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Summary of Completed Work

Work that has been completed under the existing grant has resulted in the following 12 publications:

1. M. H. Zhang, R. D. Cess and X. D. Jing, Concerning the interpretation of enhanced cloud shortwave absorption using monthly-mean earth radiation budget experiment/global energy balance archive measurements, *J. Geophys. Res.-Atmos.* **102**, pp. 25899-25905, 1997.
2. F. P. J. Valero, R. D. Cess, M. H. Zhang, et. al., Absorption of solar radiation by the cloudy atmosphere: Interpretations of collocated aircraft measurements, *J. Geophys. Res.-Atmos.* **102**, pp. 29917-29927, 1997.
3. X. D. Jing and R. D. Cess, Comparison of atmospheric clear-sky shortwave radiation models to collocated satellite and surface measurements in Canada, *J. Geophys. Res.-Atmos.* **103**, pp. 28817-28824, 1998.
4. R. D. Cess, M. H. Zhang, F. P. J. Valero, et al., Absorption of solar radiation by the cloudy atmosphere: Further interpretations of collocated aircraft measurements, *J. Geophys. Res.-Atmos.* **104**, pp. 2059-2066, 1999.
5. R. C. Yu, M. H. Zhang and R. D. Cess, Analysis of the atmospheric energy budget: A consistency study of available data sets, *J. Geophys. Res.-Atmos.* **104**, pp. 9655-9661, 1999.
6. M. S. Croke, R. D. Cess and S. Hameed, Regional cloud cover change associated with global climate change: Case studies for three regions of the United States, *J. Climate* **12**, pp. 2128-2134, 1999.

7. D. E. Waliser, R. A. Weller and R. D. Cess, Comparisons between buoy-observed, satellite-derived, and modeled surface shortwave flux over the subtropical North Atlantic during the Subduction Experiment, *J. Geophys. Res.-Atmos.* **104**, pp. 31301-31320, 1999.
8. R. D. Cess, T. Qian and M. Sun, Consistency tests applied to the measurement of total, direct and diffuse shortwave radiation at the surface, *J. Geophys. Res.-Atmos.*, **105**, pp. 54,881-24,887, 2000.
9. R. D. Cess, M. Zhang, B. A. Wielicki, D. F. Young, X-L. Zhou and Y. Nikitenko, The influence of the 1998 El Niño upon cloud-radiative forcing over the Pacific warm pool. *J. Climate*, **14**, pp. 2129-2137, 2001.
10. Y. P. Zhou and R. D. Cess, Validation of longwave atmospheric radiation models using Atmospheric Radiation Measurements (ARM) data, *J. Geophys. Res.-Atmos.* **105**, pp. 29,703-29,716, 2000. Y. Zhou and R. D. Cess, Algorithm-development strategies for retrieving the downwelling longwave flux at the Earth's surface, *J. Geophys. Res.-Atmos* (in press), 2000.
11. Y. Zhou and R. D. Cess, Algorithm-development strategies for retrieving the downwelling longwave flux at the Earth's surface, *J. Geophys. Res.-Atmos.*, **106**, pp. 12,477-12,488. 2001.

In the following the results of Paper Nos. 9 through 11 are specifically summarized.

Paper Number 9. Clouds radiatively cool the climate system through reflection of shortwave (solar) radiation from the system, while they radiatively warm the system by reducing the outgoing longwave (infrared) radiation that increases the atmospheric greenhouse effect. One of the many insights provided by data from the Earth Radiation

Budget Experiment (ERBE) involved understanding the role of clouds on the climate system through analyzing top-of-the-atmosphere (TOA) cloud-radiative forcing measurements provided by ERBE (Ramanathan et al. 1989; Harrison et al. 1990). For example, Kiehl and Ramanathan (1990) demonstrated there is a near cancellation between the longwave (LW) and shortwave (SW) components of cloud-radiative forcing (CRF) in tropical deep-convection regions, such that the net CRF is nearly zero; i.e., the SW cooling and LW heating by clouds are of similar magnitude but of opposite sign, so that $(\text{SW CRF})/(\text{LW CRF}) \approx -1$.

To investigate the possibility that the strong 1998 El Niño might have influenced the relative magnitudes of SW versus LW CRF, we have combined two datasets that provide measurements of CRF at the TOA. One set consists of ERBE measurements made over a five-year period (1985-1989) and the other CERES (Clouds and the Earth's Radiant Energy System) measurements which comprise ERBE-like measurements for the first eight months (January-August) of 1998. Thus the ERBE data encompass the 1987 El Niño, while the CERES data include the stronger 1998 El Niño which persisted until July 1998 (Wong et al. 2000). In Paper No. 9 we showed that during the 1998 El Niño there was a significant increase in the magnitude of the $(\text{SW CRF})/(\text{LW CRF})$ ratio throughout the warm pool of the tropical western Pacific (TWP), an increase which exceeds that typical of interannual variability in other years. We next selected a region contained within the warm pool for which sea-surface temperature (SST) and the clear-sky TOA emitted LW and reflected SW radiative fluxes were temporally stable. For the five ERBE years attention is directed to the same eight months (January-August) for which CERES data are available.

The five-year (January 1985 to December 1989) ERBE S-4G data consist of scanner measurements that have been temporally averaged into monthly means and spatially averaged into 2.5° by 2.5° longitude-latitude grids. In addition to TOA reflected SW and TOA emitted LW, the S-4G product also provides SW CRF and LW CRF (Ramanathan et al. 1989; Harrison et al. 1990), for which

$$\text{SW CRF} = R - R_c$$

where R denotes the TOA all-sky reflected SW and R_c that for clear skies, while

$$\text{LW CRF} = F_c - F$$

where F and F_c respectively denote the all-sky and clear-sky TOA emitted LW. Typically SW CRF is negative (cooling) and LW CRF positive (warming). The eight-month (January-August 1998) CERES ES-4G data set provides scanner measurements, made from the TRMM (Tropical Rainfall Measuring Mission) Satellite, that have been averaged into monthly-mean gridded data in the same way as the ERBE data so as to provide continuity of the two datasets. In the following we adopt eight-month means (January-August), providing five (1985-1989) ERBE means and one (1998) CERES mean. The ERBE data are from the Earth Radiation Budget Satellite, which was in a 57° orbit relative to the equator, while for the TRMM Satellite the inclination angle is 35° . Thus both satellites sampled throughout the diurnal cycle.

The NET CRF is simply the sum of the SW and LW components; i.e.

$$\text{NET CRF} = \text{SW CRF} + \text{LW CRF} \quad (1)$$

so that cancellation between SW CRF cooling and LW CRF warming ($\text{NET CRF} = 0$) can be expressed by the ratio

$$N = - (\text{SW CRF})/(\text{LW CRF}) = 1$$

If $N > 1$, then SW cooling dominates.

The SW and LW CRFs, averaged over the selected region of the warm pool, and as a function of measurement period, are shown in Figs. 1a and 1b, respectively. Figure 1c in turn demonstrates the near cancellation of the SW and LW components of CRF ($N \approx 1.08$) for the normal years (1987 and 1998 excluded). As for the tropical pacific in general (Fig. 1), N is considerably enhanced during the 1998 El Niño. If this enhancement is associated with that specific event, we should anticipate that the enhancement be greatest during the first few months of 1998, since this El Niño began to dissipate in the spring of 1998 and had essentially ended by late July. This in fact is the case, as is shown in Fig. 2 by comparing the first four months (JFMA) of 1998 to the following four months (MJJA), and with reference to the 1985 and 1989 normal periods. This further serves to emphasize that the $N = 1.25$ result in Fig. 3a is not an artifact of ERBE-CERES differences.

A change in cloud vertical structure (i.e., cloud heights and optical depths) constitutes a probable explanation for 1998 versus the ERBE period. Much can be learned by understanding the relationship between N and NET CRF, and to this end a broadband

radiative transfer model was used to show the dependence of this relationship on cloud altitude and cloud optical depth. The calculations were performed using the 4-stream Fu-Liou radiative-transfer model (Fu and Liou 1993) for a climatological mean tropical atmosphere. The solar zenith angle was set to 60 degrees, and the diurnal average reflected solar radiation was approximated using the diurnal average insolation of 413 Wm^{-2} in the tropics and with cloud albedo evaluated at the 60-degree solar zenith angle. The intent was to show a reasonable approximation of the results expected, rather than to exactly model any particular case. Results for cloud top altitudes of 15, 9, and 6 km and optical depths ranging from 0.1 to 128 are shown in Fig. 5a, in which SW versus LW CRF cancellation corresponds to the $N = 1$ and $\text{NET CRF} = 0$ intercept.

Consider first the cloud top altitude of 15 km. As cloud optical depth increases from a small value (progressing from bottom to top), both cloud emissivity and cloud albedo simultaneously increase, but with the former dominating so that NET CRF increases (net warming). As cloud optical depth continues to increase, this reverses so that cloud albedo increases more rapidly than does cloud emissivity which rapidly asymptotes to unity, ultimately resulting in $\text{NET CRF} < 0$ (net cooling). Within the confines of the model, the 15 km cloud depicted in Fig. 3a can be thought of as representing thin cirrus on the right (small optical depths) to deep-convective (optically thick) clouds at the leftward termination of the curve, and we will shortly elaborate on this termination. Because they are warmer, middle-level clouds, such as those associated with middle-level convection and decaying mesoscale convective systems, have smaller LW CRF, and hence larger N and smaller NET CRF, than do high clouds. The 9 and 6 km clouds depicted in Fig. 3a demonstrate this point, with the latter producing net cooling for all cloud optical depths.

Values of N for low clouds, such as marine stratocumulus, can considerably exceed the range given in Fig. 3a. An important point is that the model calculations of 3a are for a cloud fraction of unity. Kiehl (1994) has shown that N is invariant to cloud fraction, but NET CRF is not. If we had included cloud fraction, then for a point on any of the three curves, there would additionally be a horizontal progression of points with the same value of N , but with NET CRF going to zero as cloud fraction tends to zero.

N versus NET CRF scatter plots are given in Figs. 3b and 3c for two normal years and in Figs. 3d and 3e for the two El Niño years. Each point represents an ERBE/CERES monthly mean and 2.5° by 2.5° grid average. The two normal years (Figs. 3b and 3c) are largely consistent with our expectations from Fig. 5a for high clouds, taking into account that fractional cloud cover will move the data points toward NET CRF = 0. The data points span the range from thin cirrus on the right to deep-convective clouds on the left. Some data to the upper left exceed $N \approx 1.3$, which is the maximum value of N for deep-convective clouds (Paper No. 9) and may be due to middle-level clouds, but if so their populations are small. Moreover, the radiometric measurements cannot preclude the existence of middle-level clouds in conjunction with high-level clouds. The data for the 1987 El Niño, shown in Fig. 3d, are not appreciably different than for the two normal years (Figs. 3b and 3c), except there appear to be fewer thin cirrus which, by itself, would increase the average value of N . As shown in Fig. 3e, however, the strong El Niño of 1998 is quite different than for the other years. This argues that the changes in Fig. 3e, relative to Figs. 3b through 3d, are mainly due to a reduction in average cloud height, since the maximum value of N has increased to 1.8 for 1998.

Paper Number 10. Theoretically based radiative-transfer models are major tools used to study atmospheric radiative transfer processes associated with clouds. However, there are many uncertainties associated with cloud radiation parameterizations in current radiation models. While several recent studies have focused on the models' shortwave (SW) computation (e.g., Cess et al., 1995; Chou and Zhao, 1996; Waliser, et al., 1996), it is equally important to realize that there are large uncertainties in the longwave (LW) calculations. For example, the Intercomparison of Radiation Codes in Climate Models (ICRCCM) found that for optically thin clouds the LW difference among models can be as large as 30-80 Wm^{-2} (Ellingson et al., 1991), while a 10-20 Wm^{-2} discrepancy was found for the surface downward longwave radiation among several general circulation models (GCMs) and observations (Wild at al., 1995). It is essential that GCMs calculate the longwave flux correctly in order to project climate change caused by increasing greenhouse gases.

The validation of radiation models with in-situ measurements requires concurrent input of the atmospheric state, including cloud information, which are themselves difficult to access and subject to large uncertainty. Cloud microphysical properties are commonly estimated in terms of bulk cloud parameters measured using a variety of different instruments and techniques and the input parameters required by different models are not always provided as needed by the observations. As demonstrated in Paper No. 10, this requires careful examination and collocation of various parameters before the data can be used as input to the radiation models.

Figure 5 shows scatter plots of model calculations versus clear-sky measurements made at the Atmospheric Radiation Measurements (ARM) Southern Great Plains (SGP)

Central Facility. The clear-sky calculations of OLR (outgoing longwave radiation) and SDLW (surface downwelling radiation) were made with the National Center for Atmospheric Research Community Climate Model (CCM2/CCM3) Column Radiation Models (CRM) and the Moderate Resolution Transmittance MODTRAN radiation model. Fu-Liou model calculations from CAGEX (Charlock and Alberta, 1996) are also shown for comparative purposes. A total of 101 clear-sky soundings were available for use in the calculations, but only 57 SDLW measurements were available due to missing SDLW data. All four models produce OLR fluxes that are higher than the measurements, while their SDLW fluxes lower than the surface pygeometer measurements, with the CCM2 CRM having the largest and MODTRAN3 the smallest biases. The smaller biases for the CCM3 CRM, relative to that for CCM2, is consistent with the improvement of CCM3 CRM by inclusion of minor trace gases, so that the non-CO₂ trace gases produce about a 6-7 Wm⁻² difference in both OLR and SDLW, which is close to the 5 Wm⁻² estimate of Slingo and Webb (1992). The Fu-Liou radiation model has taken account of the radiative effects of H₂O, CO₂, O₃, CH₄, and N₂O. CFCs are not included, but CFC radiative forcing is reported to be less than 1 Wm⁻² (Charlock and Alberta, 1996). MODTRAN3 includes most of the absorbing gases, including H₂O, CO₂, O₃, N₂O, CO, CH₄, NO, SO₂, NO₂, NH₃, HNO₃, and CFCs.

The overall overestimation of OLR and underestimation of SDLW for all the models could be caused by a systematic underestimate of the models' gaseous absorption. For MODTRAN3, which includes a new and improved water vapor continuum, the minimal biases in OLR and SDLW (relative biases of 2.5% and -1.3%, respectively) do not appear to be related to the background aerosol used in that model. We have performed

sensitivity studies which show that aerosols impact SDLW much more than OLR, while MODTRAN3 has a larger OLR than SDLW bias. Also, these sensitivity studies showed that inclusion of aerosols within a model should increase SDLW by about 1 to 3 Wm^{-2} and reduce OLR by about 0.4 to 1.0 Wm^{-2} . These effects will not substantially impact the results shown in Figure 5.

Figure 6 compares the models to observations for 46 cases for which satellite observations determined the presence of a single-layered cloud. The bias and RMS errors are for all 46 overcast cloud cases, and the calculations using the Fu-Liou code are the same CAGEX calculations as shown in Figures 3a and 3d. The MODTRAN3 calculations give the best overall estimates of both OLR and SDLW, although the Fu-Liou model and the CCM3 CRM are both in quite reasonable agreement with the measurements. An interesting point is that the 7 Wm^{-2} difference between the CCM3 and Fu-Liou OLR biases (Figure 5c versus 5a) coincides with the fact that cloud scattering causes a 6 Wm^{-2} reduction in OLR within the Fu-Liou model (Fu et al., 1997).

In summary, data taken at the Atmospheric Radiation Measurement Program's Central Facility in Oklahoma have been used to validate the top-of-the-atmosphere and the surface longwave radiative fluxes for two widely used radiation models: the Column Radiation Model from the National Center for Atmospheric Research Community Climate Model (CCM), and the Moderate Resolution Transmittance (MODTRAN3) radiation code. For completeness, we have also included in these comparisons the GAGEX-computed fluxes using the Fu-Liou radiation model (Fu and Liou, 1992).

The clear sky results show some systematic underestimates of SDLW and overestimates of OLR for all models. But one should not conclude that the models actually

underestimate the greenhouse (SDLW minus OLR) effect, since we cannot exclude the possibility of insufficient input of absorbing gases. For example, the upper atmospheric water vapor is highly suspect.

For overcast-skies, an important issue to resolve when using a point-to-point radiative transfer model to represent a grid-averaged atmospheric state and radiative fluxes is to correctly collocate the various measurements taken from the satellite instruments, the surface instruments, and the sounding profiles. The TOA-surface collocation problem, in terms of cloud measurements, was illustrated in Paper No. 10 using cloud amount and base height as measured by both by the satellite and from the surface. Results from this study show that the cloud parameters taken from satellite spatial averages and surface temporal averages have large discrepancies when considered case by case. Moreover, a crucial mistake in using the surface microwave radiometer to measure cloud liquid water content is that this biases the cirrus cloud measurements, since the microwave radiometer cannot measure ice water content. Conversely, the satellite retrievals of IR optical depth (which includes both liquid and ice water content) are very useful, not only in their large area coverage, but also for more accurate high cirrus cloud information. On the other hand, the satellite technique has its own limitations, such as poor treatment of multi-layer clouds.

Paper Number 11. The surface radiation budget not only represents direct radiative forcing at the surface, but it also determines the latent and sensible heating fluxes to the atmosphere. The surface net longwave (LW) flux is further a manifestation of the greenhouse effect of the atmosphere, and it is essential to reproduce it accurately. Direct surface radiation measurements, however, are geographically sparse. Over the years many algorithms have been developed in an

effort to obtain a global surface radiation data set. The upward emission from the surface (SULW) is usually determined from the Stephen-Boltzman law given knowledge of the surface temperature and, if available, the surface emissivity. The surface downwelling longwave flux (SDLW) is the most difficult component, since it is significantly affected by the vertical distribution of clouds.

Bulk formulas have been used extensively in the past for estimating SDLW for large-scale studies. These formulas rely on surface or near-surface meteorological variables such as surface temperature and humidity, and in some cases an estimated cloud amount (Berliand, 1952; Idso and Jackson, 1969; Brusaert, 1975; Idso, 1981; Swinbank, 1963; Satterlund, 1979; Prata, 1996). Fung et al. (1984) conducted a detailed examination of several widely used bulk formulas using computations from a detailed radiative transfer model and found large discrepancies. They pointed out that cloud base height, optical thickness, and cloud type have strong impacts on the surface net longwave flux and concluded that it was not possible to use a single cloudiness factor in a bulk formula to evaluate the instantaneous SDLW for overcast skies.

Several techniques have been developed for deriving the SDLW from satellite measurements. Smith and Woolf (1983) demonstrated a regression technique where upward, downward, and net LW fluxes at various levels in the atmosphere were computed from outgoing radiances measured by several VAS (VISSR Atmospheric Sounder) channels. Schmetz et al. (1986) used the bulk formulas of Idso and Jackson (1969) and Idso (1981) to determine the SDLW using surface temperature fields from grid point analyses and cloud parameters derived

from METEOSAT data. Frouin et al. (1988) presented a radiative transfer methodology for estimating the downward longwave radiation over oceans using TOVS (TIROS Operational Vertical Sounder) meteorology together with cloud parameters derived from GOES data. Gupta et al. (1989, 1992) have developed an algorithm to derive the surface longwave fluxes from NOAA's operational sun-synchronous satellite data. This algorithm is used to produce the gridded monthly surface longwave fluxes as part of the Surface Radiation Budget (SRB) project by using the ISCCP-C1 data as its primary input data.

There have also been some attempts to obtain the surface longwave fluxes by exploring the relationships between the top-of-the-atmosphere (TOA) and surface fluxes. For clear skies, Inamdar and Ramanathan (1997) developed an algorithm that expresses a normalized surface longwave flux and TOA flux in terms of column water-vapor amount and TOA radiative measurements in the LW window (8 to 11 μ m). Stephens et al. (1994)) have utilized the outgoing LW radiation (OLR) to derive the monthly-mean SDLW based on a linear regression of SDLW/OLR with column water vapor. Their results are restricted to clear skies and ice-free ocean areas. Harshvardhan et al. (1990) used simulations from a general circulation model (GCM) to estimate a climatological ratio of cloud longwave forcing (CLF) at the TOA and at the surface along with the CLF obtained from Earth Radiation Budget Experiment (ERBE) data to obtain the surface CLF. The surface downwelling and net longwave fluxes for cloudy skies were obtained by adding the surface CLF to the clear sky fluxes, which were calculated from

soundings using a radiation model. While this technique avoids the input of cloud information, it is subject to uncertainty because of the GCM-based CLF ratios

As there is no simple relation between the surface longwave flux and the TOA flux (Ramanathan, 1986), especially for cloudy skies, satellite retrievals normally rely on retrieved atmospheric temperature, water vapor, and ozone profiles, along with cloud properties, by applying them to radiation models and/or algorithms. The disadvantage of such techniques is that they are highly dependent on the satellite retrieved atmospheric profiles and cloud parameters that are subject to large uncertainties.

Without accurate measurements of cloud amount and base temperature, estimates of the cloudy sky SDLW remains a challenge. In Paper No. 11 we presented strategies for developing an algorithm to retrieve the SDLW based on physical models as well as observational data for all-sky conditions. A suggested new algorithm requires input of surface temperature, total precipitable water vapor (PWV) and cloud liquid water path (LWP). The advantage of this technique is that the only cloud parameter required is cloud liquid water which can be measured by both surface and satellite microwave instruments.

The observational data used in this study are from the Atmospheric Radiation Measurement (ARM) Program at the U.S. Southern Great Plains (SGP) and the Tropical Western Pacific (TWP) sites. One of ARM's primary objectives is to develop and test parameterizations of atmospheric water vapor, clouds, and the surface characteristics affecting atmospheric radiation. The Central Facility of each ARM site is designed to serve this purpose by providing high spectral resolution radiometric measurements and a detailed

characterization of the atmospheric column above the facilities. The measurements include surface radiometric fluxes, clouds, aerosols, balloon-borne radiosonde profiles, regular surface meteorological parameters, and satellite observations. The data used in this study are taken from 15 IOPs (Intensive Operation Periods) from 1994 to 1998 (Table 1) at the SGP's Central Facility. The reason to use these IOP data is that GOES TOA fluxes and cloud products (Minnis et al., 1995) are only available for these periods at the time this work was conducted. The 15 IOPs span several years and cover different seasons (see Table 1).

We start by using data for six IOPs at the SGP site: April, July, and October from 1994 to 1996 (see Table 1). Figure 6 summarizes regression plots for the clear-sky data, with the clear-sky identification provided by GOES, in the same manner as reported in Paper No. 11 using model calculations, and the data provide the same conclusions as were obtained from the calculations; the multi-variable regression using surface upward longwave (SULW) and precipitable water vapor, as shown in Figure 6c, produces an excellent correlation of the data with $R = 0.978$. Likewise, Figure 7 summarizes the cloudy-sky data in a similar manner as again reported in Paper No. 11 using model calculations. Note that inclusion of cloud-base height (BH) as a cloud input parameter (Figure 7b versus 7a) provides no enhancement in the correlation. In Paper No. 11 we discussed difficulties in defining an average BH with respect to the model sensitivity studies, and such difficulties are far more pronounced when dealing with observations. The radiation fluxes, precipitable water vapor, and cloud liquid water path (LWP) are all half-hourly averages. An averaged cloud-base height over a half-hour period, however, is not

Table 1. ARM SGP Intensive Observational Periods

IOP	Period	Days
April 94	04/06-04/27	22
July 94	07/11-07/31	21
October 94	10/23-11/15	24
July 95	07/15-08/07	24
October 95	09/25-10/02, 10/08-11/01	33
April 96	04/09-05/09	31
March 97	03/01-03/31	31
April 97	04/01 -04/30	30
June 97	06/18-07/18	30
September 97	09/15-10/05	20
December 97	12/01-12/31	31
January 98	01/01-01/31	31
February 98	02/01-02/28	28
April 98	04/27-5/17	21

something straightforward when dealing with broken clouds (Mark Miller, personal communication). The cloud-base heights provided by Miller are ceilometer measurements averaged over a half-hour period, and averaged together with cloud fraction (to account for

cloud occurrence) and standard deviation of the cloud base (to account for horizontal inhomogeneity). Without using cloud fraction, the cloud-base height is certainly not representative of a half-hourly cloud status. Thus, considering the ambiguity associated with defining a temporal and spatial mean cloud-base height, it is perhaps not surprising that the inclusion of BH provides no improvement in the correlation (Figure 7b versus 7a). In contrast, the inclusion of LWP provides a modest improvement (Figure 7c versus 7a) which, in some circumstances, can prove to be significant as we later demonstrate. Presumably, as with the model sensitivity studies, this improvement is because LWP serves as a surrogate for BH, at least in an average context. Because the emissivity of clouds that impact SDLW is essentially unity, LWP is itself not a cloud input parameter, underscoring its role as surrogate for BH.

The model sensitivity tests and the observational data have both shown that the clear-sky SDLW can be derived from surface temperature (i.e., SULW) and PWV. For the cloudy sky, the inclusion of LWP further increases the accuracy of the correlation. The algorithm determined from the all-sky data for the six IOPs, combining both clear (Figure 6) and cloudy (Figure 7) data, is

$$\text{SDLW} = a + b\text{SULW} + c\text{LN(PWV)} + d(\text{LN(PWV)})^2 + e\text{LN}(1 + f\text{LWP}) \quad (1)$$

where a, b, c, d, e, f are regression coefficients with values of 123.86, 0.444, 56.16, -3.65, 5.30 and 1226.0, respectively, and the units of PWV and LWP are cm. Since $\text{LWP} = 0$ for clear skies, the functional form of this algorithm is such that it is applicable to clear skies as well.

There are nine additional IOPs from 1996 to 1998 for the SGP site which contain three winter months (Table 1), in contrast to the six original IOPs that contained no winter months for which Equation (1) was derived. Figure 8 shows scatter plots of the derived, from Equation (1), versus observed SDLW fluxes for these nine IOPs. The bias errors are minimal, while the RMS errors for all-sky sky and clear-sky are similar.

The positive outliers in Figure 9a correspond mainly to very large values of PWV. According to the ARM Instrument Mentor [*J. Liljegren*, personal communication], precipitable water larger than 5 cm is very rare in Oklahoma, and if the cloud liquid water is greater than 0.1 cm, then it is most likely raining or wet with other forms of precipitation. The large PWV data likewise correspond to LWP normally greater than 0.2, and these excursions are also largely absent for the original six IOPs. Moisture detectors were installed on the MWRs in 1993 so as to flag data as "wet" if there was moisture on the MWR window. And when moisture was detected, a heater was turned on to dry the window. But sometimes the detector dries before the window, and that might explain the positive outliers in Figure 8a.

Data for the negative outliers in Figure 8a correspond to weather that was extremely cold and dry. The average precipitable water and cloud liquid water for negative biases exceeding 50 Wm^{-2} in Figure 8a was 0.64 cm and 0.0055 cm (0.003 cm of cloud liquid water is considered clear), respectively. However, the average surface relative humidity was always greater than 92%. From the weather log we found well-developed overcast stratus fractus over the central facility on Jan 9, 1998, but the MWR instrument recorded essentially zero LWP on that day, and a data inconsistency of this type would result in the algorithm producing a negative bias. This would not, however, explain the negative biases

for clear skies shown in Figure 8b, although their population is considerably less than in Figure 8a.

The data anomalies discussed above are primarily restricted to the three winter months (DEC 97, JAN 98 and FEB 98), and correspondingly produce excessive RMS errors as demonstrated in Figure 9. For the other months the algorithm's bias and RMS errors are quite small. When the three winter month are deleted, the average RMS errors for the remaining six IOPs become 15.9 Wm^{-2} for all skies and 15.0 Wm^{-2} for clear skies.

The tropical Pacific warm pool is characterized by high surface temperatures, an extremely humid lower atmosphere, and extended cirrus clouds caused by deep convection. The SDLW does not have large a variations because of minimal variations of surface temperature and the abundant moisture in the lower atmosphere. The high cirrus clouds, predominately composed of ice and thus not contributing to the measured LWP, hardly affect the surface downwelling longwave flux. The standard deviation of SULW and SDLW for the 6 months of Manus data are only 10 and 15 Wm^{-2} , respectively, and the average PWV is 5.1 cm, much greater than for the Oklahoma site.

Figure 10 shows scatter plots of the derived SDLW versus the observed SDLW, with the derived SDLW evaluated from the algorithm based on the data for the six SGP IOPs. Since TOA clear-sky identification is not available for this data set, the LWP was used as the clear-sky criterion; a sample was regarded as clear when LWP was less than 0.003 cm. The algorithm derived from the data for the six SGP IOPs is thus very successful in replicating the observed SDLW, producing minimal bias and RMS errors. Because of the considerable differences in atmospheric state between Oklahoma and the tropical western Pacific, this suggests that the Oklahoma-derived algorithm may have

broad geographic applicability. The results shown in Figure 10 further underscore the point that a single algorithm appears to be applicable for both clear and cloudy conditions.

The TWP data also serve to demonstrate the importance of including the last term in (2), despite the fact that comparing Figure 7c versus 7a for the SGP data indicates only a modest improvement in the correlation due to this term. The "NO LWP TERM" bias and RMS errors shown in Figure 10 correspond to using the algorithm, as given by (2), with the last term deleted and reevaluating the regression coefficients a, b, c and d again using the six SGP IOPs. Figure 10a shows that including LWP provides a modest improvement in both the bias and RMS errors, as is consistent with the modest improvement in correlation for Figure 7c relative to 7a. But for clear skies, as demonstrated in Figure 10b, the improvement is substantial. This is because the "NO LWP TERM" algorithm applies to an average LWP (in this case for the six SGP IOPs), whereas for clear skies (2) is being applied to the limit $LWP = 0$. Thus the inclusion of the last term in (2) is important if an algorithm derived from all-sky data is to be applied to clear-sky conditions.

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Fig. 1. (a) Eight-month means of the SW CRF for the selected region of the warm pool and for each of the six years. (b) The same as (a) but for the LW CRF. (c) The same as (a) but for the ratio N.

Fig. 2. Four-month means of the ratio $N = -(SW\ CRF)/(LW\ CRF)$ for 1985, 1989 and 1998 for the selected region of the warm pool.

Fig. 3. (a) Model calculations of N versus NET CRF for cloud-top altitudes of 15, 9 and 6 kms. The shaded areas represent domains that are not physically possible (N denoting warming but NET CRF denoting cooling and vice versa). (b) Scatter plot of N versus NET CRF for 1985. (c) The same as (b) but for 1989. (d) The same as (b) but for 1987. (e) The same as (b) but for 1998.

Fig. 4. (a) Clear-sky OLR as determined from the Fu-Liou model compared to the measured clear-sky OLR. (b) The same as (a) but for the CCM2 CRM. (c) The same as (a) but for the CCM3 CRM. (d) The same as (a) but for MODTRAN. (e) through (h) The same as (a) through (d) but for SDLW.

Fig. 5. Comparison of the four models to measurements for overcast skies. See text for an explanation of cloud input parameters.

Fig. 6. (a) Scatter plot of SDLW versus OLR for the first six SGP's IOPS. **(b)** The same as (a) but versus SULW. **(c)** The same as (a) but versus a function whose coefficients were determined from regression.

Fig. 7. (a) The same as Figure 3c but for cloudy skies. (b) The same as (a) but with cloud-base height (BH) included in the x-axis variable. (c) The same as (a) but with cloud liquid water path (LWP) included in the x-axis variable.

Fig. 8. (a) Scatter plot of the algorithm-determined SDLW versus the measured SDLW for the SGPs all-sky measurements and the latter nine IOPs. (b) The same as (a) but for clear skies.

Fig. 9. Scatter plots of the algorithm-derived SDLW versus that measured at the SGPs for each of the latter eight IOPs.

Fig. 10. The same as Figure 8 but for the first six months of measurements made at manus Island in the Tropical Western Pacific (TWP).