Observations of stratocumulus clouds and their effect on the eastern Pacific surface heat budget along 20°S

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Abstract
Widespread stratocumulus clouds were observed on 9 transects from 7 research cruises to the southeastern tropical Pacific Ocean along 20°S, 75°-85°W in October-November 2001-2008. The nine transects sample a unique combination of synoptic and interannual variability affecting the clouds; their ensemble diagnoses longitude-vertical sections of the atmosphere, diurnal cycles of cloud properties and drizzle statistics, and the effect of stratocumulus clouds on surface radiation. Cloud fraction was 0.88 and most 10-minute overhead cloud fraction observations were overcast. Clouds cleared in the afternoon (15 h local) to a minimum of fraction of 0.7. Precipitation radar found strong drizzle with reflectivity above 40 dBZ.
Cloud base heights rise with longitude from 1.0 km at 75° to 1.2 km at 85°W in the mean, but the slope varies from cruise to cruise. Cloud base-lifting condensation level (CB-LCL) displacement, a measure of decoupling, increases westward. At night CB-LCL is 0-200 m, and increases 400 m from dawn to 16 h local time, before collapsing in the evening.

Despite zonal gradients in boundary layer and cloud vertical structure, surface radiation and cloud forcing are relatively uniform in longitude. Clouds reduce solar radiation by -160 W m\(^{-2}\) and radiate 70 W m\(^{-2}\) more downward longwave radiation than clear skies.

Coupled model intercomparison project (CMIP3) simulations of the climate of the 20th century show 40±20 W m\(^{-2}\) too little net cloud radiative cooling at the surface. Simulated clouds have correct radiative forcing when present, but models have ~50% too few clouds.

1. **Introduction**

Accurate simulation of tropical southeastern Pacific Ocean sea surface temperature (SST) is challenging for coupled general circulation models (GCMs; Mechoso et al. 1995, Davey et al. 2002, de Szoeke and Xie 2008). Warm errors of 2°C in SST are found at 20°S, 75°W in most of the Coupled Model Intercomparison Project phase 3 (CMIP3) models assessed by de Szoeke et al. (2010). Atmospheric subsidence over cool SST and high surface pressure provides a stable cap to the marine boundary layer, reducing the cloud-top entrainment rate and increasing stratiform clouds (Klein and Hartmann 1993).
The high-albedo clouds shade the ocean surface from strong tropical solar radiation, thus cooling it. This positive feedback between shallow clouds and SST helps to maintain more low stratus and stratocumulus clouds and cooler SST in the southern hemisphere than the northern hemisphere eastern tropical Pacific. The north-south symmetry is broken by trade-wind driven upwelling at the northwest-southeast slanted American coast (reviewed by Xie 2004). The clouds and their feedbacks are difficult to simulate accurately because critical turbulence and precipitation processes are parameterized in the atmospheric models. Testing models and improving parameterizations thus call for detailed observations of stratus cloud processes.

Figure 1 shows the location of cool tropical SST (shaded) and the stratocumulus cloud deck (cloud fraction, contoured) for climatological average October-November AMSR SST (Risien and Chelton 2008) and MODIS cloud fraction (Platnick et al. 2003). De Szoeké et al. (2010) shows Coupled Model Intercomparison Project (CMIP3) 20th century GCM simulations with 20-30 W m\(^{-2}\) excess net surface radiative warming compared to satellite and in situ observations. The downwelling excess solar and reduced longwave radiation is consistent with too little simulated cloud or simulated clouds with weaker radiative properties than observed. Here we explore sensitivity of observed surface downwelling radiation to cloud fraction, atmospheric temperature and water vapor profiles, cloud boundaries, and liquid water path.

Cronin et al. (2006) and Fairall et al. (2008) measure the effect of marine clouds on the tropical Eastern Pacific heat budget using observations from TAO buoys along 95° and
110°W, the Woods Hole Oceanographic Institution (WHOI) Stratus buoy at 20°S, 85°W, and ship observations from the buoy-tending cruises. We perform a similar radiative analysis for research cruises to the stratocumulus region along 20°S between 85° and 75°W. Kollias et al. (2004) documents cloud and marine boundary layer vertical structure using rawinsondes, cloud remote sensing, and aerosol sampling on a research cruise in the stratocumulus region in 2003. We combine observations in the 20°S eastern Pacific stratocumulus region from research cruises in 2001 and 2003-2008. All but one of the research cruises were in October-November.

Section 2 introduces the ship-based observations of clouds, drizzle, related surface meteorology, radiative fluxes, and atmospheric rawinsondes. Section 3 presents the mean vertical-longitude section of cloud properties. Section 4 presents the diurnal cycle of clouds and section 5 assesses the effect of clouds on the surface radiation budget. Section 6 provides an example of how these cloud observations can be used to assess clouds in 15 GCMs used for climate projection. Section 7 summarizes the conclusions.

2. Ship-based Observations

data set, and used surface flux observations from the synthesis data set to evaluate CMIP3 models. Here we summarize cloud observations from the synthesis data set.

Ships yearly serviced the Woods Hole Oceanographic Institution (WHOI) Stratus Ocean Reference Station at 20°S, 85°W (Colbo and Weller 2007) and the Chilean tsunami buoy at 20°S, 75°W. NOAA PSD scientists made observations on research cruises to the southeastern tropical Pacific Ocean along 20°S, 75°-85°W in 7 years (2001, 2003, 2004, 2006-2008). A total of eight 20°S sections (3 in 2008) by the instrumented ships are included in the NOAA synthesis data set. Tracks of the ships for the nine sections are plotted in Fig. 1. Dates of reaching destinations at 75° and 85°W are listed for each track in the key below.

For the VOCALS Regional Experiment (REx) in 2008 the NOAA ship Ronald H. Brown made 3 longitudinal along 20°S sections in 2 legs. In the first leg it made an eastward section between servicing the WHOI Stratus Station and the Chilean tsunami buoy. On the second leg the Brown made an east-west-east round trip from Arica, Chile to survey ocean eddies. All 20°S transects, except for the ocean survey in the second leg of VOCALS REx, were completed in under 3 days.

Atmospheric soundings

In all, 487 rawinsondes profiled temperature, humidity, and vector winds of the troposphere in the vicinity of 20°S. Every 4 or 6 hours a rawinsonde was released from the fantail of the ship during the cruises. The frequency of rawinsondes allows sampling
of atmospheric diurnal cycles, day-to-day variability, and gradients in longitude along 20°S. In 2005 soundings are available only west of 82°W. Maximum spacing between the rawinsonde launches is 130 km, based on a ship’s maximum speed of 12 nautical miles per hour (6.2 m s\(^{-1}\)), but usually the distance between rawinsondes is much less.

Surface meteorology and fluxes

Surface air temperature and relative humidity are measured from a mast 17.5 m above sea level at the bow of the ship. SST of the upper 5 cm of seawater is measured with a “sea snake” floating thermistor. The surface lifting condensation level (LCL) and its temperature is computed by adjusting the observed humidity and temperature to 500 m according to similarity theory, then adiabatically lifting air with those properties until it is saturated.

Downwelling solar and longwave radiative fluxes were measured by a pyranometer and a pyrgeometer mounted on an upper deck of the ship. Upwelling radiative fluxes are modeled from the albedo of the sea surface (0.05) and Planck blackbody radiation from the SST, assuming the sea surface emissivity is 0.97. Clear-sky solar fluxes were computed from the model of Iqbal (1988) using the solar zenith angle and integrated water vapor in the atmosphere. Clear-sky longwave radiation is computed from the Hare (2005) 2-parameter model based on latitude and surface specific humidity. A 3-parameter clear-sky model also using integrated water vapor has \(-2 \text{ W m}^{-2}\) weaker downwelling longwave radiation. Since the 3-parameter model is within sampling variability of the 2-parameter model, we use the simpler 2-parameter model.
Cloud remote sensing

Passive and active remote sensing instruments measured properties of clouds over the ship. Vertically pointing narrow-band microwave radiometers measured sky brightness temperature $T_B$ at 24 and 31 GHz (~1.5 and 1 cm), from which liquid water path (LWP) and water vapor path (WVP) are calculated (Zuidema et al. 2005). A Vaisala pulsed lidar ceilometer measured optical backscatter in the atmosphere, and retrieved the occurrence of clouds and cloud base height of up to three clouds along its vertically pointing beam. Cloud fraction is computed from 10-minute averages of cloud occurrence from the ceilometer.

Cloud top height was computed by a variety of in situ and remote sensing methods. A strong inversion was always present in the vicinity of 20°S, and was coincident with cloud top when clouds were present. The rawinsondes detect inversion base height from the coincident sharp rise in temperature and drop in humidity. The inversion base is identified as the minimum temperature below the strongest gradient of temperature in the lower 3 km of the troposphere. In 2001 and 2004-07 a NOAA 915-MHz wind profiler retrieved boundary layer inversion height more frequently from Bragg scattering at the gradient in atmospheric index of refraction at the inversion. In 2001, 2003, and 2008 cloud radar detected cloud top height as the highest range gate containing cloud reflectivity above the radar noise threshold. Different measures of cloud top height were found to agree when they coincided, though not all were available at any one time. Cloud
top height from cloud radar is found to be coincident within 10 m of inversion base height from radiosondes for stratocumulus clouds over the southeastern tropical Pacific.

The 10-minute 15th, 50th, and 85th percentiles of cloud base height sampled every 20 s by pulsed lidar ceilometer are recorded in the synthesis data set. These percentiles are less influenced by outliers than the mean. The ceilometer sometimes returns cloud base height from clouds above or below the stratocumulus cloud layer. Solar noise is a problem for ceilometer cloud detection for high ranges when the sun was near zenith. For stratocumulus cloud base height, we use the 85th percentile cloud base height to minimize the contribution of boundary layer shallow cumulus below the stratocumulus, and exclude cloud bases detected above the stratocumulus cloud top.

Aerosol concentrations

Aerosol number concentration (diameter $D>0.1 \times 10^{-6}$ m) was measured by Texas A&M University (TAMU, Tomlinson et al. 2007) in 2003 and 2004. In 2005, 2006, and 2007 the number is computed as the sum of aerosols counted by a Particle Measurement Systems Lasair-II. Differences between sensitivity of the two instrument platforms may affect the absolute accuracy, yet each detects relative changes in the size-resolved drop concentrations. In 2008 the Pacific Marine Environmental Laboratory (PMEL) measured aerosols aboard the ship (Covert, pers. communication). Three particle diameter bins were chosen to be uniform across both platforms: 0.1-0.3, 0.3-1.0, and $>1.0$ µm.

Accumulation mode aerosols ($D>0.1$ µm) have critical supersaturation less than 0.17%, while coarse mode aerosols ($D>1$ µm) have critical supersaturation less than 0.01%.
Aitken mode aerosol \((D<0.1\times10^{-6}\text{ m})\) concentrations were measured by TAMU and PMEL.

Radar observations

Doppler precipitation radar (C-band, 5 cm wavelength) made range-height and azimithal scans at different elevation angles every 3-10 minutes in 2001, 2004, and 2006-08. The C-band radar provides reflectivity and radial velocity within a 60 km radius of the ship.

In 2001, 2003, and 2004 the vertically pointing NOAA millimeter-wavelength cloud radar (MMCR, 8.6 mm) measured clouds and precipitation from the ship (Kollias 2004, Comstock et al. 2004, Comstock et al. 2007). In VOCALS 2008 NOAA deployed a new motion-stabilized W-band (3 mm) Doppler cloud radar sensitive enough to detect clouds and capable of measuring precipitation drop and cloudy air vertical velocities (Moran et al. 2011). In VOCALS 2008 the NOAA W-band measured high-resolution cloud and precipitation reflectivity and vertical velocities, while the C-band simultaneously sampled the larger surrounding area and mesoscale organization of precipitating structures (Yuter et al. pers. communication).

3. The Cloud-capped Boundary Layer Section Along 20°S

a. Thermodynamics and wind

Measurements from rawinsondes released from research cruises along 20°S are presented as longitude-height sections for 2001, 2003, 2004, 2006-2008 in Fig. 2. Multiple soundings from when the ship was on station have been eliminated for clarity of presentation, leaving 157 soundings shown in Fig. 2. Potential temperature and specific
humidity gradients show a well-mixed boundary layer 1.0-1.8 km deep. The mean boundary layer potential temperature over all 8 sections is 290±1 K. The boundary layer is capped by a strong (10 K) inversion, over which the free troposphere has a 6.2 K km⁻¹ stable potential temperature gradient. Slightly stable stratification within the upper boundary layer is occasionally observed, e.g. in 2008 leg 1 (Fig. 2k) east of 78°W and leg 2.2 (Fig. 2o) west of 81°W. The moist-adiabatic lapse rate of boundary layer stratocumulus clouds causes this conditionally stable potential temperature stratification. Conditionally stable potential temperature gradients are nearly ubiquitous in the top 100-400 m of the boundary layer, where clouds are almost always observed. Stable layers are also sometimes observed below the cloud layer, presumably evidence of decoupling of warmer potential temperature air in the upper boundary layer from the cooler surface mixed layer.

Fig. 2 shows boundary layer potential temperature is quite uniform in longitude, increasing approximately 2 K from 75° to 85°W. This increase corresponds to the 2°C SST gradient (Fig. 1). Temperature varies among the 20°S transects. Warmer temperature and higher specific humidity are observed in 2003 and 2004. These transects occurred later in the year, in late November and early December, when SST is seasonally higher. Free tropospheric temperature is also higher and more stably stratified in December 2004, which may contribute to capping the boundary layer to only about 1 km height. Such cases must be interpreted with caution however, since synoptic variability influences the soundings, but is not resolved by the cruise data. The three transects in 2008 show boundary layer potential temperature progressively increasing about 1 K over the month.
from October 27 to November 28. Assuming the seasonal cycle repeats each year we attribute some differences in the sampled atmospheric profiles to their time in the seasonal cycle. Though SST and the atmospheric structure also vary interannually, the 8 transects sample at best 2 El Niño/Southern Oscillation (ENSO) periods, so we cannot estimate interannual variability with statistical certainty with this data set.

Water vapor specific humidity distinctly marks the inversion between the moist boundary layer and the dry free troposphere (Fig. 2 right column). Specific humidity is 7-10 g kg\(^{-1}\) and decreases with height in the boundary layer. This specific humidity gradient is found in individual soundings. The specific humidity gradient below the cloud, where potential temperature is constant, must be achieved by dry adiabatic processes. The constant gradient of specific humidity suggests a layer mixing between two end members: the moist surface layer and the drier cloud layer. Specific humidity in the free troposphere is usually below 1 g kg\(^{-1}\) and always distinctly drier than the MABL. Combined with the increase of temperature at the inversion, the drier air results in a dramatic drop of relative humidity, from saturation in the MABL cloud to less than 5% relative humidity in the lower free troposphere (Fig. 3a)

The height of the boundary layer increases westward in 2001, 2006, 2007, and 2008 leg 2, but little or no westward tilt is evident in 2003, 2004, or 2008 leg 1. During VOCALS REx, aircraft traversed 20°S from Arica, Chile to 80°W in and above the boundary layer 15 times from 2008 October 15 to November 13 (Bretherton et al. 2010). The slope of the boundary layer depth also varied among these flights. Seeing this variability on
interannual to daily timescales, we conclude that considerable synoptic variability affects the boundary layer clouds, which the yearly research cruises sample but do not resolve.

The shaded frequency-altitude diagrams in Fig. 3 show the vertical structure of the temperature and humidity distribution of all 487 soundings within 2° of 20°S, 75-85°W. The median (solid) and mean (dashed) profiles differ from the mode (dots) of the distribution. This is especially true in the vicinity of the inversion, where the rise and fall of inversion height results in sampling properties distributed bimodally between boundary layer and free-tropospheric air, yet rarely a mixture of the two air masses. The altitude-resolved distribution of thermodynamic variables indicates the sharpness of the inversion. While the mean smoothes the inversion over 500 m, the modes of the temperature and humidity distributions jump from boundary layer to free tropospheric properties over only tens of meters, with few intermediate values.

Figure 3a shows relative humidity (RH) is about 70% at the surface, increasing linearly with height in the boundary layer. The mode of the soundings is saturated (RH=100%) in the top 500 m of the boundary layer, indicating clouds are usually present. The median and mean RH are less than 100%, skewed toward unusual soundings that are unsaturated at a given height. Thermodynamic profiles in Fig. 3 are bimodally distributed between the clear free-troposphere air and boundary layer clouds near the inversion. Arithmetic means over nonlinear transitions, such as between saturated and unsaturated air, are a poor representation of clouds. Statistics of clouds will be explored further in subsection 3b.
The mean profile of wind along 20°S has vertically uniform 7 m s\(^{-1}\) southeasterlies (\(u\) and \(v\) components each 5 m s\(^{-1}\)) throughout the boundary layer, except for a 100-m thick layer at the surface with what appears to be a logarithmic velocity profile (Fig. 4). The mean profile has uniform northwesterly shear above the inversion, reaching westerlies of \(u = 22\) m s\(^{-1}\) and northerlies of \(v = -11\) m s\(^{-1}\) at 13 km altitude in the subtropical jet. Winds change gradually across the inversion compared to the thermodynamic variables. The standard deviation of zonal wind is 3 m s\(^{-1}\) in the boundary layer and 4 m s\(^{-1}\) above the inversion. The standard deviation of meridional wind is 2 m s\(^{-1}\) in the boundary layer and 3 m s\(^{-1}\) above the inversion.

Thermodynamic atmospheric soundings are remarkably constant over the 7 years of research cruises to 20°S. We idealized a sounding to 14 significant levels based on the 487 soundings along 20°S (Table 1).

**b. Clouds**

The size and frequency of clouds critically affect the surface heat budget through their effect on surface radiation. Figure 5a shows the longitude-height section of mean MABL top, cloud base, and lifting condensation level (LCL) in 2.5° longitude bins from 75 to 85°W along 20°S. The filled gray boxes show the mean top and bottom of the cloud, while the unfilled boxes show standard deviation of the mean. The standard deviation of the mean is computed over all 10-minute samples in the selected location, but the variability is dominated by transect-to-transect variations. Each of the 9 transects is an
independent sample of synoptic, seasonal, and interannual variability. Cloud thickness averages 230 m across the transect. The MABL top height and cloud base height increase westward on average, with little change in mean cloud thickness. The range of cloud heights includes cases in which the cloud height did not increase with longitude (cf. Fig. 2). Cloud heights were at the low end of the distribution in 2004 December—late in the seasonal cycle—but did not differ much compared to cloud heights observed other years in October and November. Excluding 2004 raises mean cloud heights by about 20 m, a difference within the range of variability among October-November transects.

The LCL is the height at which we expect an undilute parcel from 500 m (transformed with similarity theory from measurements at 15.5 m) to reach saturation with respect to water vapor if it cools adiabatically as it is lifted. This is the lowest level that a cloud is likely to form. While cloud base heights increase westward, LCL remains approximately level, decreasing westward by only a statistically insignificant amount (Fig. 5a). The least-squares regression of distance between cloud base and the LCL rises 30 m per 100 km of longitude. Figure 2 shows cloud base from the ceilometer (blue dots) is sometimes within 100 m of the LCL (red dots), e.g. 2007 between 80-77°W. More often, cloud base is several hundred meters above the LCL. The model of Bretherton and Wyant (1997) predicts entrainment of warm dry air from above the boundary layer dilutes saturated cloud air and evaporates clouds from their base, creating a warm decoupled sub cloud layer. Displacement of the cloud base from the LCL is consistent with the decreasing humidity gradient between the surface mixed layer and cloud base.
Cloud fraction (gray circles Fig. 5b) shows a weak decreasing trend westward from 92% coverage at 75°W to 84% at 85°W. The middle 3 longitudes of cloud fraction and LWP are averaged together at 80°W. Liquid water path (LWP) increases westward by 40% over 10° longitude, despite decreasing cloud fraction (Fig. 5b). LWP in Fig. 5 is averaged regardless of whether a cloud is present. Average LWP conditionally sampled in the cloud would be slightly greater. Average liquid water content (LWC), increases 1 g m\(^{-3}\) km\(^{-1}\) per km of cloud thickness. For an undilute cloud condensing water above its saturation vapor pressure at the moist adiabatic lapse rate LWC would increase by 1.7 g m\(^{-3}\) per km of cloud thickness.

c. Aerosol concentrations

Coarse and accumulation mode aerosol concentration (diameter > 0.1 x 10\(^{-6}\) m) was measured in 2003-2008 (TAMU, Tomlinson et al. 2007), by a Lasair-II optical particle sampler in 2005-2007, and by researchers from University of Washington and Pacific Marine Environmental Laboratory (UW/PMEL) in 2008 (Cover, pers. communication). Aerosol concentration increases toward the coast in all the 20°S longitude cross-sections (Fig. 6a). South America is a source of dust. Its volcanoes, cities, and industries, including copper smelters, are a source of sulfur dioxide, an aerosol precursor gas. Mean concentration and standard deviation west and east of 80°W are displayed either side of the longitude series. There appear to be systematic differences between TAMU and the Lasair-II aerosol concentrations sampled, with lower concentrations measured by the Lasair-II. Histograms of the concentration at 85°W show a wide range of concentrations in 25 cm\(^{-3}\) bins from 0 to 300 cm\(^{-3}\) from the TAMU observations, and a narrower range of
lower concentrations ($75\pm25\ \text{cm}^{-3}$) from the Lasair-II (Figs. 6b and c). The two
ingstruments sampled distinct air masses each year they were used, which may have had
different aerosol concentrations. From the distributions of the TAMU and Lasair-II
aerosol concentrations, there may be a bias between the two sensors. Nevertheless the
gradient of aerosol concentration toward the coast is unambiguous. Aerosol concentration
is $150-250\ \text{cm}^{-3}$ at $75^\circ\text{W}$, and mostly less than $100\ \text{cm}^{-3}$ and relatively uniform west of
$82^\circ\text{W}$.

Hypotheses conceived prior to VOCALS assumed that marine air masses at $85^\circ\text{W}$ were
pristine. In fact the coarse and accumulation mode aerosol concentration at $85^\circ\text{W}$ is about
half the concentration at $75^\circ\text{W}$. Rather than pristine, the aerosol concentration at $85^\circ\text{W}$ is
variable, with standard deviation comparable to that at $75^\circ\text{W}$.

Liquid water path (LWP) decreases toward the coast (Fig. 5) while aerosols increase.
Aerosol concentration decreases as they are removed by precipitation and diluted farther
over the open ocean. LWP increases westward due to thicker low clouds. By separating
LWP at $85^\circ\text{W}$, where most of the aerosol concentrations were measured, into different
aerosol concentration bins, we should be able to see any emergent relationship between
LWP and aerosol concentration. Compared to the anticorrelation of aerosol concentration
and LWP over longitude, there is no systematic relationship between LWP and aerosol
concentration at $85^\circ\text{W}$. Competing cloud-aerosol effects could lead to this lack of a
relationship. Higher aerosol number increases LWP through cloud lifetime effect
(Albrecht 1989), while high LWP clouds remove aerosols by precipitation.
d. Remote sensing of clouds and rain

Instruments of different wavelengths and scanning strategies are used to detect clouds and precipitation. Sensitive lidar (905 nm) ceilometers and cloud radars (3-mm W-band and 8.6-mm MMCR) have a range of order 10 km and detect clouds overhead pointing vertically. The W-band cloud radar used a vertical resolution of 25 m. Larger precipitation particles have higher reflectivity and can be detected with longer wavelength radar (5 cm C-band) at farther ranges. Strong precipitation occupies a small area compared to the widespread southeastern tropical Pacific stratocumulus clouds. With low elevation angle scans the C-band radar samples these infrequent events over a 60 km radius area.

Figure 7 shows the fraction of sky detected above the threshold on the horizontal axis, for research cruise legs along 20°S. Reflectivities lower than the receiver noise have been excluded from the fraction. More clouds are detected as the method becomes more sensitive at lower thresholds.

The linear scale of Fig. 7a emphasizes sensitivity and total cloud amount detected by each instrument. The ceilometer, W-band cloud radar, and C-band precipitation simultaneously sensed clouds and precipitation in 2008 VOCALS leg 2. The ceilometer uses optical backscatter (km\(^{-1}\) steradian\(^{-1}\), top axis) while the radars used reflectivity (dBZ) units. The ceilometer is the most sensitive, detecting cloud fraction of 0.94 above the threshold of 0.05 km\(^{-1}\) sr\(^{-1}\). The W-band cloud radar detects cloud fraction of 0.8
above -32 dBZ, and the C-band radar detects 0.61 cloud fraction above -20 dBZ. The MMCR from 2003 is more sensitive than the W-band used in 2008. Though different clouds were observed in 2003 and 2008, the reflectivity distributions are similar between the MMCR and the W-band cloud radar (Fig. 7b).

The true fraction probably resembles the maximum of the fraction measured by the ceilometer, the W-band, and the C-band radars in Fig. 7b. Only the ceilometer reliably detects clouds with reflectivity less than -30 dBZ. The cloud radars detect cloudy and drizzling columns with reflectivity -30 to -10 dBZ. All three radars had adequate sensitivity and sampling from -25 to 15 dBZ, and their probability densities match.

Radar receivers saturate at high reflectivity. At 500 m range the W-band cloud radar saturates at 33 dBZ (Moran et al. 2011), yet the number columns the W-band radar observes above even 20 dBZ is insignificant. We suspect the vertically pointing radars see too little sky to sample rare strongly precipitating events.

The C-band radar scans over a larger area and samples infrequently occurring cases of high reflectivity. C-band fraction has a wide shoulder with 0.3% of pixels with reflectivity above 20 dBZ. This shoulder decays more slowly with reflectivity than the probability density in the 0-10 dBZ range. More than $10^5$ pixels (0.03% of the total) are over 50 dBZ, which corresponds to roughly $N=10^4$ 1-mm raindrops per meter$^3$ for a typical lognormal drop size distribution. Such reflectivity is unexpectedly high for warm
clouds, and implies warm microphysical processes generate large precipitation drops in southeastern tropical Pacific stratocumulus cloud region.

4. Diurnal Cycle

Cloud layer

Low cloud fraction $c$ is estimated in 10-minute averages from the pulsed lidar ceilometer, which points vertically with a narrow field-of-view. To focus on stratocumulus clouds, only clouds below 2 km are counted in $c$. This excludes infrequent high clouds and noise contamination at higher ranges. The laser ceilometer ranges cloud base reflectivity within a narrow field of view (<1 degree) directly overhead. Averaging the overhead cloud fraction time series over an arbitrarily long time should give a mean cloud fraction that approaches the entire sky cloud fraction. We choose a sampling interval of 10 minutes to obtain a representative overhead cloud fraction and resolve cloud variability. Clouds being so widespread, 71% of 10-minute overhead cloud fractions were totally cloudy ($c=1$), and only 6% were clear ($c=0$), leaving 23% partly cloudy scenes ($0<c<1$).

Ceilometer cloud fraction $c$ for the 7 years is composited on the local hour of the day (Fig. 8a triangles). Mean cloud fraction goes through a single cycle each day, with maximum of 0.96 at 4 in the morning and partly clearing to a cusp-shaped minimum of $c=0.68$ in the afternoon (15 local). Cloud fraction is greater than 0.8 apart from 6 hours of partial clearing in the afternoon. Shading in Fig. 8a shows fraction of observations by hour-of-day that are overcast ($c=1$, dark gray), partly cloudy ($0<c<1$, light gray), or clear
Afternoon clearing occurs with a 36% decrease in the number of overcast observations.

While cloud base remains relatively constant throughout the day, cloud top varies from 1.27 to 1.43 km (Fig. 8b). Cloud thickness is 340 m in the early morning (0-6 h local) and 230 m in the afternoon (12-18 h local, Fig. 8b). If entrainment were solely responsible for the growth of cloud top at night we would expect entrainment of dry air would also thin the cloud layer, evaporating its base. Thickening of the cloud layer when the cloud top is growing indicates that variation in subsidence, and not entrainment alone, contributes to the diurnal variation in cloud top height. Figure 8c shows liquid water path (LWP) mean (circles) and hourly 15, 30, 50, 70, and 85th percentiles. The LWP distribution is positively skewed, biasing the mean toward higher values. The daily cycle of LWP is in phase with cloud thickness, but changes relatively more than cloud thickness.

The ratio of LWP to cloud thickness is the vertical average liquid water content (LWC, g m\(^{-3}\)) of the cloud. Squares in Fig. 8c show vertically averaged LWC for all sky (black), and normalized by cloud fraction for the in-cloud average LWC when a cloud is present (gray). The diurnal cycle of average LWC mostly follows the cycle of LWP, but increases faster around 18 h when LWP increases but clouds stay relatively thin until later in the evening. In fact in-cloud vertical average LWC is lowest at in the early afternoon (2-3 h). Average LWC should increase nearly linearly with cloud thickness. From the moist adiabatic lapse rate (6.5 K km\(^{-1}\)) and change in saturation LWC with temperature (5.5 \(\times\) 10\(^{-4}\) g m\(^{-3}\) K\(^{-1}\)), we estimate vertical average LWC should increase 1.7
g m\(^{-3}\) per kilometer of cloud thickness for an undilute moist saturated air parcel. The least squares fit of average LWC to cloud thickness variations over the diurnal cycle explains 0.6 g m\(^{-3}\) LWC per kilometer of cloud height. The difference is likely the result of several plausible factors: Cloud fraction or the vertical extent of clouds may be overestimated by the remote sensing instruments, clouds are diluted with warm or dry air, and/or liquid water is precipitating out of the cloud. Precipitation must evaporate or reach the surface before it is no longer sensed as LWP by the microwave radiometer.

Subcloud layer variability

Measurements of clouds and surface air temperature and humidity, combined with conservation of heat and humidity, give us a thermodynamic definition of cloud decoupling from surface layer air. This is complementary to definitions of kinematic decoupling, measured as a minimum in vertical velocity variance or negative buoyancy flux below cloud base.

The vertical structure of clouds and the sub-cloud boundary layer varies regularly over the day. Surface lifting condensation level (LCL) is computed by adjusting the observed humidity and temperature to 500 m according to similarity theory, then adiabatically lifting air with those properties until it is saturated. The frequency distribution of displacement of the cloud base height observed by the ceilometer from the LCL is composited for each local hour of the day in Fig. 9. The mode of cloud base–LCL displacement (circles, Fig. 9a) along 20°S reaches a maximum of 375 m at 14-15 local and is near zero at 23-6 local. Figure 9b-d isolates the diurnal cycle in 3 longitude ranges.
The $e^{-1}$ correlation time scale of the raw cloud base-LCL (CB-LCL) time series is 15 hours. Much of this is due to the diurnal cycle. The correlation time scale of the anomalies drops to 6 hours when mean diurnal and zonal variations are removed. Though observations are correlated for several consecutive hours, observations each day are independent of observations from other days.

Despite changes in its amplitude, the pattern of the diurnal cycle is largely the same at different longitudes. At each longitude CB-LCL displacement is relatively low and constant during the night hours, only beginning to increase after sunrise at 6 local. The displacement increases gradually through the morning until the mid afternoon. In each longitude range maximum CB-LCL displacements are seen in the mid-afternoon, with width of the CB-LCL displacement also increased in the afternoon. The mode of CB-LCL displacement drops sharply about an hour before sunset at 17 hours local. All the cloud base height-LCL distributions are bimodal in the late afternoon. Most at 85°W observations at 9-16 local show cloud base height 400-650 m above the LCL, about twice the displacement at 75°W. Even at night, the mode CB-LCL displacement is 100-200 m at 85°W, compared to near zero at 75°W.

Black lines in each panel of Fig. 9 show normalized frequency distributions of CB-LCL displacement for longitude bins in Table 2. Consistent with the mean LCL and cloud base height in Fig. 5, the peak of the CB-LCL distribution rises westward for the longitude bins in Fig. 9. At 75°W (Fig. 9d) cloud base height is a median of 75 m above the LCL,
indicating subcloud mixed layers with minimal humidity gradient and nearly adiabatic lapse rate. At 80°W cloud base height is a median of 150 m above the LCL, and at 85°W median displacement is 240 m. Cloud base-LCL displacement is more broadly distributed to higher values 85° than at 75°W.

Comparisons of cloud base height and LCL suggest three mechanisms of thermodynamic decoupling. The first is a uniform offset of the cloud base distribution upward from the LCL, demonstrated by the upward shift of the mode of the cloud base height-LCL displacement from 75° to 85°W. As one moves westward, this mode of relatively well-coupled clouds are slightly more decoupled from the surface layer. Second, the distributions in Fig. 9 suggest the width of the distribution broadens so that CB-LCL displacement is more variable. The increased width indicates some cloud base parcels have been diluted by larger quantities of warmer, drier air. Dilution by warmer drier air is episodic and affects different clouds by a randomly varying amount. The offset and width of the cloud base height distribution are independent of the LCL. Third, perhaps there is height-dependent decoupling, with CB-LCL displacement correlated to the height of the LCL itself. Diurnal composites in Fig. 10 suggest height-dependent decoupling only in the afternoon, the most decoupled phase of the diurnal cycle.

Figure 10 shows joint distributions of cloud base height and LCL for different longitudes and phases of the diurnal cycle. Columns are sorted by longitude; rows by 6-hour time bin (0-6, 6-12, 12-18, and 18-24 local). Early morning (0-6 local) at 75°W is the most thermodynamically coupled with 500-1000 m cloud bases forming at the LCL (Fig. 10i).
Cloud bases are most displaced from the LCL at 85°W. Afternoons at 75° and 85°W (Figs. 11c,k) seem to show height-dependent decoupling, with larger displacements for higher LCL, but even for these cases the correlations of CB-LCL displacement to the LCL is unconvincing. While height-dependent decoupling would increase the regression slope, wider distribution of cloud base height weakens the regression. Though LCL is a lower bound for cloud base height, regressions of cloud base height on CB-LCL are weak.

Separation of cloud base from the LCL is somewhat coordinated with diurnal changes in cloud fraction and LWP (Fig. 8). On the whole CB-LCL displacement is largest in the afternoon while cloud fraction is lowest. Cloud fraction begins to increases in the late afternoon (15-18 h), a couple of hours before the fall of CB-LCL displacement. This could be an indication that the rapid recoupling of the cloud to the surface layer is driven by buoyancy flux from cloud-top radiative cooling after the stabilizing effects of shortwave radiation are reduced in the late afternoon.

**Diurnal-longitude structure of clouds and tropospheric waves**

Models (Garreaud and Muñoz 2005, Rahn and Garreaud 2010) and satellite observations (O’Dell 2008, Zuidema and Painemal 2009, O’Neill et al. 2011) show a first-harmonic diurnal cycle in liquid water path (LWP) and cloud top height at 20°S 85°W. A strong semidiurnal cycle is found at 75°W. Ship measurements composited hourly by local solar time provide the highest available temporal resolution of the diurnal cycle along 20°S. These data confirm modeling and remote sensing observations of the diurnal and semidiurn oscillations of LWP and cloud top height. Figure 11a-f are contoured
Hovmöller diagrams of key boundary layer and cloud properties as a function of local solar time vs. longitude between 75° and 85°W. Data are composited in the three longitude bins of Table 2. Mean zonal gradients (Fig. 5) exist throughout the diurnal cycle with less cloud, higher SST, higher humidity, higher cloud tops, and higher liquid water to the west.

The diurnal and semidiurnal harmonics explain most of the variance of boundary layer variables (Fig. 12). The first harmonic explains more than 70% of the variance of the diurnal cycle of cloud fraction, SST, and specific humidity (Fig. 12a-c). The diurnal and semidiurnal components of the cloud fraction explain about the same fraction of variance as the cosine of solar zenith angle: 0.84 for diurnal and 0.15 for semidiurnal. At 85°W SST has diurnal and semidiurnal components in about this proportion, perhaps because cloud fraction is low enough for solar absorption to drive SST. Cloud fraction and SST lag incoming solar radiation by about 6 hours.

The first diurnal harmonic explains more than 90% of the variance of the surface specific humidity, but water vapor path (WVP) has a strong semidiurnal rather than diurnal cycle at 75°W. Thus boundary layer cloud and water vapor do not integrate away high-frequency specific humidity variability. Diurnal variations in water vapor also do not follow saturation vapor pressure of the SST. The peak surface specific humidity lags the peak mid-afternoon SST by about 2 hours (Fig. 11b-c), but the specific humidity minimum occurs 2 hours before sunrise, leading SST. Water vapor path is correlated more strongly to cloud top height than to surface specific humidity. Because of the strong
specific humidity interface at the inversion, water vapor path is modulated more by
boundary layer height than internal changes in the boundary layer specific humidity.

As previously found in models and satellite observations, cloud top height, LWP, and
WVP have significant semidiurnal cycles at 20°S 75°W. All are linked more strongly to
the extent of the boundary layer than to intrinsic properties within the boundary layer.
The semidiurnal component of the boundary layer height comes from an offshore
propagating “upsidence” wave in tropospheric velocity, predicted by models to originate
from diurnal heating over the Andes.

We test the ship data for evidence of propagation from 75° to 85°W with cyclic lag
correlation (Fig. 12g-i). LWP and cloud top height at both 75° and 85°W have significant
first-harmonic diurnal cycles in phase with the solar cycle, with no time lag between 75°
and 85°W. Only water vapor path (WVP) has a lag of ~10 hours between 85° and 75°W,
but the correlations between 75° and 85°W are weak because the semidiurnal cycle
dominates at 75°W and the diurnal cycle dominates at 85°W.

An explanation posited for the lack of semidiurnal cycle at 85°W is destructive
interference of the local solar cycle and the propagating upsidence wave. This
explanation requires the semidiurnal component of the wave at 85°W to be the same
magnitude and of opposite phase as the local semidiurnal cycle driven by solar heating.
Gravity wave speeds of 30 m s$^{-1}$ would give a time delay of 10 hours between 75° and
85°W. This is longer than the 6 hours it takes for the semidiurnal wave to reverse phase
locally, inconsistent with the destructive interference hypothesis. We propose that either
spatial interference between waves generated over Peru and Chile, and/or dissipation of
the semidiurnal upsidence wave, could be responsible for the small semidiurnal cycle at
85°W.

5. Surface Cloud Radiative Forcing

Clouds influence the transmission of solar radiation. From downwelling solar radiation
measured by pyranometer $S$ and modeled clear-sky solar radiation $S_1$ (Iqbal 1988) we
compute the full solar transmissivity ratio $t = S/S_0$. Figure 13 shows the transmissivity for
all conditions (black) and cloudy (blue), partly cloudy (gray), and clear (magenta)
conditions as a function of the time of day. Each point represents a 10-minute average
realization along 20°S, 75-85W. Observations near dawn and dusk are less reliable
because $S_0$ is small due to the low solar elevation. We find in practice $t$ can be only be
measured during daylight when the denominator $S_0$ is above 25 W m$^{-2}$. Solar
transmissivity is important only during significant daylight. Curves average realizations
in 10-minute bins by local time of day and are low pass filtered with a 1 hour time scale.

As cloud fraction decreases in the afternoon, transmissivity increases, reaching a
maximum of 0.7 at 13 local. Even though clouds decrease until 15 local, the transmissivity
decreases from 13, perhaps because clouds are more effective scatterers for lower solar
elevation angles. All-sky and partly cloudy transmissivity is similar to cloudy transmissivity,
because of the mostly cloudy conditions. Clear-sky conditional transmissivity is near unity,
but slightly less on average, perhaps because the narrow field of view of the ceilometer
classifies some partly cloudy scenes as clear. It is rare but possible for $t$ to be greater than
unity, in the case that direct sunlight reaches the radiometer through gaps in clouds, while
the clouds scatter additional indirect sunlight into the radiometer.

Variability of downwelling radiation measured on the ship was linked to variation of the
clouds. The downwelling surface solar radiation $S$ is the average of the clear-sky flux $S_0$
and the cloudy-sky flux $S_1$ weighted by the cloud fraction $c$,

$$S = (1 - c)S_0 + cS_1. \quad (1a)$$

The “maximum” cloud forcing

$$S_1 - S_0 = (S - S_0)/c \quad (1b)$$
is the cloud forcing in the case that cloud fraction $c=1$ (Walsh and Chapman 1998). In
practice $(S - S_0)/c$ is subject to error when the cloud fraction is small.

Filled circles in Fig. 14a show mean downwelling solar radiative flux $S$ for full-sky
conditions averaged in 2.5° longitude bins; filled circles in Fig. 14d show mean full-sky
longwave radiation. Boxes in Fig. 14 show sampling standard deviation as a measure of
variability within each longitude bin, and whiskers within the boxes show standard
deviation of the mean. The standard deviation of spatial and temporal variability sampled
by radiometers within a given longitude range is larger than the systematic zonal gradient
of solar radiation.

Open circles in Fig. 14 show mean clear-sky solar flux $S_0$ from the model of Iqbal (1988)
and clear-sky longwave radiative flux $R_0$ from the model of Hare (2005), parameterized
as surface water vapor, temperature, and latitude independent of the in situ radiative flux
observations. The clear-sky model simulates $S_0$ surface solar radiation in the absence of
clouds. All solar fluxes are zero at night. We average the solar flux over each local day to
average out instantaneous changes of $S_0$ due to the diurnal cycle of solar zenith angle. We
require at least 94% of samples to be available for a valid daylight average (5-19 hours
local). We then average over the entire day assuming zero solar radiation at night.

Longwave fluxes do not depend fundamentally on the solar cycle, so we use all 10-
minute means directly. Gray points in Fig. 14 show daily average solar radiation and 10-
minute average longwave radiation that are used for the longitude averages. Radiative
fluxes in Fig. 14b and e are shown only for times when the ceilometer 10-minute average
cloud fraction is overcast ($c=1$). Fluxes in Fig. 14c and f are shown only for clear
overhead ceilometer cloud fraction ($c=0$). The conditional average solar fluxes in Fig.
14b and c require some care not to alias the diurnal cycle.

Since the cloud fraction is not uniform throughout the day, the solar cycle of $S_0$ aliases
onto cloudy- or clear-sky conditional average solar fluxes. To avoid this we statistically
reconstruct cloudy-sky flux $S^-_1$ for non-cloudy times, and clear-sky flux $S^-_0$ for non-clear
times from the diurnal time-dependent conditional cloudy and clear-sky transmissivity,

$$\tilde{t}^-_1 = \langle t(c=1) \rangle^{\text{time}}$$
$$\tilde{t}^-_0 = \langle t(c=0) \rangle^{\text{time}},$$

where $< >^{\text{time}}$ represents a diurnal composite as a function of the time of day.

Curves in Fig. 13 show the diurnal average solar transmissivity $\tilde{t}^-$ (black), clear sky
transmissivity $\tilde{t}^-_0$ (magenta), cloudy-sky transmissivity $\tilde{t}^-_1$ (blue), and partially cloudy
transmissivity (gray). The diurnal cycle is composited on local time in 10-minute intervals and low pass filtered to remove variability shorter than 1 hour.

The empirical reconstructions of cloudy and clear conditional solar fluxes are then the product of the empirical transmissivity curves and $S_0$ derived from a clear-sky model:

$$S'_{\text{c}} = \tau'_c S_0$$

$$S'_{\text{o}} = \tau'_o S_0.$$ (2)

The conditional clear- and overcast-sky solar flux reconstructions are suitable for averaging daily without diurnal aliasing. Fig. 14b and c show daily mean reconstructions averaged in longitude, $\langle S'_{\{0,1\}} \rangle^{\text{day}}_{\text{lon}}$. Since the sky is completely cloudy 67% of the time, the cloudy conditional average is quite similar to the full-sky average. Overcast solar fluxes (Fig. 14b) are approximately 25 W m$^{-2}$ lower, and only half as variable as compared to full-sky (Fig. 14a). Downwelling longwave fluxes in cloudy conditions are 10 W m$^{-2}$ stronger than the full-sky average.

In clear conditions fluxes collapse approximately to the clear sky model (Fig. 14c,f).

Solar clear-conditional fluxes are slightly weaker than the clear sky model due to undersampling of cloud fraction by the narrow field of view of the ceilometer. Clouds undetected by the ceilometer reduce the solar radiation reaching the pyranometer. This also explains average transmissivities less than 1 when the ceilometer detected clear skies (Fig. 13). Longwave fluxes in clear conditions are statistically indistinguishable from the clear sky model.
De Szoeke et al. (2010) find CMIP3 general circulation models all overestimate downwelling solar radiation by at least 40 W m$^{-2}$ in the region, partially compensated by excessive net upwelling longwave radiation. We examine the effect of clouds on the radiative fluxes by computing downwelling cloud radiative forcing from the methods and data of Fig. 14. Incident solar cloud forcing $S - S_0$ is the difference between the observed and modeled clear-sky downwelling solar and longwave radiation (Fig. 10a,d). Table 3 summarizes multi-cruise-year average radiative fluxes and cloud forcing for observations within the region 18.5-21.5°S, 73.75-86.25°W (all longitude bins along 20°S). Average solar cloud forcing $S - S_0$ is -133 W m$^{-2}$.

We estimate maximum solar cloud forcing (MSCF) from the 1b and from the conditional cloudy solar reconstruction $S^-_1$ (Table 3). Daily averaging $<S - S_0>$ and $<c>$ before dividing their quotient reduces problems associated with small denominators $c$ and prevents diurnal aliasing, but daily average $<c>$ is not representative of the clouds affecting the radiation. For example, about half the contribution to $<c>$ is overnight while $S - S_0$ is zero. Thus cloud fraction at night affects the maximum cloud forcing $<S - S_0>/<c>$ even though it has no effect on the solar radiation. When instantaneous cloud fraction is used instead, maximum cloud forcing estimates $(S - S_0)/c$ are ill conditioned by small or zero cloud fraction $c$. In light of these problems we employ the conditionally reconstructed cloudy sky flux $S^-_1$ to get the maximum cloud forcing. Reconstructed $S^-_1$ is likely to be representative of all clouds because of the large number of fully cloudy realizations composited in the conditional cloudy reconstruction.
The research cruise in 2004 took place in December, later in the seasonal cycle than the other cruises in October-November. Observed and modeled solar fluxes are comparable (within 10 W m$^{-2}$) between the 2004 December cruise and the other years. However, cloud fraction is considerably less in December, and the increased number of clear observations make maximum cloud forcings calculated from 1b unreliable. This method estimates a physically absurd average cloud forcing of $-488$ W m$^{-2}$ for 2004 December, impossibly 100 W m$^{-2}$ stronger than the clear-sky solar flux. The conditionally reconstructed cloud forcing $S_1^- - S_0$ is minimally affected by removing December data from the all-cruise average because the conditional reconstruction of transmissivity and hence $S_1^-$ are representative of October-November conditions. We suspect $S_1^-$ is not representative of clouds in December, and there are not enough cloudy data in December 2004 to make an independent radiative reconstruction for that season. Thus 2004 December reconstructed maximum solar cloud forcings $S_1^- - S_0$ in Table 3 and $R_1^- - R_0$ in Table 4 may be too weak to explain the stronger cloud forcing (even when there were fewer clouds) in December.

Surface thermal longwave cloud forcing is about $+60$ W m$^{-2}$ (Table 4). While small cloud fraction is still a problem, longwave radiation has a weak diurnal cycle and so is not subject to diurnal aliasing effects. The empirical longwave $R_1^-$ is simply the conditional average for cloudy skies. Because the diurnal variability is weak, daily average cloud fraction is representative, so $<R_0-R>/<c>$ is a good estimate of maximum longwave cloud forcing (MLCF). MLCF computed this way and from the conditional cloudy sky radiation $R_1^- - R_0$ agree well, at about 70 W m$^{-2}$. 
Surface longwave radiation is strongly correlated with the amount of clouds, indicating variations in emission of the clear atmosphere and in cloud brightness temperature only weakly influence the downwelling radiation. Figure 15, a scatter plot of surface downwelling longwave radiation versus cloud fraction, suggests surface longwave radiation is a sum of clear sky and a nearly constant maximum cloud forcing weighted by the cloud fraction,
\[ R = R_0 + c(R_1 - R_0). \] (3)
Small dots in Fig. 15 indicate cloud fractions of 0 or 1. Radiative forcing for partial cloud fraction lies on a mixture line between clear and cloudy end points. Cloud fraction explains 75% of the variance of downwelling longwave radiation. Modeled clear-sky longwave radiation explains only 16% of the variance of observed radiation in clear conditions. Clouds undetected by the narrow-beam ceilometer modify longwave radiation from its clear-sky value. Only approximately 10% of the surface longwave radiation variance could be explained by variations in emission from the bases of the clouds themselves. Considering this insignificant fraction of variance, it is not surprising that cloud base brightness temperature estimated from the longwave radiation is uncorrelated with ceilometer cloud base height retrievals. Apparently slight changes in either the clear-sky or cloud radiance temperatures have only a small effect on the longwave radiation, while cloud fraction modulates the large difference between clear and cloudy longwave radiation.
We show daily mean longwave and solar cloud radiative forcing against daily mean cloud fraction in Fig. 16. Daily mean cloud fraction is mostly greater than 0.88 and always greater than 0.3. As expected from the strong correlation of 10-minute mean downwelling longwave radiation to cloud fraction (Fig. 15), and from relatively constant clear-sky downwelling longwave radiation, daily cloud fraction explains daily mean longwave cloud forcing well (Fig. 16, open circles). Cloud forcing is the product of the maximum cloud forcing and the cloud fraction; for longwave radiation, \( R - R_0 = c(R_1 - R_0) \).

Assuming the maximum cloud forcing is constant, we model the effect of cloud fraction on the cloud radiative forcing with straight lines connecting the maximum longwave and solar cloud forcing at \( c=1 \) to zero cloud forcing for \( c=0 \). The correlation of daily longwave cloud forcing with cloud fraction is \( r=0.93 \).

Daily solar cloud radiative forcing is explained less well by daily cloud fraction (dots) than longwave cloud forcing, with a correlation coefficient of only \( r=-0.76 \). The solar cloud forcing in Fig. 16 suggests a nonlinear dependence on cloud fraction, with stronger maximum cloud forcing on days when there is more cloud fraction. The nonlinear dependence on cloud fraction is explained by considering cloud forcing as a product of cloud fraction, transmissivity, and clear-sky solar radiation:

\[
S - S_0 = c(S_1 - S_0) = c(t - 1)S_0.
\] (4)

Solar cloud forcing depends nonlinearly on daily mean cloud fraction because of the negative correlation of \( c \) with \( S_0 \) due to afternoon clearing during strong clear-sky solar radiation. The solar-weighted daily cloud fraction, \( c_S = \langle S_0 c \rangle / \langle S_0 \rangle \) is more representative of the cloud fraction when solar radiation is actually incident. Solar-
weighted daily cloud fraction is usually less than ordinary cloud fraction because most clearing is observed in the afternoon (Fig. 8). Crosses in Fig. 16 show somewhat more linear dependence of solar cloud radiative forcing on the solar-weighted cloud fraction compared to ordinary cloud fraction. The correlation coefficient of solar cloud forcing with solar-weighted cloud fraction is \( r = -0.83 \). Additionally, stratocumulus clouds are expected to be more opaque when cloud fraction is higher and geometrically and optically thinner when they are more patchy and clearing. This negative correlation of cloud fraction \( c \) and transmissivity \( t_1 \) also results in nonlinear dependence of cloud forcing on cloud fraction in (4). Straight lines in Fig. 16 connect zero cloud forcing to the maximum cloud forcing from Table 4.

We compare daily mean solar and longwave cloud forcing in the cloud forcing phase diagram of Fig. 17. Filled circles are from October-November cruises to the 20°S, 75-85°W region. We assume there is zero longwave and solar cloud forcing when there are no clouds, and therefore the model line (black) extends from the origin to the average maximum cloud forcing, as in Fig. 16. Our observations agree with the regression line from 4 years of buoy observations at 20°S, 85°W (Cronin et al. 2006), and fall in the stratocumulus and trade cumulus region of the cloud-forcing phase space (Fairall et al. 2008). Even for October-November, when the cloud regime is strongly stratocumulus, observed cloud forcing falls in a wide range of the solar-longwave phase space. Daily average cloud forcing from outside the 20°S region (crosses), or from December (circles) do not show significantly different solar vs. longwave cloud forcing behavior.
6. Surface Cloud Forcing in General Circulation Models

Solar radiation is the only warming term over most of the ocean. Longwave radiation damps SST anomalies. Cloud radiative forcing modulates solar and longwave radiation, and is an important term in the upper ocean heat budget, especially beneath the extensive tropical stratocumulus cloud decks. De Szoeke et al. (2010) ranked models by solar radiation absorbed by the ocean. Here we focus specifically on attributing model radiative forcing anomalies to errors in cloud radiative forcing. We evaluate cloud radiative forcing in ocean-atmosphere coupled general circulation models (GCMs) from the third Coupled Model Intercomparison Project (CMIP3) with the cloud radiative forcing observed along 20°S from the 7 years of cruises.

Since the surface heat budget responds to the net radiative heat flux, surface cloud forcing is reduced slightly by surface absorption coefficients. For longwave this is the emissivity of the sea surface $\varepsilon_s = 0.97$; for solar radiation it is the complement of the surface albedo $1-\alpha_s = 0.945$. Thus net longwave and solar cloud forcings are respectively defined,

\[ \text{LCF} = \varepsilon_s (R - R_0) \]
\[ \text{SCF} = (1 - \alpha_s) (S - S_0). \]

Figure 1 compares surface longwave and shortwave cloud forcing climatology for October-November from 15 CMIP3 coupled GCMs with the ship observations. In situ observations from the WHOI buoy at 20°S, 85°W (Cronin et al. 2006), remote sensing from the ISCCP Flux Data set, and the 7 years of ship observations agree well. For the
ISCCP FD and Stratus buoy data sets we use MODIS cloud fraction (Fig. 1). Surface solar cloud forcing in all CMIP3 models (-20 to -90 W m\(^{-2}\)) lacks the observed strength (-120 W m\(^{-2}\)). Longwave cloud forcing offsets about 50% of solar cloud forcing for observations and for the consensus of models. Longwave errors offset solar errors in the same proportion. Correlated among models at \(r=-0.63\), longwave cloud forcing errors compensate about half of shortwave cloud forcing errors. Two models have 60 W m\(^{-2}\) too-weak surface solar cloud forcing, yet have longwave cloud forcing close to the observed, resulting in 60 W m\(^{-2}\) too much downwelling radiation absorbed at the surface.

Surface cloud forcing errors are associated with deficiencies of cloud fraction. Model longwave and solar cloud forcing are correlated to cloud fraction at 0.82 and -0.72. The models mimic the proportionality of cloud forcing to cloud fraction (Fig. 18), falling near the line between the observed cloud forcing and cloud fraction and zero cloud forcing for clear skies. This suggests that simulated clouds have the right cloud forcing when present, but too few clouds in the models result in too little cloud forcing cooling.

7. **Summary**

Models have too strong net radiation because they have too few clouds. On the whole, simulated clouds have approximately the right cloud radiative forcing when present. Longwave cloud forcing offsets 50% of shortwave cloud forcing in observations. Conveniently longwave cloud forcing error also offsets 50% of shortwave cloud forcing error in models. In CMIP3 models, errors in cloud amount dominate any errors that could be attributed to cloud albedo or aerosol indirect effect issues.
Observations show that daily average downwelling longwave radiation and longwave cloud forcing are proportional to the cloud amount, indicating relatively constant maximum longwave cloud forcing, i.e. constant radiation from clouds when present. Surface solar radiative forcing responds linearly to cloud amount to some degree, but has a nonlinear tendency for stronger solar radiative forcing at higher cloud fraction, partly explained by clearing during high clear-sky solar flux in the afternoon.

Daily solar cloud forcing along 20°S falls within a wide range about the stratocumulus regime ($|S-S_0| = 2(R-R_0)$) of the longwave-solar forcing phase diagram (Fairall et al. 2008). The range stretches from the warmer trade cumulus regime ($|S-S_0| = 3(R-R_0)$) to the cooler midlatitude cloud regime, with a relatively smaller proportion of solar to longwave cloud forcing. The 20 days with anomalously low ratio of solar to longwave cloud forcing ($|S-S_0| < 2(R-R_0) - 10 \text{ W m}^{-2}$) have about 50 W m$^{-2}$ weaker maximum solar cloud forcing but almost no change in their longwave cloud forcing (gray symbols, Fig. 16). A 3% decrease in their median cloud fraction accounts for some of the reduction in solar forcing, and the reduction in solar forcing is amplified because the reduction of clouds is mostly in the afternoon.

The midlatitudes have colder atmosphere and weaker solar radiation than the tropics. Clear periods between midlatitude storms correspond to cold, dry, descending air and less emissive atmospheric conditions. The relative increase in surface longwave cloud forcing puts the storm tracks in the $|S-S_0| < 2(R-R_0) - 10 \text{ W m}^{-2}$ region of the cloud forcing phase.
diagram (Fig. 17). Unlike midlatitude clouds, the tropical stratocumulus clouds observed in this study have weaker solar maximum cloud forcing than typical stratocumulus and similar longwave cloud forcing. This might result from tenuous clouds that allow a large amount of solar radiation through but are nevertheless strongly emissive in the thermal infrared.

The effects of aerosols on the solar radiation could account for some variance in the solar cloud forcing, perhaps explaining deviations from the typical stratocumulus cloud-forcing phase space regime (Fig. 17). Overestimation of the clear sky solar flux by not considering the aerosol direct effect diminishing clear sky radiation would result in overestimation of our solar cloud forcing. Cloud albedo aerosol indirect effect (Twomey 1974) affects solar flux measured by the solar radiometers, increasing the strength and variability of the solar cloud forcing, accounting for more variance from the stratocumulus regime in the cloud forcing phase space.

Thermodynamic atmospheric soundings are remarkably constant over the 7 years of research cruises to 20°S. A 14-level idealized sounding based on 487 soundings along 20°S is available for model studies (Table 1).

Cloud base-LCL displacement is a thermodynamic index of decoupling. This displacement increases both westward with longitude, and during the daylight hours, as summarized by the schematic of Fig. 19. Larger displacements of 500-1000 m also grow increasingly common to the west. Afternoon CB-LCL displacement is 400 m greater than
the displacement at night. The westward rise of cloud base coincides with higher and
more variable CB-LCL displacement. For deep subcloud layers in the afternoon, cloud
base is less likely to be found near the LCL.

Jones et al. (2011) find decoupling is correlated to cloud thickness. For all 20°S
observations we find cloud top height, cloud base height, and LCL to be correlated (Table
5). Cloud top and cloud base height, correlated at $r=0.7$, both explain CB-LCL
displacement, cloud base more so than cloud top (CB-LCL depends explicitly on CB).
Cloud top height does not explain any additional variance in the CB-LCL displacement,
but it is consistent with deeper boundary layers being more decoupled. The gradient in
longitude explains some but not all of the variance in cloud top, cloud base, LCL, and
CB-LCL displacement.

Semidiurnal cycles of variables related to boundary layer height are observed at 75°W,
but not 85°W. Lag correlations of diurnal cycles in the ship data do not show coherent
propagation from 75° to 85°W. Moreover the phase of such tropospheric gravity waves is
not consistent with constructive interference with the local diurnal cycle at 75°W and
destructive interference at 85°W. Other mechanisms by which the semidiurnal cycle in
variables related to boundary layer height can weaken at 85°W relative to 75°W include
dissipation of waves as they propagate offshore, and spatial interference of diurnally-
forced waves emitted from different source locations, e.g. from the Peruvian and Chilean
Andes on either side of the Arica Bight. Though the ship data provide excellent time
resolution of the diurnal cycle, the spatial sampling of the composites is rather coarse.
Satellite observations may be able to resolve spatial interference from the waves. Stability in the inversion is an effective waveguide for tropospheric gravity waves. Energy of waves traveling along the inversion could be dissipated by entrainment mixing free tropospheric air into the boundary layer at the inversion.

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References

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Figure 1. October-November satellite sea surface temperature (SST) and cloud fraction climatology in the southeastern tropical Pacific Ocean from the AMSR-E and MODIS instruments, respectively. Colored lines indicate tracks of NOAA research cruises.
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Figure 4. Zonal ($u$) and meridional ($v$) wind distributions with height (shades), and median wind profile (black line) from the 20°S soundings. Dashed lines indicate the sampling standard deviation of wind profiles.

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Figure 7. (a) Fraction of sky with reflectivity above a threshold, excluding noise. (b) Probability density as the fraction of full sky per bin (width 2 dBZ or 0.0175 km$^{-1}$ sr$^{-1}$) on a logarithmic scale. Vertically pointing instruments such as the W-band cloud radar (dot-dashed), millimeter cloud radar (MMCR, gray), and ceilometer (triangles) are compared to the C-band scanning radar (solid black). The threshold for the ceilometer is in optical backscatter units (km$^{-1}$ sr$^{-1}$, top axis).

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Figure 14. Solar and longwave surface downwelling radiation measured along 20°S
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cumulus regimes (gray, Fairall et al. 2008), and the regression at the Stratus buoy (dashed, Cronin 2006) are shown.

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<table>
<thead>
<tr>
<th>height (km)</th>
<th>potential temperature (K)</th>
<th>RH (%)</th>
<th>u (m s(^{-1}))</th>
<th>v (m s(^{-1}))</th>
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<tbody>
<tr>
<td>0.0</td>
<td>290.0</td>
<td>70</td>
<td>-4</td>
<td>3.0</td>
</tr>
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<td>290.0</td>
<td>71</td>
<td>-5</td>
<td>4.5</td>
</tr>
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<td>0.9</td>
<td>290.5</td>
<td>99</td>
<td>-5</td>
<td>4.5</td>
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<td>1.4</td>
<td>291.5</td>
<td>99</td>
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<td>2.0</td>
</tr>
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<td>1.0</td>
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<td>5.0</td>
<td>322.0</td>
<td>2</td>
<td>4</td>
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<td>8.0</td>
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<td>17.8</td>
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<td>1</td>
<td>-2.0</td>
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<td>470.0</td>
<td>4</td>
<td>-5</td>
<td>-0.5</td>
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Table 2. Total and minimum diurnal (per local hour) sampling for the composite 20°S transect.

<table>
<thead>
<tr>
<th>nominal longitude</th>
<th>85°W</th>
<th>80°W</th>
<th>75°W</th>
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<tbody>
<tr>
<td>longitude range (°W)</td>
<td>86.25-83.75</td>
<td>83-75-76.25</td>
<td>76.25-73.25</td>
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<tr>
<td>total hours</td>
<td>1118</td>
<td>456</td>
<td>438</td>
</tr>
<tr>
<td>minimum hours per local hour</td>
<td>45</td>
<td>17</td>
<td>17</td>
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Table 3. Mean surface clear-sky solar radiation, solar radiation, cloud forcing, and estimates of maximum cloud forcing ± the standard error of the mean (W m\(^{-2}\)). Angle brackets indicate daily averages. Columns show the average for all 7 years of cruises, the average for the 6 years of cruises in October-November, and the average of the cruise in 2004 December. Standard errors of the mean less than the least significant digit (e.g. 1 W m\(^{-2}\)) are not listed.

<table>
<thead>
<tr>
<th>solar (W m(^{-2}))</th>
<th>all cruises</th>
<th>no December</th>
<th>2004 December</th>
</tr>
</thead>
<tbody>
<tr>
<td>(S_0)</td>
<td>358±1</td>
<td>357±10</td>
<td>366±2</td>
</tr>
<tr>
<td>(S)</td>
<td>225±5</td>
<td>226±5</td>
<td>215±12</td>
</tr>
<tr>
<td>(S-S_0)</td>
<td>-133±5</td>
<td>-131±5</td>
<td>-150±12</td>
</tr>
<tr>
<td>mean (c)</td>
<td>0.86</td>
<td>0.88</td>
<td>0.74</td>
</tr>
<tr>
<td>((S-S_0)/c)</td>
<td>-219±19*</td>
<td>-158±4**</td>
<td>-488±77</td>
</tr>
<tr>
<td>(&lt;S-S_0&gt;/c)</td>
<td>-153±5</td>
<td>-146±5</td>
<td>-205±17</td>
</tr>
<tr>
<td>(S_{-}_S_0)</td>
<td>-162±2</td>
<td>-162±2</td>
<td>-167±2</td>
</tr>
</tbody>
</table>

*in all cruises 116 hours of observations (of 1963 hours) were disregarded for having \(c=0\), resulting in unbounded \((R-R_0)/c\).  

**in Oct-Nov 115 of 1727 hours had \(c=0\).
Table 4. Surface longwave clear-sky radiation, longwave radiation, cloud forcing, and estimates of maximum cloud forcing (W m\(^{-2}\)) as in Table 2.

<table>
<thead>
<tr>
<th>longwave (W m(^{-2}))</th>
<th>all cruises</th>
<th>no December</th>
<th>2004 December</th>
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<tbody>
<tr>
<td>(R_0)</td>
<td>316</td>
<td>315</td>
<td>329</td>
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<tr>
<td>(R)</td>
<td>375</td>
<td>373±1</td>
<td>391±1</td>
</tr>
<tr>
<td>(R - R_0)</td>
<td>59±1</td>
<td>59±1</td>
<td>62±1</td>
</tr>
<tr>
<td>((R - R_0)/c)</td>
<td>75±2</td>
<td>66±1</td>
<td>134±13</td>
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<tr>
<td>(&lt;R - R_0&gt;/c)</td>
<td>68</td>
<td>66</td>
<td>84</td>
</tr>
<tr>
<td>(R^{+1} - R_0)</td>
<td>69</td>
<td>69</td>
<td>69±1</td>
</tr>
</tbody>
</table>
Table 5. Correlations of cloud geometry: cloud top, cloud base (CB), lifting condensation level (LCL), cloud thickness (top-CB), and CB-LCL displacement. Correlations weaker than 0.3 (in parentheses) do not differ from zero with 95% statistical significance, assuming an autocorrelation time scale of 6 hours for 43 degrees of freedom.

<table>
<thead>
<tr>
<th></th>
<th>top</th>
<th>CB</th>
<th>LCL</th>
<th>top-CB</th>
<th>CB-LCL</th>
<th>longitude</th>
</tr>
</thead>
<tbody>
<tr>
<td>top</td>
<td>1</td>
<td>0.73</td>
<td>0.45</td>
<td>(0.26)</td>
<td>0.37</td>
<td>-0.37</td>
</tr>
<tr>
<td>CB</td>
<td>0.73</td>
<td>1</td>
<td>0.52</td>
<td>-0.48</td>
<td>0.60</td>
<td>-0.46</td>
</tr>
<tr>
<td>LCL</td>
<td>0.45</td>
<td>0.52</td>
<td>1</td>
<td>(-0.15)</td>
<td>-0.37</td>
<td>-0.31</td>
</tr>
<tr>
<td>top-CB</td>
<td>(0.26)</td>
<td>-0.48</td>
<td>(-0.15)</td>
<td>1</td>
<td>-0.38</td>
<td>(0.17)</td>
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<tr>
<td>CB-LCL</td>
<td>0.37</td>
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<td>-0.38</td>
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<td>-0.46</td>
<td>-0.31</td>
<td>(0.17)</td>
<td>(-0.20)</td>
<td>1</td>
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