North Pacific Cloud Feedbacks Inferred from Synoptic-Scale Dynamic and Thermodynamic Relationships

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ABSTRACT

Daily satellite cloud observations and reanalysis dynamical parameters are analyzed to determine how midtropospheric vertical velocity and advection over the sea surface temperature gradient control midlatitude North Pacific cloud properties. Optically thick clouds with high tops are generated by synoptic ascent, but two different cloud regimes occur under synoptic descent. When vertical motion is downward during summer, extensive stratocumulus cloudiness is associated with near-surface northerly wind, while frequent cloudless pixels occur with southerly wind. Examination of ship-reported cloud types indicates that midlatitude stratocumulus breaks up as the boundary layer decouples when it is advected equatorward over warmer water. Cumulus is prevalent under conditions of synoptic descent and cold advection during winter. Poleward advection of subtropical air over colder water causes stratification of the near-surface layer that inhibits upward mixing of moisture and suppresses cloudiness until a fog eventually forms. Averaging of cloud and radiation data into intervals of 500-hPa vertical velocity and advection over the SST gradient enables the cloud response to changes in temperature and the stratification of the lower troposphere to be investigated independent of the dynamics. Vertically uniform warming results in decreased cloud amount and optical thickness over a large range of dynamical conditions. Further calculations indicate that a decrease in the variance of vertical velocity would lead to a small decrease in mean cloud optical thickness and cloud-top height. These results suggest that reflection of solar radiation back to space by midlatitude oceanic clouds will decrease as a direct response to global warming, thus producing an overall positive feedback on the climate system. An additional decrease in solar reflection would occur were the storm track also to weaken, whereas an intensification of the storm track would partially cancel the cloud response to warming.

1. Introduction

Clouds have a large impact on earth’s radiation budget because they typically reflect more shortwave (SW) radiation and emit less longwave (LW) radiation to space than does the unobscured surface. Despite their importance, the sign and magnitude of cloud feedbacks on the climate system are some of the largest uncertainties of future climate change. It is presently not known whether cloud properties will change so as to exacerbate or mitigate global warming (Moore et al. 2001). One of the primary reasons global climate models (GCMs) produce substantially different predictions of the magnitude of future global warming is that they do not correctly and consistently simulate clouds. Not surprisingly, differences in climate sensitivity are closely related to particular aspects of various cloud parameterizations employed in the GCMs (Senior and Mitchell 1993). Errors in simulated cloud variability on interannual and longer time scales are directly related to errors in the representation of processes controlling cloud properties on daily time scales (Norris and Weaver 2001). The limitations of GCM studies thus motivate investigation of cloud feedbacks using observational data. Careful documentation of the dependence of cloud properties on related dynamic and thermodynamic parameters using daily observations will provide insight into how clouds might respond to dynamic and thermodynamic changes associated with global warming.

Clouds over midlatitude oceans are of particular interest since during summer they produce net cloud radiative forcing that is more negative than anywhere else...
on earth (Harrison et al. 1990). These clouds are associated with extratropical cyclones in the storm track. Synoptic ascent generates thick frontal cloudiness that can reflect as much as 200 W m$^{-2}$ of SW radiation more than clear sky. Extensive stratocumulus cloudiness under synoptic descent following the cold front typically reflects 100 W m$^{-2}$ more than clear sky (Weaver and Ramanathan 1997). Despite the radiative importance of these clouds, GCMs poorly simulate them (Klein and Jakob 1999; Tselioudis and Jakob 2002). Common errors are overproduction of cloud optical thickness and height in the ascent regime and underproduction of cloud cover in the descent regime (Norris and Weaver 2001).

In addition to vertical motion, advection over the sea surface temperature (SST) gradient separating subpolar and subtropical gyres changes low-level cloud properties, especially during summer. The climatological distribution of surface-observed low cloud types suggests that poleward advection of warm subtropical air over colder water causes stratification of the near-surface layer that leads to the suppression of cumulus clouds and eventually the formation of fog and stratus. Equatorward advection of midlatitude stratocumulus over warmer water causes decoupling of the boundary layer and a transition to cumulus (Norris 1998a,b). Anomalies in SST cause shifts in the cloud transition region and consequently anomalies in low-level cloud cover (Norris and Leovy 1994). These cloud-cover anomalies can in turn radiatively reinforce the SST anomalies (Norris et al. 1998). Another thermodynamic factor besides SST affecting cloud properties is the static stability of the lower troposphere (Klein and Hartmann 1993; Weaver and Ramanathan 1997). Ascertain the individual influence of each of the above meteorological parameters on cloud properties is complex since interannual anomalies in the strength and location of the storm track covary with SST anomalies (Norris 2000), and SST anomalies affect the static stability of the lower troposphere.

The strategy of the present study is to separately characterize dynamical and thermodynamical effects on cloud properties using a method similar to that of Bony et al. (2004). The North Pacific region is examined since it is better sampled than the Southern Ocean and is more zonally uniform than the North Atlantic, which simplifies the analysis. The time period of the investigation is the months of January, April, and July during 1984–2001 and October during 1983–2000. Daily values of cloud amount, cloud optical thickness (τ), and cloud-top pressure (CTP) are averaged within small intervals of midtropospheric vertical velocity and advection of the near-surface wind over the SST gradient to determine how synoptic forcing in conjunction with the SST gradient control cloud properties. The cloud data in each vertical velocity and advection interval are then subdivided into terciles of anomalously cold or warm SST and anomalously strong or weak lower-tropospheric static stability. The average of the data within the middle terciles represents “normal” conditions, and the difference between the averages of the upper and lower terciles represents how cloud properties change with temperature or stratification. This procedure effectively calculates the partial derivatives of cloud fraction, τ, and CTP with respect to vertical motion, advection, SST/atmospheric temperature, and static stability. Long-term cloud feedbacks on the climate system may be inferred from these synoptic-scale relationships. Although present-day observations cannot describe the state of the future climate, the results of this study can be used to estimate, for example, how clouds might respond were SST to increase or the intensity of the storm track were to change. They can also be used to constrain GCM simulations of cloudiness over midlatitude oceans.

2. Data

The International Satellite Cloud Climatology Project (ISCCP) Data (D1) product provides values of cloud fraction, mean visible τ, and mean CTP in 2.5° × 2.5° grid boxes every 3 h (Rosso et al. 1996; Rosso and Schiffer 1999). These properties were retrieved from geostationary and polar-orbiting weather satellites starting in July 1983 and currently ending in September 2001. Besides gridbox mean values, the D1 data include the frequency of pixels in each grid box classified into six τ and seven CTP intervals, subsequently reduced to the same three τ and three CTP intervals used by ISCCP D2 data. Because each 4–7-km pixel is assumed to be completely cloud filled or cloud free, these frequencies correspond to the fractional coverage of clouds with low, middle, or high tops and thin, intermediate, or thick τ. High clouds are defined as those with tops above the 440-hPa level, midlevel clouds with tops between 680 and 440 hPa, and low-level clouds with tops below the 680-hPa level. Thin clouds are defined as those with τ < 3.55, intermediate clouds with 3.55 < τ < 22.63, and thick clouds with τ > 22.63. Although ISCCP data are available every 3 h, only the hours closest to local noon (0000 and 2100 UTC over the North Pacific) are used in order to minimize solar zenith angle and cloud retrieval errors.

One disadvantage of ISCCP cloud observations is that they do not accurately recognize the morphological characteristics of low-level clouds (Hahn et al. 2001).
Knowledge of cloud morphology, determined through visual identification of low cloud type, is useful because it provides qualitative information about vertical structure and meteorological processes in the boundary layer (Norris 1998a). For this reason the satellite data are supplemented by surface synoptic reports collected in the Extended Edited Cloud Report Archive (EECRA; Hahn and Warren 1999) and originally processed from the Comprehensive Ocean–Atmosphere Data Set (COADS; Woodruff et al. 1987). These reports, made primarily by Volunteer Observing Ships, include cloud type at low, middle, and high levels ($C_L$, $C_M$, and $C_H$) and present weather (ww) identified according to the World Meteorological Organization (WMO) code (WMO 1987). To increase the number of observations contributing to $2.5^\circ \times 2.5^\circ$ gridbox averages, data from all observing hours (0000, 0600, 1200, and 1800 UTC) are used. Since it was necessary to discard many nighttime observations due to poor illumination (Hahn et al. 1995), the diurnal mean is biased toward daytime. EECRA data are currently not available after 1997.

Impacts of cloud properties on the top-of-atmosphere radiation budget are assessed using daily mean values of SW, LW, and net cloud radiative forcing (CRF) in $2.5^\circ \times 2.5^\circ$ grid boxes, previously used in Norris and Weaver (2001), and originally obtained from the Earth Radiation Budget Experiment (ERBE; Barkstrom et al. 1989). These data are available for 1985–89. CRF was calculated by subtracting daily mean clear-sky radiative flux from daily mean all-sky radiative flux in each grid box, and monthly mean values of clear-sky flux were used when daily clear-sky values were not available due to extensive cloud cover (C. Weaver 2004, personal communication). It should be noted that ERBE daily means were actually calculated from a few instantaneous measurements during the day (Barkstrom 1984), and thus could include errors associated with aliasing of diurnal cloud variations. This is unlikely to produce a systematic bias in the present study, however, because the satellite orbit precesses and values from widely separated days are averaged into the composites. Moreover, diurnal variations in cloud cover are not large over the midlatitude North Pacific (Cairns 1995; Rozendaal et al. 1995). Radiative flux is defined as positive downward such that SW CRF is negative and LW CRF is positive.

The National Centers for Environmental Prediction (NCEP) Department of Energy (DOE) Atmospheric Model Intercomparison Project (AMIP)-II Reanalysis (R2) provides values of vertical velocity, horizontal winds, and temperature on standard pressure levels at $2.5^\circ \times 2.5^\circ$ grid spacing every 6 h (Kanamitsu et al. 2002). These are based on a variety of measurements of the atmospheric state brought into dynamical consistency by a numerical weather prediction model. Even though vertical velocity is diagnosed by the model rather than directly measured, negligible differences occur between statistical cloud relationships based on NCEP Reanalysis I, NCEP Reanalysis II, and ECMWF Re-Analysis vertical velocity (as was also found by Norris and Weaver 2001). Since the reanalysis grid is offset from the ISCCP grid by $1.25^\circ$ in latitude and longitude, reanalysis parameters were bilinearly interpolated to the ISCCP grid and linearly interpolated to ISCCP times.

Weekly mean SST values for $1^\circ \times 1^\circ$ grid boxes are obtained from version 2 of the Reynolds optimal interpolation (OI) dataset, a combination of in situ and satellite measurements of SST (Reynolds et al. 2002). These data were also bilinearly interpolated to the ISCCP grid. Lower-tropospheric static stability (LTS) is defined as $\theta_{700} - \theta_{SST}$, where $\theta_{700}$ is the potential temperature at the 700-hPa level and $\theta_{SST}$ is the potential temperature at the surface (calculated from SST and SLP).

3. Spatial composites

Figure 1 shows climatological July and January distributions of total cloud amount, SST, and 1000-hPa wind. Cloud amount and SST are close to zonally uniform over the central North Pacific, and the gradient in cloud amount is collocated with the gradient in SST. In general, cloudiness over the North Pacific is most prevalent where the ocean is cold and least prevalent where it is warm. Maximum midlatitude cloud amount is slightly greater during July than during January, but the area of extensive cloud amount does not extend as far equatorward during July. Although the climatological near-surface wind is westerly or southwesterly over the region of the strongest SST gradient, the frequent passage of extratropical cyclones causes large variability in daily wind direction.

The synoptic cloud response to advection over the SST gradient was investigated by compositing ISCCP cloud properties and reanalysis meteorological parameters in a manner similar to that of Lau and Crane (1995, 1997). The compositing parameter used in this study is maximum SST advection, defined as $-\mathbf{V} \cdot \nabla \theta_{SST}$, where $\mathbf{V}$ is 1000-hPa wind and $\nabla \theta_{SST}$ is the gradient of SST. For every July or January during 1984–2001 and every longitude grid point within the region $30^\circ$–$50^\circ$N, $155^\circ$–$215^\circ$E (the smaller box demarcated in Fig. 1), the latitude and day of month with the greatest positive SST advection were identified. The boundaries of the selection region were chosen to emphasize the
midlatitude North Pacific and to avoid influences of land. Values of SST, 1000-hPa wind, and 500-hPa pressure vertical velocity ($\omega_{500}$) at the center of the composite, consistent with strong warm advection over a primarily meridional SST gradient. As expected from quasigeostrophic theory, the area of poleward motion substantially overlaps the area of upward motion at the 500-hPa level. High-top optically thick clouds are located where midtropospheric upward motion is strongest, as previously documented by Lau and Crane (1995) in their composites centered on locations of very optically thick clouds. One new aspect in Fig. 2a, however, is the enhanced frequency of clear-sky southeast of the ascent core during July, where near-surface warm advection occurs beneath midtropospheric descent and weak ascent. Movement of warm subtropical air over colder water likely promotes stratification of the near-surface layer and inhibits the upward flux of moisture needed to sustain cloud growth against subsidence drying. Clear sky becomes less frequent and low-level clouds become more frequent farther north, suggesting that fog and shallow stratus form as the stratified near-surface layer moistens to saturation (Norris 1998a; Norris and Klein 2000). During January, the area of enhanced clear sky occurs on the northern side of the SST gradient (Fig. 2b).

Low-level clouds are prevalent under midtropospheric descent occurring in the cold sector west of the high-top optically thick frontal clouds. These clouds are emphasized by creating composites centered on maximum negative advection over the SST gradient during July and January (Figs. 2c,d). July low-level cold advection clouds generally have greater optical thickness than those in January. The frequency of optically thin clouds and clear sky increases from north to south during July, implying the breakup of stratocumulus clouds as they are advected over warmer water under decreasing midtropospheric subsidence (Fig. 2c). The midlevel clouds occasionally seen on the southern side of the SST gradient during July are probably averaged from narrow bands of higher clouds that occur along cold fronts.

Morphological low cloud types and weather conditions identified in synoptic reports provide qualitative information about the state of the boundary layer not available from the ISCCP data. These surface observations were composited in the same manner as the ISCCP data, except that averaging was carried out over all surface observations within $\pm 12$ h of the composite time and only during 1984–97. The resulting composites are displayed in Fig. 3, and the close similarity between spatial patterns of SST, 1000-hPa wind, and $\omega_{500}$ in Figs. 2a,b display meteorological and cloud conditions associated with maximum positive advection over the SST gradient during July and January. The 1000-hPa wind has a very strong northward component at the center of the composite, consistent with strong warm advection over a primarily meridional SST gradient.
2 and 3 indicate that the difference in averaging time periods has little effect on the composites. Figures 3a,b display frequencies of occurrence of several low cloud types and weather conditions associated with strong warm SST advection during July and January. Not surprisingly, precipitation and nimbostratus are more frequent under the region of strong upward motion, especially during January when ascent is much stronger. During July, small cumulus clouds occur more frequently than usual southeast of the ascent core. This enhanced cumulus presence does not contradict the increased frequency of ISCCP clear sky (Fig. 2a) since the field of view of a surface observer is larger than an ISCCP pixel and the cumuli may be small or widely scattered. Although not shown, surface observers do identify completely clear sky slightly more often in this sector compared to the average over all meteorological conditions. The increased frequency of fog 10° east of the ascent core confirms that the low-level ISCCP clouds detected in that region are indeed advection fog (also documented by Lau and Crane 1997). Fog also frequently occurs underneath frontal clouds in the ascent core during July, but is exaggerated due to basing composites on maximum warm SST advection rather than the actual wind advection.

Fig. 2. Composite spatial distributions of 1000-hPa wind (arrows), SST (nearly horizontal lines), 500-hPa pressure vertical velocity (other solid and dashed lines), and ISCCP cloud anomalies (color) centered on locations within the region 30°–50°N, 155°–215°E where advection of the 1000-hPa wind over the SST gradient is (a) maximum positive during Jul, (b) maximum positive during Jan, (c) maximum negative during Jul, and (d) maximum negative during Jan. Composites were constructed from local noon data during 1984–2001. The SST contour interval is 2°C with a thick line for the 16°C isotherm. The vertical velocity contour interval is 20 hPa day⁻¹ for Jul and 40 hPa day⁻¹ for Jan with negative (upward) contours dashed, positive (downward) contours solid, and no zero contour. Each 2.5° x 2.5° grid box in the plot is filled with 25 pixels, and each pixel represents an additional 2% cloud amount or clear-sky frequency beyond the climatological value for the ISCCP category associated with that color (see legend in figure). Only cloud anomalies statistically significant at 95% are shown, and negative cloud anomalies are not plotted.
than on another parameter such as maximum upward motion.

Low cloud types and weather conditions associated with strong cold SST advection are presented in Figs. 3c,d. Stratus, stratocumulus, and cumulus clouds are very prevalent in the region of strong subsidence and northwesterly flow following the frontal nimbostratus and precipitation. These clouds occur in a generally well-mixed boundary layer capped by a temperature inversion (Norris 1998a). As seen in climatologies (Norris 1998b), stratus and stratocumulus clouds are relatively more prevalent during July, whereas cumulus clouds are relatively more prevalent during January due to the larger negative air–sea temperature difference that generates strong shallow convection. One interesting feature of Fig. 3c is the enhanced frequency of mixed cumulus and stratocumulus followed by small cumulus southwest of the composite center. Mixed cumulus and stratocumulus clouds preferentially occur when the subcloud layer is decoupled from the cloud layer by a thin layer of weak stratification (Norris 1998a). Shallow cumulus convection intermittently penetrates this transition layer and provides the moisture that sustains the overlying stratocumulus deck against entrainment drying (Nicholls 1984). Previous studies investigating cloudiness over eastern subtropical oceans have noted a transition in low cloud type from stratocumulus alone to cumulus under stratocumulus to cumulus alone as the boundary layer is advected by steady trade winds over increasingly warmer water (Wyant et
al. 1997; Norris 1999b). Increased condensational heating resulting from increased latent heat flux promotes decoupling, as does deepening of the boundary layer due to weakening LTS and subsidence. The co-occurrence of equatorward advection over warmer water, a north–south decrease in downward vertical motion, and a north–south transition from mostly stratocumulus to mixed cumulus and stratocumulus seen in Fig. 3c suggests that similar processes occur on synoptic time scales over the midlatitude North Pacific.

4. Dynamical and thermodynamical composites

The spatial distributions of cloud and meteorological variables displayed in Figs. 2 and 3 indicate that the dominant dynamical and thermodynamical parameters associated with cloud properties are vertical velocity, advection, and SST. These parameters are useful proxies for many of the actual processes generating cloud condensate that are not accurately observed or modeled by the reanalysis. Vertical motion is closely tied to the advective tendency of total water mixing ratio, and advection over an SST gradient greatly influences boundary layer turbulent mixing. The dependence of cloud properties on vertical velocity, advection, vertically uniform temperature changes, and vertical stratification changes can be determined by dividing each parameter into multiple intervals and averaging daily cloud values within the intervals. For simplicity, \( \omega_{900} \) and SST advection \((-V \cdot \nabla SST)\) are the only dynamical parameters considered. Spatial composites of meteorological fields (not shown) show that vertical velocity anomalies at 700, 500, and 300 hPa occur at nearly the same horizontal location, suggesting lack of sensitivity to the particular level chosen for vertical velocity. Temperature anomalies exhibit a similarly close vertical alignment, but there is less correspondence between wind fields at different levels due to baroclinicity. Examination of correlations between gridpoint cloud and wind anomalies indicates that, after removing the relationship to 500-hPa vertical velocity, 1000-hPa wind has the largest influence on cloud properties through its interaction with SST gradient. The different effects on cloud properties caused by vertically uniform changes in temperature and changes in vertical stratification are examined by holding LTS constant and letting SST vary in the first case and holding SST constant and letting LTS vary in the second case.

Following the two-variable composite technique of Ringer and Allan (2004), dynamical composites of cloud amount, \( \tau \), and CTP were created by classifying every local noon ISCCP value during July 1984–2001 and within the region 25°–55°N, 145°–225°E (the larger box demarcated in Fig. 1) into 50 hPa day\(^{-1}\) intervals of \( \omega_{900} \) and 2.5°C day\(^{-1}\) intervals of SST advection. ISCCP values during January were similarly classified into 120 hPa day\(^{-1}\) intervals and 4°C day\(^{-1}\) intervals. The total number of available ISCCP values in January or July was 214,272, and each \( \omega \)-advection bi-interval was required to have at least 20 values. ISCCP data in each bi-interval were further classified into terciles of SST anomalies and terciles of LTS anomalies, with middle terciles of SST and LTS corresponding to normal conditions. Use of anomalies was required by the large climatological meridional variation of SST and LTS that would otherwise cause a strong geographical bias in the results. Furthermore, anomalies were standardized separately for each grid box by dividing by the standard deviation prior to classification. This additional step ensures that the upper and lower terciles are not biased toward those areas of the spatial domain where standard deviations of SST or LTS are larger. Were standardization not applied, average cloud amount for both “cold” and “warm” SST terciles could be larger than average cloud amount for the normal tercile merely because climatological cloud amount and SST variability are greater in the northern part of the compositing region. Although some absolute anomalies classified into upper or lower terciles may actually be smaller than other absolute anomalies classified into the middle terciles, standardization does not qualitatively change the differences between average cloud properties for upper and lower terciles.

Figure 4 displays the frequency distributions of \( \omega_{900} \) and SST advection for normal SST and LTS during July and January. Vertical velocity and advection are close to normally distributed with near-zero median values, and the linear correlation between \( \omega_{900} \) and SST advection is \(-0.31\) for July and \(-0.42\) for January. It should be noted that not all parts of the spatial domain contribute equally to each \( \omega \)-advection bi-interval (e.g., large \( \omega \) and advection values preferentially come from the region of the mean storm track and strongest SST gradient).

Figure 5 displays cloud amount, \( \tau \), and CTP for normal SST and LTS conditions averaged into \( \omega \)-advection bi-intervals with weighting by gridbox area. Since radiative flux is nonlinearly related to optical thickness, \( \tau \) values were converted to reflectivity values using an ISCCP lookup table before averaging and then converted back. This procedure ensures that mean values of \( \tau \) correctly represent mean radiative impacts, and were it not applied, domain-averaged \( \tau \) values would be about 30% larger. Differences between two domain-averaged \( \tau \) values would also be 30% larger. Cloud amount in July and January is nearly 100% for all con-
ditions of upward motion and for downward motion when cold advection is strong. Downward motion with warm or near-zero advection has the least cloud amount, corresponding to areas of enhanced frequency of clear sky east of the ascent regions seen in Fig. 2. July $\tau$ is also least in the downward warm advection quadrant (Fig. 5c), but January $\tau$ is equally small for warm and cold advection under conditions of synoptic descent (Fig. 5d). Consistent with the generation of thick clouds with high tops by synoptic ascent, CTP decreases and $\tau$ increases when upward motion becomes stronger (Figs. 5e–f). Note that CTP is likely to be underestimated by 50–80 hPa (cloud top is too high) under conditions of strong downward motion since the Television Infrared Observation Satellite (TIROS) Operational Vertical Sounder (TOVS) temperature profile used by the ISCCP algorithm is inaccurate when a temperature inversion is present (Rossow and Schiffer 1999). Table 1 lists mean cloud properties associated with middle SST and LTS terciles calculated by averaging over the distributions in Fig. 5 with weighting by the frequencies of occurrence in Fig. 4 ($\tau$ and CTP were additionally weighted by cloud amount in each $\omega$–advection bi-interval). Mean values for the months of April and October are included as well.

Figure 6 shows composite SW, LW, and net CRF for July and January obtained from daily ERBE data during 1985–89 and calculated in the same manner as for Fig. 5. Note that fewer data are available since the ERBE time period is only about one-fourth as long as the ISCCP time period. The SW and net CRF are much less negative during January than during July due to the lesser solar zenith angle and length of day. Consistent with Fig. 5, SW CRF is greatest where cloud amount and $\tau$ are greatest and least where they are least (Figs. 6a,b). Similarly, the magnitude of LW CRF inversely varies with CTP (Figs. 6c,d). July net CRF is always negative, and the largest values occur when synoptic ascent generates thick frontal clouds and when synoptic descent and cold advection generate low-level stratocumulus clouds (Figs. 6e and 3c). Contrasting, January net CRF is much weaker and sometimes positive, particularly in the downward warm advection quadrant that has enhanced cirrus frequency with few low clouds (Figs. 6f and 2b). Table 1 lists mean SW, LW, and net CRF associated with middle SST and LTS terciles for the months of July and January.

Effects of a vertically uniform increase in temperature on cloud properties and CRF were estimated by subtracting the average of ISCCP and ERBE values corresponding to the lower SST and middle LTS terciles from the average corresponding to the upper SST and middle LTS terciles. The data were restricted to the middle tercile of LTS in both cases to exclude effects resulting from changes in atmospheric stratification. Thus, the cloud differences result from identical temperature changes at 700 hPa and the surface. Statistical significance was determined by first calculating the difference between averages of two sets of randomly chosen values (with replacement) within each $\omega$–advection bi-interval. The number of values in each random set was the same as the number of values contributing to the upper SST/middle LTS or lower SST/middle LTS terciles. This was repeated 1000 times, and the random differences were placed in ascending order. Composite differences were then deemed statistically significant if
Fig. 5. Composite ISCCP cloud amount (%) for (a) Jul and (b) Jan, composite ISCCP \( r \) for (c) Jul and (d) Jan, and composite ISCCP CTP (hPa) for (e) Jul and (f) Jan during 1984–2001 for \( \omega \)-advection bi-intervals in the region 25°–55°N, 145°–225°E. Averages are calculated only for those values associated with the middle terciles of standardized SST anomalies and standardized LTS anomalies.
they were smaller than the 26th or larger than the 975th random difference.

Figure 7 displays statistically significant differences between ISCCP cloud properties composited on upper (warm) and lower (cold) SST terciles. During July, cloud amount and \( \tau \) are generally lower for warmer conditions (Figs. 7a,c), consistent with findings from previous studies of midlatitude cloud variability over the North Pacific (e.g., Norris and Leovy 1994; Tselloudis et al. 1992; Weare 1994). Less cloud amount and \( \tau \) imply smaller negative SW CRF for warmer temperature, as confirmed by Fig. 8a (note that positive anomalies occur because mean SW CRF is negative). CTP increases and LW CRF decreases with increasing temperature during July (Figs. 7e and 8c). Since the weakening of LW CRF is less than the weakening of SW CRF, July net CRF for the most part becomes less negative (positive anomalies) with warming. Even though changes in January cloud amount and \( \tau \) are small and largely insignificant (Figs. 7b,d), SW and net CRF uniformly become more positive (or less negative) with increasing temperature (Figs. 8b,f). Table 2 lists differences between mean warm and cold cloud properties averaged over all \( \omega \)-advection bi-intervals with weighting by frequency. For purposes of comparability between different months, cloud differences in Table 2 have been divided by the corresponding SST difference. In every month, warmer temperature is associated with a reduction in mean cloud amount, a reduction or no change in mean \( \tau \), a weakening of mean SW CRF, a weakening of mean LW CRF, and a positive shift in mean net CRF.

Effects of increasing LTS were estimated by subtracting the average of lower LTS and middle SST terciles from the average of upper LTS and middle SST terciles. Since the correlation between 700- and 500-hPa temperature is 0.84, variations in LTS largely correspond to opposite changes in SST and free-troposphere temperature. Although figures are not shown for reasons of space, Table 3 lists the differences between mean cloud properties for conditions of strong stratification and weak stratification, normalized by the corresponding LTS difference. In all months, cloud amount and \( \tau \) are slightly larger when stratification is stronger. Examination of strong-minus-weak stratification differences in \( \omega \)-advection composites indicates that the increase in \( \tau \) is greatest for strong ascent and cloud amount is sometimes reduced under downward motion. Mean SW CRF and net CRF are slightly more negative (i.e., stronger) under conditions of enhanced LTS.

One possible confounding factor in the preceding analyses is correlated meridional variations in cloud properties and meteorological parameters. Previous research by Norris (2000) documented the frequent co-occurrence of interannual shifts in the latitude of the SST gradient and in the latitude of the storm track as well as associated nimbostratus clouds. Since northward movement of the location of the SST gradient results in a local warm SST anomaly, a coincidental movement from the south of cloudiness with different characteristics could create an apparent dependence of cloud properties on temperature. This effect is partially mitigated by the incorporation of shifts in the local dynamical distributions by the compositing on \( \omega_{500} \) and SST advection, but Table 2 may nonetheless overestimate the cloud change per degree warming. A poleward shift in storm-track cloudiness would also weaken annual average SW CRF simply because of the lesser insolation at higher latitudes. Unfortunately, it is very difficult to observationally distinguish changes in cloudiness directly due to temperature from meridional shifts in cloudiness indirectly related to temperature. Consequently, the results of this study will be most applicable to the question of the cloud response to global warming if the storm track and SST gradient also move poleward.

5. Cloud response to dynamical changes

Aside from variability in thermodynamical parameters, average cloud properties can change merely in response to altered frequency distributions of \( \omega_{500} \) and SST advection. This effect can be explored by holding constant the normal SST/LTS composite cloud properties displayed in Figs. 5 and 6 while averaging them with weightings different than those displayed in Fig. 4. The

### Table 1. Average properties for middle SST and LTS terciles.*

<table>
<thead>
<tr>
<th>Month</th>
<th>Cloud amount (%)</th>
<th>( \tau )</th>
<th>CTP (hPa)</th>
<th>SW CRF (W m(^{-2}))</th>
<th>LW CRF (W m(^{-2}))</th>
<th>Net CRF (W m(^{-2}))</th>
<th>SST (°C)</th>
<th>LTS (K)</th>
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<td>-96</td>
<td>28</td>
<td>-68</td>
<td>18.9</td>
<td>13.1</td>
</tr>
</tbody>
</table>

The type of dynamical change considered is amplification and reduction of the standard deviations of the $\omega_{500}$ and SST advection distributions by an increment equal to one-tenth of the original standard deviation. One limitation of this approach is the arbitrary specification of the magnitudes of dynamical changes. Since it has not yet been established that a $+0.2 \sigma$ shift in the standard deviation of $\omega_{500}$ and SST advection is as equally likely as a $+1$ K shift in temperature or LTS, cloud responses to changes in these parameters cannot be directly com-

Fig. 6. Same as in Fig. 5, except for ERBE SW CRF for (a) Jul and (b) Jan, LW CRF for (c) Jul and (d) Jan, and net CRF for (e) Jul and (f) Jan during 1985–89.
pared. This shortcoming was surmounted by assessing the relative sizes of parameter shifts through calculation of regional average interannual standard deviations of monthly mean temperature and LTS and interannual standard deviations of the intramonthly standard deviation of $\omega_{500}$ and SST advection. The resulting values, recorded in Table 4, indicate that typical interannual anomalies are indeed on the order of 1 K and 0.2 $\sigma$.

Fig. 7. Same as in Fig. 5, except for the average of ISCCP values associated with the lower (cold) SST and middle LTS terciles subtracted from the average of ISCCP values associated with the upper (warm) SST and middle LTS terciles. Only differences significant at the 95% level are shown.
The first step to applying the dynamical transformation was classification of the $\omega_{500}$ values into ascending percentiles ($i = 0–99$) followed by classification of corresponding SST advection values within each $\omega_{500}$ percentile into their own ascending percentiles ($j = 0–99$). Since $\omega_{500}$ and SST advection are correlated, use of a two-stage process prevented modifications to the $\omega_{500}$ distribution from inadvertently introducing modifica-
Average difference between middle SST/middle LTS (strong stratification) and middle SST/lower LTS (weak stratification)

<table>
<thead>
<tr>
<th>Month</th>
<th>Cloud amount (K)</th>
<th>τ (K)</th>
<th>CTP (hPa K⁻¹)</th>
<th>SW CRF (W m⁻² K⁻¹)</th>
<th>LW CRF (W m⁻² K⁻¹)</th>
<th>Net CRF (W m⁻² K⁻¹)</th>
<th>SST anomaly (K)</th>
<th>LTS anomaly (K)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Jan</td>
<td>0.1</td>
<td>0.0</td>
<td>-6</td>
<td>+4.4</td>
<td>-1.0</td>
<td>+3.4</td>
<td>+1.8</td>
<td>+0.1</td>
</tr>
<tr>
<td>Apr</td>
<td>-1.1</td>
<td>-0.1</td>
<td>-6</td>
<td>+4.4</td>
<td>-1.0</td>
<td>+3.4</td>
<td>+1.8</td>
<td>+0.1</td>
</tr>
<tr>
<td>Jul</td>
<td>-2.6</td>
<td>-0.3</td>
<td>+6</td>
<td>+9.4</td>
<td>-2.5</td>
<td>+6.9</td>
<td>+2.5</td>
<td>-0.2</td>
</tr>
<tr>
<td>Oct</td>
<td>-1.6</td>
<td>0.0</td>
<td>-2</td>
<td>0</td>
<td>-2.5</td>
<td>+6.9</td>
<td>+2.5</td>
<td>0.0</td>
</tr>
</tbody>
</table>

* Same as in Table 1. Cloud properties have been divided by the SST difference, and bold indicates differences that are significant at the 95% level.

Table 3. Average difference between middle SST/upper LTS (strong stratification) and middle SST/lower LTS (weak stratification) terciles.

<table>
<thead>
<tr>
<th>Month</th>
<th>Cloud amount (K)</th>
<th>τ (K)</th>
<th>CTP (hPa K⁻¹)</th>
<th>SW CRF (W m⁻² K⁻¹)</th>
<th>LW CRF (W m⁻² K⁻¹)</th>
<th>Net CRF (W m⁻² K⁻¹)</th>
<th>SST anomaly (K)</th>
<th>LTS anomaly (K)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Jan</td>
<td>+0.2</td>
<td>+0.1</td>
<td>-4</td>
<td>-0.1</td>
<td>+0.1</td>
<td>-0.1</td>
<td>0.0</td>
<td>+8.9</td>
</tr>
<tr>
<td>Apr</td>
<td>+0.5</td>
<td>+0.2</td>
<td>-4</td>
<td>0</td>
<td>-0.2</td>
<td>-1.2</td>
<td>0.0</td>
<td>+9.3</td>
</tr>
<tr>
<td>Jul</td>
<td>+0.3</td>
<td>+0.1</td>
<td>-4</td>
<td>-1.0</td>
<td>-0.2</td>
<td>-1.2</td>
<td>0.0</td>
<td>+5.6</td>
</tr>
<tr>
<td>Oct</td>
<td>-0.2</td>
<td>+0.2</td>
<td>0</td>
<td>0</td>
<td>0.0</td>
<td>0.0</td>
<td>0.0</td>
<td>+7.4</td>
</tr>
</tbody>
</table>

* Same as in Table 1. Cloud properties have been divided by the LTS difference, and bold indicates differences that are significant at the 95% level.
the combined changes in \( \omega \) and advection variance is nearly the same as the sum of the individual effects listed in Table 5. Although previous studies have separately examined trends in the frequency and trends in the intensity of extratropical cyclones (e.g., McCabe et al. 2001), the present analysis can only address changes in the strength of storms since the \( \omega \)-advection distributions do not contain information on the number of cyclones that contribute to them.

Of course, it is possible that the relative magnitudes of long-term trends will differ from those of interannual anomalies. The GCM study of Dai et al. (2001) predicts that between the 1990s and the 2090s central North Pacific temperature will increase by about 2 K, the standard deviation of 2–8-day SLP variance will decrease by about 10% (\(-0.1 \sigma\)), and negligible changes will occur in LTS. These trends suggest that vertically uniform warming will produce the largest change in cloud properties: less cloud amount, less \( \tau \), and less negative (or more positive) net CRF. The weakening of the storm track will produce similar but smaller cloud changes. Somewhat contrasting results are found in the GCM study of Carnell and Senior (1998), who documented a 2-K rise in northern midlatitude temperature and a tendency for fewer but stronger cyclones due to greater latent heating. In this case, the expected cloud response to the apparent intensification of the storm track will partially cancel the expected cloud response to increased temperature. Assuming that the relationships documented in this study are also influential under large-scale climate change, the large sensitivity of cloud properties to temperature (Table 2) suggests that the overall response will be decreases in cloud amount, \( \tau \), and the magnitudes of SW and LW CRF. Net CRF will become more positive, implying a positive cloud feedback on temperature. Trends in GCM cloudiness are not discussed here because Norris and Weaver (2001) found substantial deficiencies in their simulation over the midlatitude North Pacific at synoptic and interannual time scales.

### 6. Discussion and conclusions

The differences in cloud properties associated with changes in thermodynamic and dynamic parameters are consistent with the synoptic cloud relationships illustrated by the spatial composites. Stratocumulus cloudiness breaks up when it is advected equatorward over warmer subtropical water because the subcloud layer becomes decoupled from the cloud layer, thus inhibiting the upward flux of moisture needed to sustain the cloud deck. Due to the increase in saturation vapor pressure, higher temperature increases the amount of latent heat released in cloud condensation and thereby enhances the increase of stratification between subcloud and cloud layers (i.e., decoupling; Wyant et al. 1997). A warm SST anomaly additionally acts to shift the main SST gradient poleward such that cloud breakup occurs sooner. The reduction of upward moisture flux also contributes to a thinning of the cloud layer. For poleward advection, higher temperature de-

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**Table 4. Interannual standard deviations of monthly means and intramonthly standard deviations.**

<table>
<thead>
<tr>
<th>Month</th>
<th>SST (K)</th>
<th>LTS (K)</th>
<th>( \omega_{500} ) intramonthly std dev (( \sigma ))</th>
<th>SST advection intramonthly std dev (( \sigma ))</th>
</tr>
</thead>
<tbody>
<tr>
<td>Jan</td>
<td>0.7</td>
<td>1.7</td>
<td>0.19</td>
<td>0.21</td>
</tr>
<tr>
<td>Apr</td>
<td>0.6</td>
<td>1.7</td>
<td>0.19</td>
<td>0.22</td>
</tr>
<tr>
<td>Jul</td>
<td>0.8</td>
<td>1.3</td>
<td>0.19</td>
<td>0.21</td>
</tr>
<tr>
<td>Oct</td>
<td>0.8</td>
<td>1.3</td>
<td>0.18</td>
<td>0.22</td>
</tr>
</tbody>
</table>

* Same as in Table 1. Dynamical parameters have units of \( \sigma \), the std dev of \( \omega_{500} \) or SST advection distributions.

**Table 5. Difference between a \(+0.1 \sigma\) and a \(-0.1 \sigma\) change in the standard deviation of the (top rows) \( \omega_{500} \) or (bottom rows) SST advection distributions for middle SST and LTS terciles.**

<table>
<thead>
<tr>
<th>Month</th>
<th>Amount (%)</th>
<th>( \tau ) (hPa)</th>
<th>CTP (hPa)</th>
<th>SW CRF (W m(^{-2}))</th>
<th>LW CRF (W m(^{-2}))</th>
<th>Net CRF (W m(^{-2}))</th>
<th>( \omega_{500} ) ( \Delta \sigma ) (hPa day(^{-1}))</th>
<th>SST advection ( \Delta \sigma ) (°C day(^{-1}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>Jan</td>
<td>+0.2</td>
<td>+0.3</td>
<td>-7</td>
<td>-0.7</td>
<td>+0.8</td>
<td>+0.1</td>
<td>+39</td>
<td>+1.1</td>
</tr>
<tr>
<td>Apr</td>
<td>+0.6</td>
<td>+0.5</td>
<td>-7</td>
<td>-0.7</td>
<td>+0.8</td>
<td>+0.1</td>
<td>+31</td>
<td>+0.9</td>
</tr>
<tr>
<td>Jul</td>
<td>+0.8</td>
<td>+0.2</td>
<td>-9</td>
<td>-3.6</td>
<td>+1.2</td>
<td>-2.3</td>
<td>+16</td>
<td>+0.9</td>
</tr>
<tr>
<td>Oct</td>
<td>+1.2</td>
<td>+0.3</td>
<td>-7</td>
<td>-3.6</td>
<td>+1.2</td>
<td>-2.3</td>
<td>+32</td>
<td>+1.0</td>
</tr>
<tr>
<td>Jan</td>
<td>+0.6</td>
<td>+0.1</td>
<td>-2</td>
<td>-0.4</td>
<td>+0.3</td>
<td>+0.1</td>
<td>+1.1</td>
<td>+0.9</td>
</tr>
<tr>
<td>Apr</td>
<td>+0.4</td>
<td>0.0</td>
<td>-3</td>
<td>-0.4</td>
<td>+0.3</td>
<td>+0.1</td>
<td>+1.0</td>
<td>+0.9</td>
</tr>
<tr>
<td>Jul</td>
<td>-0.9</td>
<td>+0.2</td>
<td>0</td>
<td>-1.2</td>
<td>+0.3</td>
<td>-1.0</td>
<td>+1.0</td>
<td>+1.0</td>
</tr>
<tr>
<td>Oct</td>
<td>+1.0</td>
<td>+0.1</td>
<td>-1</td>
<td>-1.2</td>
<td>+0.3</td>
<td>-1.0</td>
<td>+1.0</td>
<td>+1.0</td>
</tr>
</tbody>
</table>

* Same as in Table 1. Here \( \sigma \) is the original std dev of \( \omega_{500} \) or SST advection distributions, and \( \Delta \sigma \) is the difference between the modified std dev.
lays the onset of fog formation and therefore maintains
the regime of a smaller amount of optically thinner
clouds. Cloud amount, optical thickness, and the mag-
nitude of SW CRF all decrease with rising temperature
under most conditions of vertical velocity and SST ad-
vection. More research is needed to determine why this
is so. Although the magnitude of LW CRF also de-
creases, the change in SW CRF is larger, so net CRF
becomes less negative (or more positive) with warmer
temperature. Stronger stratification in the lower tropo-
sphere (or a smaller lapse rate) is associated with more
cloud amount and optical thickness.

A change in the variance of the $o_{500}$ distribution can
also alter mean cloud properties as a consequence of
their nonlinear dependence on vertical velocity. Since
cloud optical thickness and cloud-top height increase
more strongly with upward motion than they decrease
with downward motion, more frequently occurring
strong ascent and descent will produce on average
greater optical thickness and top height. The impact of
greater optical thickness on SW CRF is larger than the
impact of cloud-top height on LW CRF, so net CRF
becomes more negative when $o_{500}$ variance becomes
larger. Opposite changes result from a weakening of
$o_{500}$ variance.

The previously documented synoptic cloud responses
to changes in vertically uniform temperature, stratifica-
tion, $o_{500}$, and SST advection can be used to infer the
response of midlatitude oceanic clouds to climate
change. Presuming that concentrations of greenhouse
gases in the atmosphere continue to increase, it is very
likely that both the atmosphere and the ocean will
warm. If cloud properties respond to global warming in
the same way that they do to synoptic variations in
temperature, then the results of this study suggest that
cloud amount and optical thickness will decrease over
the North Pacific Ocean at all times of year and allow
more solar radiation to be absorbed by the earth. The
implied consequence increase in outgoing LW radia-
tion due to weaker LW CRF only partially compensates the SW
gain; consequently, clouds in this region and season act
as a positive feedback on the climate system. Clouds
may also respond indirectly to global warming through
changes in atmospheric dynamics. For example, polar
amplification of global warming implies less baroclinic-
ity and an associated weakening of the storm track,
which, according to calculations in this study, will result
in smaller mean optical thickness and mean cloud-top
height. The radiative effects of these changes are an
increase in absorbed SW and a smaller decrease in out-
going LW, again contributing to a positive cloud feed-
back. Alternatively, global warming implies more water
vapor and latent heating, which may act to intensify
cyclones, produce more cloud amount and optical
thickness, and thus partially cancel the cloud response
to warmer temperature.

To better understand the processes governing the
cloud response to changes in temperature and other
parameters over the North Pacific, as well as extension
of the analysis to other regions of the world, more re-
search is needed. The thermodynamical and dynamical
connections to cloudiness documented in this study
provide strong constraints for the validation of GCM
cloud simulations, and it would be particularly useful to
evaluate leading models by how well they reproduce
the observed synoptic relationships. This would help
assess the reliability of cloud processes in GCMs, many
of which may be important for future climate change
cloud feedbacks.

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pheric Sciences Data Center (see online at coswes-
balarc.nasa.gov), reanalysis data from the Web site of
the NOAA–CIRES Climate Diagnostics Center (see
online at www.cdc.noaa.gov), and SST data from the
International Research Institute/Lamont Doherty
Earth Observatory climate data library Web site (see
online at http://iridl.ldeo.columbia.edu). The authors
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