I (4–40 μm) as a function of narrowband radiance $R(T_e)$ for 10 CEPEX ER-2 flights. $R(T_e)$ is calculated from measurements of T_e coincident with I using equation (9). The quadratic least squares regression of I against R given by equation (15) is shown by the dashed line.

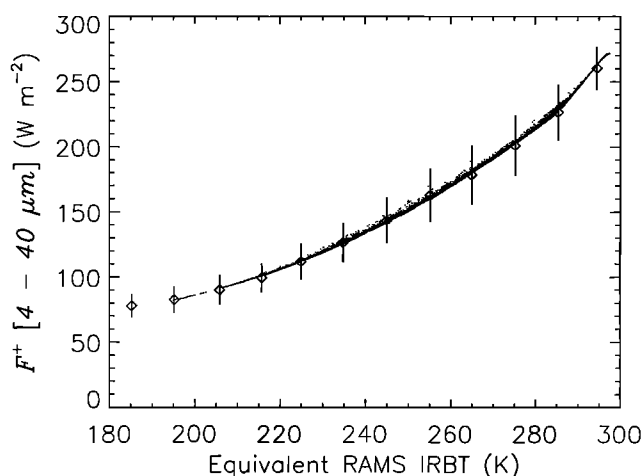


Figure 10. Upwelling longwave flux F ($4\text{--}40\ \mu\text{m}$) as a function of IRBT at the ER-2 altitude. Each point is an average over a 9×9 pixel region centered on the location of the ER-2 aircraft for satellite observations collocated with 10 CEPEX ER-2 flights. The equivalent RAMS IRBT has been estimated from the GMS IRBT using equations (13) and (14). The diamonds show the mean value of F from aircraft observations averaged over 10-K intervals in IRBT. The error bars indicate the $1\text{-}\sigma$ range of the instantaneous values of F falling within each IRBT bin.

The lower value of l is determined from the ratio of the mean flux to mean intensity averaged over all RAMS measurements with $T_e < 285$. This value of l is applied to most of the optically thick clouds since the IRBT for these clouds is generally much less than 285 K. The upper value of l is determined from the ratio of the mean flux to mean intensity for clear-sky scenes. The clear-sky scenes have been identified by combining data from the cloud-detection lidar and the shortwave radiometer on the ER-2 [Collins *et al.*, 1996]. The variation of l with T_e given in (17) closely approximates the variation of l with T_e derived from ER-2 measurements (not shown).

3.3. Error Estimates for Broadband Fluxes

The retrieval of longwave flux F from GMS IRBT in the band pass of the ER-2 radiometers is completely specified by (10)–(17). The value of the flux is appropriate for the average cruising altitude of the ER-2. The relation of IRBT to the retrieved flux is shown in Figure 10 for satellite observations coincident with the CEPEX aircraft observations. The variation of F with T_e obtained directly from the aircraft radiometers is shown for comparison. The aircraft fluxes have been averaged over 10-K intervals in T_e to clarify the comparison. The difference between the retrieved and measured values of F (averaged over each 10-K interval) is less than $4\ \text{W/m}^2$. Within each 10-K interval the RMS difference between the instantaneous RAMS fluxes and the mean RAMS flux is less than $\sigma_{\max}(F) = 23\ \text{W/m}^2$. Since the retrieval is uniquely determined by T_e , $\sigma_{\max}(F)$ is an upper estimate of the RMS errors in the instantaneous fluxes retrieved from GMS. The value of $\sigma_{\max}(F)$ includes effects of inhomogeneities in the cloud field (e.g., cloud boundaries) on scales much larger than a GMS IR pixel. An analysis of the MAS IR imagery from CEPEX shows that the mean variance in IRBT within a single GMS pixel is 2.5 K. If the variance in IRBT is limited to twice this value, or 5 K, over the field of view of the MAS instrument

(approximately 20 km in radius), the RMS difference between the instantaneous and mean RAMS fluxes is less than $\sigma(F) = 10\ \text{W/m}^2$. This value is adopted as the RMS error in the GMS retrieval in the analysis that follows.

The algorithm does not include a correction for variations in water vapor absorption that are not detected by the narrow-band satellite radiometer. Corrections of this type have been implemented using the $6.7\text{-}\mu\text{m}$ water vapor channel on geostationary satellites [Chen *et al.*, 1991], but unfortunately, the GMS 4 imager does not include a $6.7\text{-}\mu\text{m}$ channel. Corrections based upon gridded analyses of atmospheric humidity do not substantially reduce the RMS errors between the empirical estimates and simultaneous measurements of OLR [Minnis *et al.*, 1991]. The omission of a correction for variations in humidity should primarily affect the higher values of OLR, which are generally associated with longer atmospheric path lengths.

In order to estimate TOA fluxes $F^+(0)$ from F , the contributions to the flux from the stratosphere and from wavelengths outside the band pass of the ER-2 instruments must be included. An empirical relationship between F and $F^+(0)$ has been obtained from simulations of clear-sky and cloudy scenes using Lowtran 7 applied to CEPEX radiosonde profiles (section 2.4). Uncertainties in boundary-layer and stratospheric humidity introduce uncertainties of 1 and $4\ \text{W/m}^2$ in $F^+(0)$, respectively [Weaver *et al.*, 1994; Collins and Inamdar, 1995]. The latter uncertainty is particularly relevant for estimating the OLR from measurements at the tropopause. The calculations show that $F^+(0)$ and F are highly correlated, with

$$F^+(0) = 0.981F + 20.96. \quad (18)$$

The all-sky fluxes $F^+(0)$ used to compute the TOA cloud forcing have been averaged on the same 0.25° grid as the clear-sky fluxes. If the method for deriving longwave flux is applied to other geographic regions or other time periods, the bias and RMS errors may be larger. This is a common disadvantage of techniques based upon narrowband satellite measurements [Minnis *et al.*, 1991].

3.4. Comparison of OLR Estimates

Independent estimates for OLR have been developed for TOGA COARE using algorithms developed by Minnis *et al.* [1991] and Zhang *et al.* [1995]. Broadband measurements are available from the wide-field-of-view (WFOV) radiometers on the ERBS satellite for COARE and CEPEX. In order to test the present procedure for calculating OLR, the procedure has been applied to the TOGA COARE region and the results compared with the other estimates. Time and area-mean values for the COARE Intensive Observing Period (IOP) are given in Table 2. The IOP averages span November 1, 1992, to February 28, 1993. The area averages are for the large scale array (LSA) (10°N – 10°S , 135°E – 175°E), the outer sounding array (OSA) (10°N – 10°S , 147.5°E – 167.5°E), and the intensive flux array (IFA) (0°N – 5°S , 152.5°E – 157.5°E). Please note that the NASA and ISCCP estimates are based upon interim versions of these data sets.

The results show that the present algorithm is consistent with other estimates to better than $10\ \text{W/m}^2$ for daily-mean area averages. The RMS differences between the daily-mean OLR from the present scheme and the International Satellite Cloud Climatology Project (ISCCP) FC data set range from $2.5\ \text{W/m}^2$ for the LSA to $6.5\ \text{W/m}^2$ for the IFA. The instantaneous RMS differences are $15.4\ \text{W/m}^2$. In order to determine

what fractions of these differences are explained by distinctions between the algorithms and spatiotemporal sampling, the method based upon ER-2 measurements has been applied to the same ISCCP DX pixel data used to construct the ISCCP FC OLR. The ISCCP DX pixel data set is subsampled to 30-km resolution with 3-hourly intervals between successive images. Comparison of the ISCCP fluxes and the fluxes from our full and reduced-resolution data sets shows that 62–68% of the flux offsets are explained by differences between the ISCCP methodology and the present algorithms. The larger differences between the ERBE and other values for the IFA appear to be related to sampling, however. When the National Center for Atmospheric Research (NCAR) and NASA data sets are sampled at the same times as the ERBS overpasses of the IFA, the fluxes from these data sets are 202 and 203 W/m^2 , respectively. These findings suggest that the frequency of convection in the ERBS sample over the IFA is larger than the true mean frequency for the IOP.

4. Determination of the f Factor

The f factor has been computed using (1)–(3) by collocating the estimates of $C_l(0)$ from the GMS 4 satellite with $C_l(p_s)$ from the instruments on the *John Vickers* and the NOAA P-3. The relation of f derived from the FTIR data to $C_l(0)$ is shown in Figure 11. Each point in the figure corresponds to a single cloud observation. The errors in the estimates of f are calculated assuming a 6 W/m^2 uncertainty in $C_l(p_s)$ (section 2.1) and a 10 W/m^2 uncertainty in $F^+(0)$ and $F_c^+(0)$ (section 2.5). The measurements for middle- and high-level clouds with no low clouds present are identified using the WMO cloud types plotted in Figure 2. The midlevel clouds include nimbostratus, altostratus, and altocumulus, and the high-level clouds include cirrostratus, cirrus, or cirrocumulus. The remaining measurements have low-level clouds, specifically stratocumulus, cumulus, or stratus clouds, either in isolation or in combination with clouds at higher altitudes. The results indicate that the instantaneous values of f for the cloud types associated with convection are generally less than 0.25, although the estimates for optically thin clouds are highly uncertain. The mean value of f for the midlevel and high-level clouds is 0.13. The mean is calculated as the ratio of the average $C_l(p_s)$ to average $C_l(0)$. The mean value of f for all clouds measured by the FTIR is 0.27.

The relation between $C_l(0)$ and f derived from the pyrge-

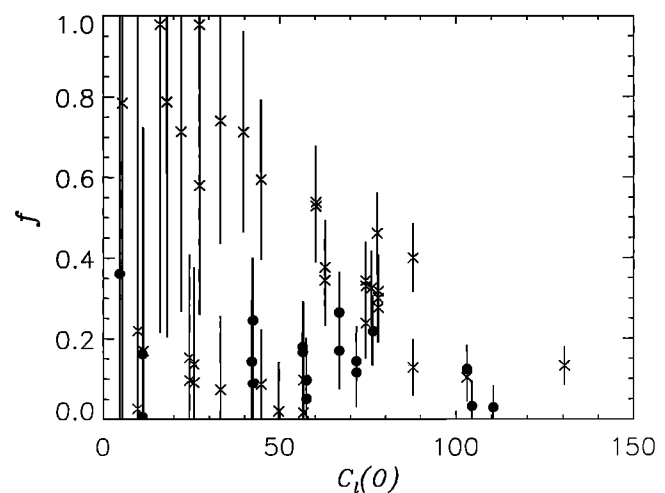


Figure 11. The f factor computed from collocated FTIR and GMS fluxes as a function of $C_l(0)$ at the location of the FTIR measurement. Circles denote observations corresponding to midaltitude clouds (WMO types Ns, As, and Ac) and high-altitude clouds (WMO types Cs, Ci, Cc). Crosses denote observations with low clouds in the field of view (WMO types St, Sc, or Cu).

ometers on the NOAA P-3 is shown in Figure 12. Each point in the figure represents a 5-min average from the low-altitude flight legs listed in Table 1. The errors in the estimates of f are calculated assuming a 6.4 W/m^2 uncertainty in $F_c^-(p_s)$ [Collins *et al.*, 1996], a 15 W/m^2 uncertainty in $F^-(p_s)$ (section 2.2), and a 10 W/m^2 uncertainty in $F^+(0)$ and $F_c^+(0)$ (section 2.5). Because a cloud classification is not available for the P-3 data, the observations associated with convective cloud systems have been determined from GMS imagery. The clouds in each GMS infrared image have been identified using a cloud classification algorithm [Boer and Ramanathan, 1997]. The procedure classifies each connected region with IRBTs less than 285 K into one of eight different cloud types based upon the minimum

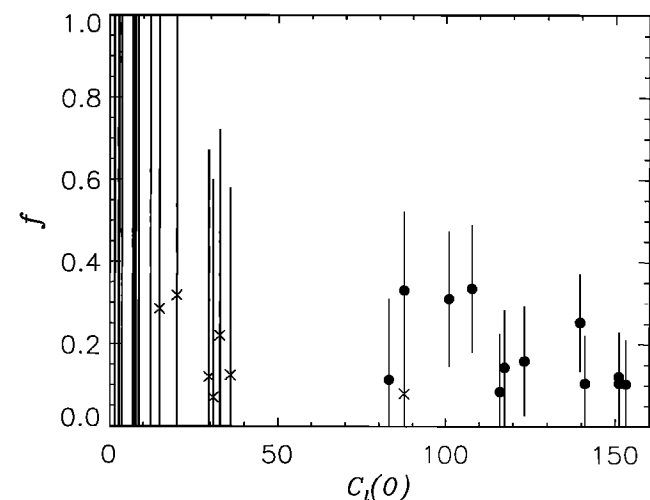


Figure 12. The f factor calculated from measurements from the low-altitude flight segments listed in Table 1. The points represent averages over 300 s. Circles denote observations underneath deep convective clouds identified in the nearest GMS image.

Table 2. Time and Area-Averaged Values of Outgoing Longwave Radiation for the COARE Intensive Observing Period

Data Set	Large Scale Array	Outer Sounding Array	Intensive Flux Array
ISCCP	220	219	215
NASA	226	226	222
NCAR	223	224	220
ERBE	218	218	208
σ	2.5	3.1	6.5

The ISCCP, NASA, and NCAR values are derived following Zhang *et al.* [1995], Minnis *et al.* [1991], and the present paper, respectively. ERBE values are from wide-field-of-view radiometers on ERBS. Units are W/m^2 . Here σ denotes the RMS differences between area-averaged daily-mean outgoing longwave radiation from the ISCCP and NCAR data sets. The NASA and ISCCP values are from interim versions of these data sets.

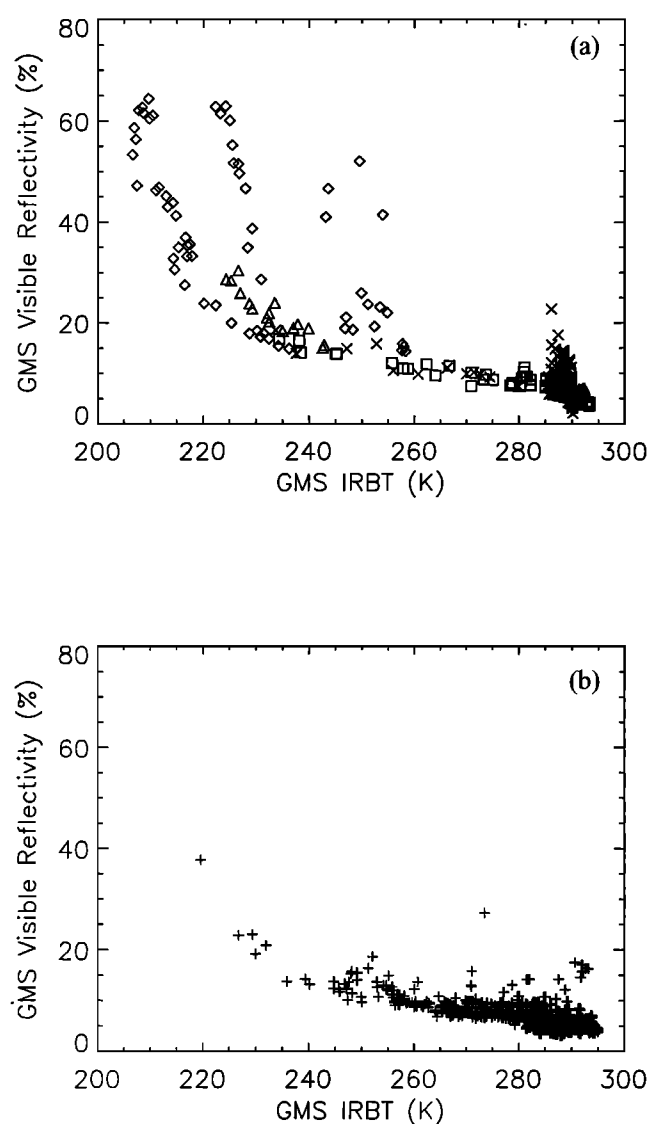


Figure 13. (a) Cloud-top GMS reflectance and IRBT for the low-level flight segments in Table 1. Equivalent visible and infrared radiances for the Sun and satellite positioned along the local zenith are plotted. Individual flights are denoted by symbols: squares, March 15; crosses, March 25; diamonds, April 3; and triangles, April 4. (b) The GMS reflectances and IRBTs for cirrus clouds above 11 km detected by the ER-2 cloud lidar. The visible and infrared radiances are corrected to the same Sun-satellite geometry as in Figure 13a. Points are selected by requiring that the cloud-base and cloud-top altitudes for all cloud layers exceed 11 km for 5 km (the radius of a GMS pixel on the surface) along the ER-2 ground track. The data are from 10 CEPEX ER-2 flights.

IRBT within the region. Brightness temperatures less than 240 K are often associated with convective systems [Laing and Fritsch, 1993]. The P-3 observations collocated with clouds with a minimum IRBT below 240 K are indicated in Figure 12. The average value of f for these clouds is 0.17.

There is also indirect satellite evidence that the P-3 observations include measurements of upper level clouds. The visible and infrared satellite radiances closest to the low-altitude P-3 observations are plotted in Figure 13a. The IRBTs have been corrected for the effects of limb darkening, and the vis-

ible radiances have been converted to equivalent values for the Sun and satellite positioned along the local zenith using the GMS angular dependence models [Boer and Ramanathan, 1997]. The normalization of the radiances facilitates combining observations from different geographic regions and different times of day. In Figure 13b the GMS radiances for upper level clouds identified by the cloud lidar on the NASA ER-2 are plotted for comparison. The temperature at the base of these clouds is below the threshold for the homogeneous nucleation of ice, and therefore the cloud sample is composed primarily of cirrus. The similarity of the satellite radiances in Figures 13a and 13b is consistent with the presence of high-altitude cirrus clouds in the field of view of the P-3 instrumentation.

Lower bounds on cloud-base and cloud-top altitudes may also be inferred from mid-IR radiance measurements from the P-3 and GMS satellite. The mean mid-IR brightness temperatures from the zenith pointing PRT-5 and the GMS imager for each data point shown in Figure 12 are plotted in Figure 14. For comparison, the PRT-5 brightness temperatures have been calculated for clouds with cloud-base altitudes of 1.5, 4, and 7 km inserted into the radiosonde profiles from the *John Vickers*. The mean and RMS brightness temperatures are 289.6 ± 0.6 , 282.2 ± 0.9 , and 263.6 ± 3.3 K, respectively. Clear-sky scenes had a mean brightness temperature of 249.6 ± 5.8 K, with 97% of the brightness temperatures below 263 K. Similar results were obtained from calculations based upon profiles extracted from the European Centre for Medium-Range Weather Forecasts (ECMWF) analysis data set collocated with the P-3. These calculations and Figures 13 and 14 indicate that the clouds that were not classified as convective are probably thin cirrus. This follows from the fact that the PRT-5 could not reliably distinguish these clouds from clear-sky scenes together with the low visible reflectivities and high IRBTs measured by the GMS satellite. However, Figure 14 shows that some of the bases of convective clouds measured by the P-3 extended to within a few kilometers of the ocean surface. This is confirmed by three-dimensional radar profiles of the clouds taken on several of the flights (P. Willis, private communication, 1994). The cloud-base altitude for most clouds was greater than or approximately equal to 4 km.

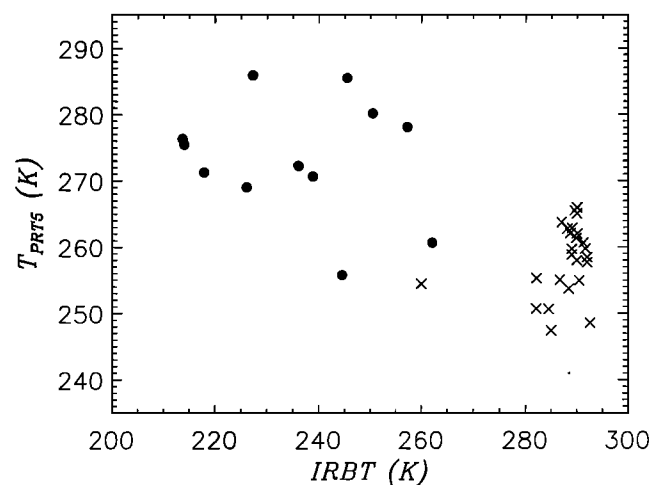


Figure 14. Mean GMS and PRT-5 brightness temperatures for each observation plotted in Figure 12. The points represent averages over 300 s. Circles denote observations underneath deep convective clouds identified in the nearest GMS image.

Table 3. Average Cloud Longwave Forcing for March 7 to April 7, 1993, Between 160°E and 160°W for Three Ranges in Latitude

Latitude Range	$C_l(0)$, $T_e < 240$ K	A_c , $T_e < 240$ K	$C_l(0)$ Mean
5°S–5°N	101	38.5	57
10°S–10°N	101	39.2	59
20°S–20°N	97	29.6	48

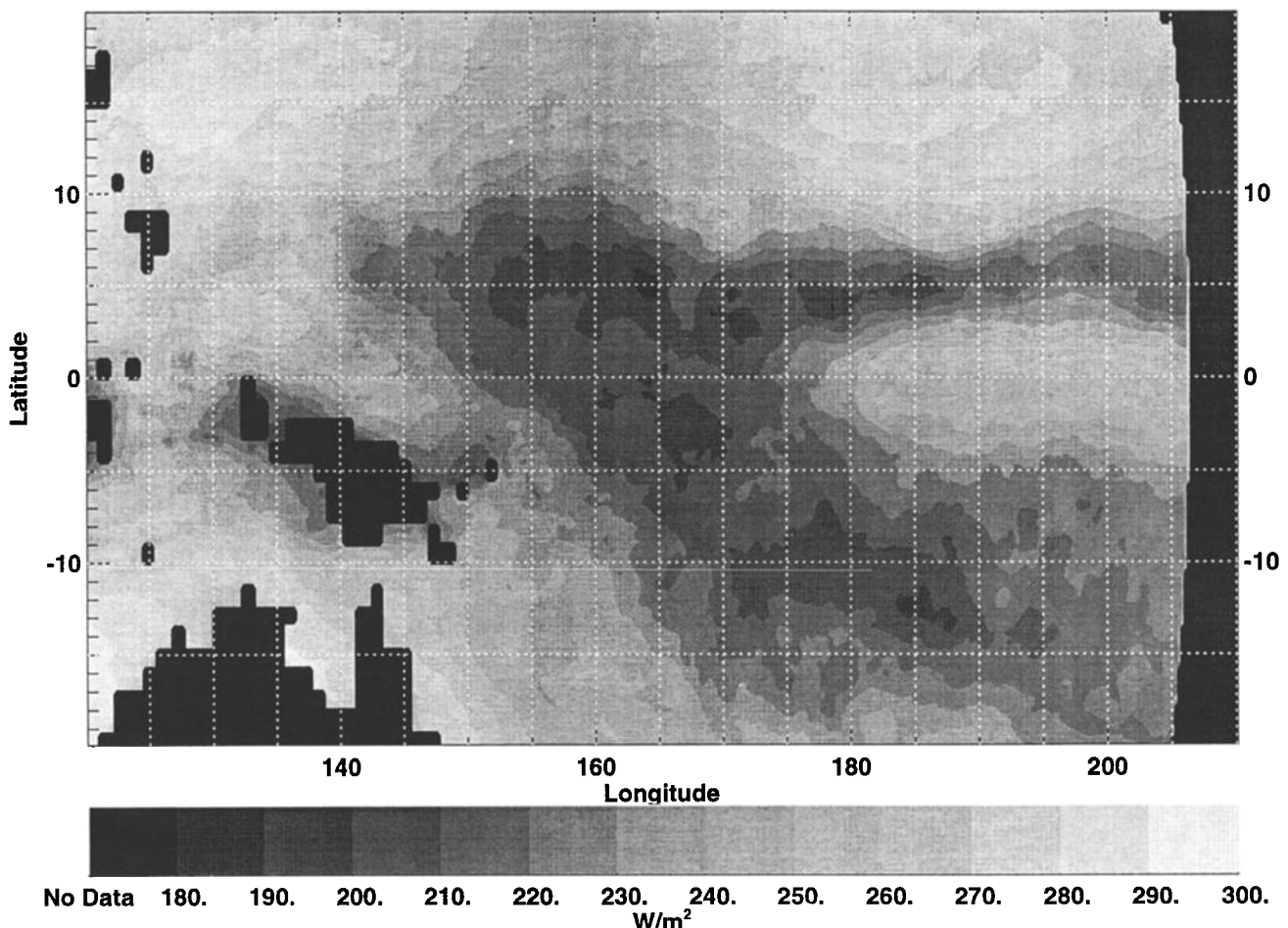
The two middle columns are estimates for clouds with minimum IR brightness temperatures below 240 K. The A_c column is the percentage of each region covered by convective cloud systems. The cloud forcing is in units of W/m^2 .

5. Discussion and Conclusions

The estimates of f from the in situ measurements may be used to estimate the radiative effects of convective cloud systems on the ocean surface over the central equatorial Pacific. The monthly-mean cloud forcing derived from the GMS data is given in Table 3. The cloud forcing by convective cloud systems identified in the GMS imagery represents the largest component in the regional mean $C_l(0)$. For example, in the latitude range 10°S–10°N, the area-mean $C_l(0)$ associated with convective systems is 40 W/m^2 , approximately 67% of the mean value including all clouds. The dominant contribution by deep

convective systems to the longwave forcing is related to the rapid decrease in outgoing longwave radiation (OLR) with the effective radiating temperature of the cloud tops. Since the mean of f for the middle and upper level clouds is $f = 0.15 \pm 0.02$, the CEPEX results suggest that the change in the downwelling longwave flux at the ocean surface due to these clouds is approximately 6 W/m^2 . This represents the smallest radiative component of the surface energy budget [Collins *et al.*, 1996] and is less than the indirect effects of convection on the greenhouse effect of the atmosphere at the ocean surface [Lubin, 1994].

The relation of the CEPEX observing period to the climatological conditions in the central Pacific may be examined using the longwave fluxes derived from the GMS data. A map of the mean outgoing longwave radiation for the CEPEX observing period is shown in Figure 15. The map indicates that there was significant convective activity in a zonal band centered at approximately 5°N and in a region aligned along a northwest-southeast axis that intersects the zonal band west of the date line. These regions correspond to the Intertropical Convergence Zone (ITCZ) and the Southern Tropical Convergence Zone (STCZ) observed in long-term satellite climatologies of OLR [e.g., Philander, 1990]. The mean SST and longwave cloud forcing for the CEPEX observing period are compared with the mean values of these quantities for March 1985 to March 1989 in Figures 16a and 16b. The SSTs are

**Figure 15.** Average outgoing longwave radiation (OLR) (in W/m^2) from March 5, 1993, to April 5, 1993 derived from GMS data.

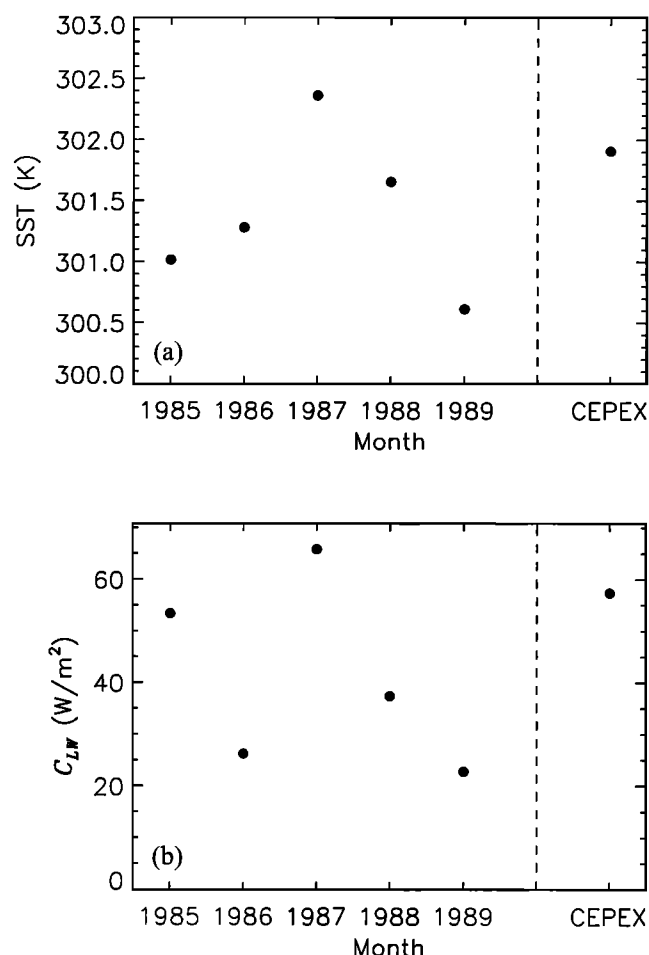


Figure 16. (a) Average National Meteorological Center sea surface temperature (SST) [Reynolds, 1988] for the months of March of 1985–1989 and for the CEPEX time period. The SSTs are the mean values for the region 5°S–5°N, 160°E–160°W. (b) Cloud longwave forcing for the months of March of 1985–1989 from ERBE and for the CEPEX time period derived from GMS. The values are averages for the same region as in Figure 16a.

taken from the blended analysis of Reynolds [1988], and the cloud forcing is derived from the monthly mean Earth Radiation Budget Experiment (ERBE) data set [Barkstrom and Smith, 1986]. The values are averaged over the equatorial region 5°S–5°N, 160°E–160°W. Almost all the surface measurements used to estimate the f factor were collected in this region. Figure 16 shows that the SST and cloud radiative effects were similar to those in March 1987 during the 1986–1987 El Niño. The similarity is not surprising, since CEPEX was conducted during the most recent El Niño episode, lasting from 1991 to 1995 [Trenberth and Hoar, 1996]. These results suggest that the cloud distributions and cloud radiative effects during CEPEX are comparable to the previous period of elevated SSTs in the central Pacific.

A new method for deriving OLR from operational satellite measurements has been developed and validated using radiometric measurements from an aircraft in the lower stratosphere. The advantage of this approach is that the spectral and angular integrations involved in deriving a broadband flux from narrowband satellite data may be checked using multiple

instruments on the same platform. Unlike methods based solely on satellites, the present method has been validated against direct, calibrated measurements of the upwelling broadband hemispherical flux. Future applications of this technique will include validation of the radiation budgets computed by GCM simulations of the CEPEX and TOGA COARE experiments.

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