Subsidence Warming as an Underappreciated Ingredient in Tropical Cyclogenesis. Part I: Aircraft Observations

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ABSTRACT

The development of a compact warm core extending from the mid-upper levels to the lower troposphere and related surface pressure falls leading to tropical cyclogenesis (TC genesis) is not well understood. This study documents the evolution of the three-dimensional thermal structure during the early developing stages of Typhoons Fanapi and Megi using aircraft dropsonde observations from the Impact of Typhoons on the Ocean in the Pacific (ITOP) field campaign in 2010. Prior to TC genesis, the precursor disturbances were characterized by warm (cool) anomalies above (below) the melting level (~550 hPa) with small surface pressure perturbations. Onion-shaped skew $T$–log\(p\) profiles, which are a known signature of mesoscale subsidence warming induced by organized mesoscale convective systems (MCSs), are ubiquitous throughout the ITOP aircraft missions from the precursor disturbance to the tropical storm stages. The warming partially erodes the lower-troposphere (850–600 hPa) cool anomalies. This warming results in increased surface pressure falls when superposed with the upper-troposphere warm anomalies associated with the long-lasting MCSs/cloud clusters. Hydrostatic pressure analysis suggests the upper-level warming alone would not result in the initial sea level pressure drop associated with the transformation from a disturbance to a TC. As Fanapi and Megi intensify into strong tropical storms, aircraft flight-level (700 hPa) and dropsonde data reveal that the warm core extends down to 850–600 hPa and has some characteristics of subsidence warming similar to the eyes of mature TCs.

1. Introduction

One of the defining characteristics of a tropical cyclone (TC) is its warm core. The development of a deep-troposphere warm temperature anomaly and low-level cyclonic circulation leading to tropical cyclogenesis (TC genesis) has been the subject of many previous studies. Observational and modeling studies have shown that organized mesoscale convective systems (MCSs) can lead to the development of a mid-upper-troposphere (~200–500 hPa) warm anomaly and a midlevel (~500–600 hPa) vorticity maximum (Chen and Frank 1993; Fritsch et al. 1994). However, there is less agreement regarding the factors that contribute to the development of warming in the lower troposphere (~600–1000 hPa).

Large-scale environment conditions that are conducive to TC development include (but are not limited to) low vertical wind shear of the horizontal wind, enhanced cyclonic circulation and moisture, sufficient Coriolis force, and a warm ocean (Gray 1968; McBride and Zehr 1981; DeMaria et al. 2001). These conditions are often associated with the development of numerous long-lasting MCSs, especially in the presence of preexisting tropical disturbances like easterly waves and monsoon troughs (Gray 1998). Nearly all TC genesis cases over the western North Pacific are associated with mesoscale cloud clusters (with cloud-top temperature < 208 K) lasting at least 8 h or longer (e.g., 8-h clusters; Kerns and Chen 2013). These long-lasting clusters represent large, organized MCSs with extensive regions of stratiform precipitation (Chen et al. 1996). Observations of these MCSs have shown enhanced midlevel (~500–600 hPa) vorticity within the stratiform region in midlatitudes during summertime (Brandes 1990) and in the tropics (Bousquet and Chong 2000).

Numerical modeling studies have provided some explanations for the development of the midlevel vortex and mid-upper-troposphere warming within MCSs. Chen and Frank (1993) have shown that the mesoscale
cyclogenesis is largely due to warming in the saturated stratiform region of MCSs enhanced by inertial stability (Schubert and Hack 1982). Others have shown that the lower-troposphere cooling is related to evaporative cooling of precipitation (Mapes and Houze 1995; Bister and Emanuel 1997; Schumacher et al. 2004). Based on a numerical modeling experiment, Bister and Emanuel (1997) proposed that an intense, long-lived, mesoscale “showerhead” can produce a low-level cold core prior to TC genesis. They proposed that the low-level cooling could lead to enhanced air–sea fluxes that can, in turn, eliminate the low-level cool anomaly from below, which would favor subsequent deep convection. However, there have been few observations during the early TC life cycle to evaluate this hypothesis so far.

Similar to the temperature anomalies within MCSs, the upper-troposphere warm and lower-troposphere cool anomalies have been observed in many tropical disturbances (Reed and Recker 1971; Reed et al. 1977; McBride 1981; Bessho et al. 2010; Komaromi 2013; Zawislak and Zipser 2014). To some extent, this observed thermodynamic structure reflects the collective contribution from latent heating and evaporative cooling in multiple long-lasting MCSs within the large-scale disturbances (Ritchie and Holland 1997; Simpson et al. 1997).

The question of how a deep warm core develops during TC genesis remains unanswered. Observational and numerical modeling studies of tropical and mid-latitude MCSs provide some possible explanations for lower-troposphere warming. Chen and Frank (1993) described the development of a wake low associated with lower-troposphere subsidence warming on the edge of a saturated region where a downward development of a mesovortex with a deep warm core occurred in a numerical simulation of mesoscale cyclogenesis. Wake lows were also observed previously in MCSs (Zipser 1977; Johnson and Nicholls 1983; Johnson and Hamilton 1988; Johnson et al. 1989). A schematic based on the model simulation from Chen and Frank (1993) is shown in Fig. 1. In organized MCSs, subsidence warming is often related to mesoscale inflow jets and mesoscale subsidence regions beneath the melting level, both of

FIG. 1. A schematic depiction of a mature midlatitude MCS, following Chen and Frank (1993). The mid-upper-level warm anomaly (black W) and vorticity center (black V), as well as the lower-troposphere adiabatic subsidence warming (red W) are indicated. The skew $T$–$\log p$ diagram in Fig. 13 of Chen and Frank is overlaid with the rear inflow jet, which shows the onion thermodynamic profile in the subsiding region.

### TABLE 1. Summary of the aircraft missions and dropsondes used in this study. Storm intensity is indicated as precursor disturbance (DB), TD, or TS.

<table>
<thead>
<tr>
<th>Storm/mission</th>
<th>Storm intensity</th>
<th>Flight level (hPa)</th>
<th>First–last dropsonde</th>
<th>Sondes</th>
</tr>
</thead>
<tbody>
<tr>
<td>Fanapi/F1</td>
<td>DB (15 kt)</td>
<td>350</td>
<td>1916 UTC 12 Sep–0253 UTC 13 Sep</td>
<td>31</td>
</tr>
<tr>
<td>Fanapi/F2</td>
<td>DB/TD (20–25 kt)</td>
<td>350</td>
<td>2137 UTC 13 Sep–0240 UTC 14 Sep</td>
<td>21</td>
</tr>
<tr>
<td>Fanapi/F3</td>
<td>TD/TS (30–35 kt)</td>
<td>350</td>
<td>2055 UTC 14 Sep–0146 UTC 15 Sep</td>
<td>26</td>
</tr>
<tr>
<td>Fanapi/F4</td>
<td>TS (55 kt)</td>
<td>700</td>
<td>2234 UTC 15 Sep–0353 UTC 16 Sep</td>
<td>26</td>
</tr>
<tr>
<td>Megi/M1</td>
<td>TD/TS (30–35 kt)</td>
<td>350</td>
<td>1840 UTC 12 Oct–0052 UTC 13 Oct</td>
<td>30</td>
</tr>
<tr>
<td>Megi/M2</td>
<td>TS (55 kt)</td>
<td>700</td>
<td>0006–0452 UTC 14 Oct</td>
<td>27</td>
</tr>
</tbody>
</table>
which result in characteristic onion-shaped profiles on thermodynamic diagrams (Zipser 1977; Smull and Houze 1987; Houze 1977; Leary 1984; Johnson and Hamilton 1988; Johnson et al. 1989), which are also illustrated in the inset on the left side of Fig. 1. Rear inflow jets are often found on the trailing/rear edges of strong, mature squall-line systems (Bousquet and Chong 2000; Smull and Houze 1987). However, such strong squall lines are relatively less common in the tropics.

Subsidence warming can still occur in MCSs without strong inflow jets. Another commonly observed feature is an unsaturated mesoscale downdraft beneath the melting layer in the stratiform region of MCSs. As hydrometeors fall from the saturated anvil cloud into subsaturated air, they partially evaporate. Initially, this evaporative cooling makes the air parcel negatively buoyant. However, in light precipitation, the evaporative cooling cannot keep the air parcel saturated and, as a result, it warms during the descent because of adiabatic compression. The adiabatic compression overcomes the evaporative cooling, resulting in net warming and drying beneath the anvil cloud—the distinctive onion-shaped profile on thermodynamic diagrams (Zipser 1977; Leary 1980; Jorgensen et al. 1991; Brandes and Ziegler 1993; Correia and Arritt 2008). Aircraft data from GATE show that evaporative cooling of rainfall

![Fig. 2. Storm tracks and dropsonde locations for (a) Typhoon Fanapi and (b) Typhoon Megi. The JTWC best track is indicated by the black line with circles for every 6 h. Best track starts at 1200 UTC 14 Sep 2010 for Fanapi and 1800 UTC 12 Oct for Megi. JTWC invest locations are indicated by I every 6 h, starting at 0600 UTC 13 Sep for Fanapi and 1200 UTC 11 Oct for Megi. The location of Guam is indicated by G.](http://journals.ametsoc.org/doi/pdf/10.1175/JAS-D-14-0366.1)

![Fig. 3. Infrared satellite image (color shading) with 8-h cloud cluster tracks for (a) Fanapi and (b) Megi. (a) IR image is for 0200 UTC 14 Sep during F2, and the cluster tracks are from 1100 UTC 13 Sep to 0000 UTC 17 Sep. In (b), the IR image is for 0200 UTC 13 Oct during M1, and the cluster tracks are from 1000 UTC 11 Oct to 0400 UTC 15 Oct. White circles are hourly cloud cluster locations, and the circle sizes are proportional to the 208-K IR area. Cloud cluster centroid tracks are drawn in magenta. Magenta dots are the initial locations of the 8-h clusters.](http://journals.ametsoc.org/doi/pdf/10.1175/JAS-D-14-0366.1)
and melting in the melting layer induce a nearly uniform area of descent in stratiform rain areas (Leary and Houze 1979; Leary 1980). Using a numerical model, Brown (1979) determines that this evaporative cooling could induce \( \sim 10 \text{ cm s}^{-1} \) of descent in the stratiform rain area, similar to the observation by Gamache and Houze (1982). In the Bow Echo and Mesoscale Convective Vortex Experiment (BAMEX) descent of 10–50 cm s\(^{-1}\) (several meters per second in strong cases) is found in the trailing stratiform region (Correia and Arritt 2008). Other observational studies have confirmed the predominance of mesoscale unsaturated downdrafts in the stratiform rain areas of tropical (Jorgensen et al. 1991) and midlatitude summer (Brandes 1990; Brandes and Ziegler 1993; Biggerstaff and Houze 1991) MCSs.

Unsaturated mesoscale downdrafts, inflow jets, and related subsidence warming in MCSs have been well documented, but their contribution to TC genesis has not been investigated. While subsidence warming is a well-known feature of mature TCs (Malkus 1958; La Seur and Hawkins 1963; Halverson et al. 2006), how early in the TC life cycle these features develop is not well known. Stossmeister and Barnes (1992) showed that, in Tropical Storm Isabel (1985), a new circulation center developed in a relatively warm, dry area beneath the downwind cloud anvil outside of the radius of maximum winds and separate from the strongest convection. They hypothesized the role of subsidence warming similar to mesolows. Simpson et al. (1998) find that, for Hurricane Daisy (1958) in the Atlantic and
Typhoon Oliver (1993) in the Coral Sea, partial eyewalls developed near the subsidence region adjacent to the large MCSs near the center. Heymsfield et al. (2006) point out that mesoscale subsidence was present at 700–500 hPa over the exposed circulation center of highly sheared Atlantic Tropical Storm Chantal (2001). More recently, Dolling and Barnes (2012) describe the warm-core development associated with the transition of Tropical Storm Humberto (2001) from a tropical depression to a tropical storm. The formation of the warm core is due in part to subsidence warming beneath a trailing anvil at the periphery of an area of persistent deep convection. The pronounced subsidence warming in that case (with aircraft-measured descent of $3 \text{ m s}^{-1}$) may have been related to a dynamically forced system, such as a descending inflow jet similar to that shown in Fig. 1. These studies suggest that subsidence warming plays a role in the tropical storm stage, but they do not focus on the precursor disturbance and tropical depression stages. Nevertheless, recent dropsonde data analysis shows that there is often a relative minimum in relative humidity in the low levels within the large-scale moist disturbance envelope leading up to TC genesis (Komaromi 2013; Davis and Ahijevych 2013; Zawislak and Zipser 2014), which is consistent with subsidence-induced warming in the lower troposphere.

In this study, we use airborne observations collected in two developing TCs during the Impact of Typhoons on Ocean in the Pacific (ITOP) field campaign in 2010 (D’Asaro et al. 2014) to further examine the evolution of temperature, sea level pressure (SLP), and low-level circulation. Here, TC genesis is defined as the initial formation of a tropical depression in the Joint Typhoon Warning Center (JTWC) best-track data. The focus will be on one underappreciated aspect of organized deep convection during TC genesis—the thermodynamic role of mesoscale subsidence warming induced by multiple long-lived MCSs. The organization of this paper is as follows. Section 2 describes the data and methods of analysis. Section 3 gives an overview of two ITOP
storms: Typhoon Fanapi and Typhoon Megi. Section 4 presents the three-dimensional evolution of winds, moisture, and temperature at the early stages of these two typhoons. These results are summarized in terms of the contribution to sea level pressure falls in section 5. Finally, conclusions are presented in section 6.

2. Data and methods

a. Aircraft data

During the ITOP field campaign from August to October 2010, U.S. Air Force C-130 aircraft were deployed into the tropical disturbances prior to TC genesis and early storm stages of Typhoons Fanapi and Megi from their base on Guam (Table 1, Fig. 2). Fanapi and Megi were also subsequently sampled as tropical storms and mature typhoons. Several of these missions were conducted at 350 hPa (Table 1), allowing the GPS dropsondes to sample a deep layer of the troposphere, as described by Elsberry and Harr (2008), which is generally not available in routine operational missions into pre-TC disturbances. Typhoon Fanapi was the most thoroughly sampled system, with a series of four daily missions from when it was a pre-TC disturbance to when it was an intensifying tropical storm (Table 1, Fig. 2a). The first three missions, referred to as F1, F2, and F3, respectively, were square-spiral surveillance missions during the pregenesis and early tropical storm stages of Fanapi. Dropsondes were generally deployed every 60–75 nm (110–140 km) in the square-spiral missions. F4 was conducted when Fanapi was a tropical storm. For F4, an alpha pattern at 700 hPa was conducted. Typhoon Megi was sampled as a minimal tropical storm (M1) and
again the next day (M2) as an intensifying tropical storm with maximum sustained winds of 55 knots (kt; 1 kt = 0.51 m s$^{-1}$) (Table 1, Fig. 2b). For M2, a hybrid square spiral–butterfly pattern was conducted from 700 hPa. For the F4 and M2 missions conducted at 700 hPa, the flight-level data provide fine horizontal-resolution data (1 Hz) within the developing warm core in the lower troposphere. The aircraft flight-level temperature (from the Rosemount sensor) and dewpoint data were corrected for sensor wetting using the method of Zipser et al. (1981) and Eastin et al. (2002).

b. Radiosonde data

During ITOP, radiosondes were released from Guam (PGAC: 13.5°N, 144.8°E; marked “G” in Fig. 2), every 12 h as part of the routine National Weather Service observations. The Guam soundings provide upstream environmental data as a context for the disturbed weather sampled by the dropsondes. The Guam mean temperature during August–October 2010 is used as a reference profile to help identify the warm and cool anomalies associated with the areas of disturbed weather. Additionally, 12-hourly soundings were available from the following stations in the western North Pacific: Yap (PTY: 9.5°N, 138.1°E), Ponape (PTP: 7.0°N, 158.2°E), Koror (PTR: 7.3°N, 134.5°E), Chuuk (PTK: 7.4°N, 151.8°E), and Majuro (PMK: 7.1°N, 171.3°E). The Guam ITOP mean profile was similar (within 0.5°C) to the grand mean profile of all of these stations during the ITOP period (not shown). The Guam mean profile is considered to be a representative reference profile for the study region during ITOP.
The Guam mean profile mainly reflects undisturbed weather conditions that are similar to those in the environment surrounding the disturbances sampled during ITOP. Throughout this study, temperature anomalies are all relative to the August–October 2010 Guam mean profile. Note that this reference profile is mainly to facilitate the intercomparison of the various missions, not to identify the distinctive onion-shaped profiles.

c. Satellite data and cloud cluster tracking

Hourly infrared (IR) temperature data are used to track cloud clusters based on the method used in Williams and Houze (1987) and Chen et al. (1996). The satellite data are from the IR channel (10.5–11.5 μm) of the geostationary Multifunctional Transport Satellites (MTSAT). The data are available at 0.05° resolution. Cloud clusters are identified as contiguous features with IR brightness temperatures below 208 K. They are tracked forward and backward in time whenever they overlap by at least 50% or 5000 km² between consecutive images. The center of the cloud cluster is defined as the geometric centroid of the 208-K area. Long-lived cloud clusters are generally associated with large areas of stratiform precipitation (Chen et al. 1996). See Kerns and Chen (2013) for more details regarding the application of the cloud cluster tracking for TC genesis.

d. Wind analysis

Horizontal pressure level analyses of winds and temperature were constructed using the dropsonde data. The aircraft sampled the system over the course of several hours (Table 1), during which the storm was moving. Additionally, the sondes traveled some horizontal distance between the time they were launched and the time they hit the sea. Therefore, when using the sondes to construct a composite wind analysis at the nominal analysis time (0000 UTC), a correction was made for the storm motion. The storm motion correction was determined as follows. The JTWC best-track data from 6 h prior and 6 h after the nominal
analysis time (e.g., 1800 and 0600 UTC coordinates for the nominal analysis time at 0000 UTC) were used to determine the average storm motion during the flight. For the pregenesis stage, the “invest” center locations were used. For F1, invest coordinates were not available until 0600 UTC, and the best-track coordinates from 0600 and 1200 UTC were used for the storm motion correction to the 0000 UTC analysis time. To implement the storm motion correction, the dropsonde locations were shifted by the displacement the storm would have moved through during the time between the dropsonde time and the nominal time. The storm motion correction was applied to each level independently.

To show the characteristics of the system circulation, layer-mean winds for 840–860, 690–710, and 490–510 hPa were calculated. The wind observations within each layer were weighted by the logarithm of pressure. If there were no dropsonde data within the 20-hPa thick layer, the layer-mean winds were considered to be missing. F4 and M2 did not have temperature and wind data above 700 hPa.

e. Temperature and geopotential height analysis

For the pressure level temperature and height analyses, a two-dimensional objective analysis was conducted at each pressure level independently using two-dimensional natural neighbor interpolation with Delaunay triangulation (Sibson 1981). In this approach, triangles are constructed using the dropsonde locations, and the points within each triangle are weighted averages based on the overlap of the grid cells and the triangles. The dropsonde temperature and altitude data were first interpolated to 10-hPa resolution using linear interpolation in the logarithm of pressure. For studying the warm-core development of incipient TCs, in which deviations of 1°–2°C are significant, temperature anomalies can be easier to interpret than temperature profiles themselves. As discussed above, the Guam mean profile is used to calculate the vertical profiles of
temperature anomalies. The estimated error of the temperature sensors on the GPS dropsondes is 0.2°C (Hock and Franklin 1999), and the features of interest deviate 0.5°C–2.0°C from the reference profile.

f. Hydrostatic surface pressure falls

To quantify the potential contribution of system warming to the SLP falls as a function of height, an analysis of the contribution of temperature deficits at all altitudes was performed, using hydrostatic balance. This analysis is similar to that of Dolling and Barnes (2014). The hydrostatic SLP difference between two profiles is

$$\Delta SLP = SLP_{\text{dropsonde}} - SLP_{\text{ref}}$$

$$= \int_{\text{TOA}}^{\text{TOA}} -g[\rho_{\text{dropsonde}}(z) - \rho_{\text{ref}}(z)] \, dz,$$

where TOA is the top of the atmosphere, $g$ is the acceleration due to gravity, $\rho$ is the density, and $z$ is the altitude. The subscripts “dropsonde” and “ref” refer to the individual dropsonde and the Guam mean, respectively. The term inside the integral represents the contribution of each altitude’s virtual temperature deviation to the SLP drop relative to the Guam mean reference sounding. Density was calculated from temperature and pressure using the ideal gas law. Dropsonde data were not available above ~350 hPa, so the full upper-level contribution to the SLP falls could not be determined. For these calculations, the dropsonde data were first interpolated to 50-m resolution. The interpolation was linear in height.

3. Overview of Typhoons Fanapi and Megi

Typhoon Fanapi formed at 1200 UTC 14 September 2010 about 900 km east of the Philippines (Fig. 2a).
Fanapi intensified to 105 kt before making landfall in Taiwan. The storm was the most thoroughly sampled storm during ITOP with a series of seven C-130 missions. Several long-lived (e.g., 8 h) clusters could be tracked as the precursor disturbance moved westward (Fig. 3a). This study focuses on the first four missions from disturbance to tropical storm (Table 1, Fig. 2). The first mission into Fanapi (F1) was 33–40 h prior to tropical depression (TD) formation. The system had convection mostly at its periphery (Fig. 4a) and an open wave structure at this time (Fig. 4b). The wave axis and a low pressure trough were oriented from the west-southwest to the east-northeast. The second mission (F2) was conducted 9–15 h prior to initiation in the JTWC best track. A nearly closed circulation was present at 850 and 700 hPa (Fig. 4d), and the deep convection was more concentrated near the center (Fig. 4b). The wave axis and a low pressure trough were oriented from the west-southwest to the east-northeast. The second mission (F2) was conducted 9–15 h prior to initiation in the JTWC best track. A nearly closed circulation was present at 850 and 700 hPa (Fig. 4d), and the deep convection was more concentrated near the center (Fig. 4b). According to the best track, the storm attained minimal tropical storm (TS) intensity (35 kt) during F3, and it had an intensity of 55 kt during F4. During F3 and F4, a single large cloud cluster was present, which could be tracked moving generally west-northwestward until 17 September, which was the central dense overcast of Fanapi (Fig. 3a). As expected, the aircraft encountered a well-defined circulation center with convection near the center during F3 and F4 (Figs. 4e–h).

Typhoon Megi formed 380 km west-southwest of Guam at 1800 UTC 12 October 2010 (Fig. 2b). The precursor invest system and the associated mesoscale cloud clusters passed just south of Guam (Fig. 3b). The long-range National Weather Service radar from Guam indicated widespread stratiform rain associated with the cloud clusters passing to the south (not shown). According to the JTWC best track, the system was a TD when the C-130 first reached it around 1800 UTC 12 October (M1); however, it was not declared a TD in real-time operational JTWC advisories for another 12 h. Aircraft data show a well-defined circulation center with convection organized near the center during M1 (Figs. 5a, b). According to best track, Megi developed into a tropical storm at 0600 UTC 13 October, soon after M1. Not surprisingly, a single coherent cloud cluster could be tracked moving to

**FIG. 10.** As in Fig. 9, but for the dropsonde released at 0109 UTC 14 Sep 2010 during F2.
the west-northwest for several days, which is the central dense overcast of Megi (Fig. 3b). During M2, it had reached a best-track intensity of 55 kt. A tight circulation with convection concentrated at the center was found at this time (Figs. 5c,d). Note that, during M1, the storm was at a similar stage of development as Fanapi was during F3: a borderline TD/TS. When the C-130 reached Megi during M2, it was already a tropical storm with intensity of 55 kt, similar to Fanapi during F4. Therefore, the C-130 missions considered in this study included a precursor disturbance (15 kt), a borderline TD (20–25 kt), two minimal TSs (35 kt), and two 55-kt TSs.

In both cases, there were multiple MCSs (208-K cloud clusters) that existed simultaneously and evolved in time (either continuously or discontinuously) over several days prior to TC genesis (Figs. 3a,b). This mode has been commonly observed over the western Pacific (e.g., Kerns and Chen 2013).

4. Thermodynamic evolution of TC genesis

a. Precursor disturbance stage: Pre-Fanapi

F1 sampled the precursor disturbance of Fanapi (Figs. 4a,b). The dropsonde winds indicated an open wave structure with a southwest-to-northeast tilt (Fig. 4b). The convection associated with pre-Fanapi was broad (Fig. 4a) and concentrated in several separated cloud clusters (Fig. 3a). The wave axis bisected the area of convection, though there was a ~300-km-wide area of weaker convection along the wave axis centered near 14°N, 138°E (Fig. 4a). The low pressure center, as determined from the dropsonde data, was located between this clear-air area and the area of convection to the southeast (Figs. 4a,b). Onion profiles were found throughout the disturbance—in particular, near the incipient low pressure center.

The onion thermodynamic structure of subsidence warming is illustrated by the dropsonde released near the circulation center at 0139 UTC during F1 (Fig. 6). Below the melting level (~550 hPa), there is expected to be a competition between evaporative cooling and unsaturated adiabatic subsidence warming. From the melting level down to 800 hPa, the relative humidity decreases from near 100% to below 70%, and temperatures are up to 2°C cooler than the reference profile (Fig. 6d). This cooling relative to the environment is related to evaporation as the rain falls from the melting layer into the lower-RH air. At 850 hPa, the adiabatic subsidence warming becomes dominant, and the profile
becomes slightly warmer than the reference profile. This is also the level where the driest (i.e., the fattest) part of the onion is found (Fig. 6b). There, the relative humidity is reduced to below 70%, which is significantly drier than the reference profile mean RH of 80%–85% (Fig. 6c). While the onion profile was not as intense (i.e., large onion) as observed in strong tropical squall-line systems (Zipser 1977) or midlatitude systems (Smull and Houze 1987), it was associated with a distinct minimum in relative humidity (Fig. 6c) and temperatures several degrees Celsius warmer than without the onion-shaped profile (Figs. 6b,d). Finally, below 900 hPa, the boundary layer is moister than the onion, with relative humidity around 80% (Fig. 6c). The cooler boundary layer air is likely modified by cool air from saturated convective downdrafts (Zipser 1977).

The moist static energy (MSE) profile is at a minimum at around 850 hPa, which is the same level as the onion feature (Fig. 6c). Typically, the minimum MSE is found in the mid troposphere at around 600 hPa, as depicted in the Guam mean MSE profile (black line). Since MSE is conserved during adiabatic descent, the fact that the minimum is located significantly lower in the troposphere is consistent with adiabatic subsidence.

A dropsonde was released at 0235 UTC 13 September near the incipient center at the edge of the convective anvil clouds (Fig. 7a). The thermodynamic profile suggests that the low pressure minimum was related to warming aloft superposed with subsidence warming in the lower troposphere (Fig. 7b). The onion layer was located between 900 and 700 hPa. In addition to the onion warm anomaly in the lower troposphere, the upper levels above ~500 hPa were warmer than the reference profile, with a nearly moist adiabatic temperature profile (Figs. 7b–d), likely because of latent heat release in the upper-level anvils. Similar to the onion profile discussed above, the reduced MSE extends down to ~900 hPa (Fig. 7e).

The upper-level warming at the center may not be enough to induce surface pressure falls (relative to the

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**Fig. 12.** As in Fig. 7, but for the dropsonde released at 0041 UTC 15 Sep 2010 during F3. The contours in (a) with low (L) and high pressure (H) labels are as in Fig. 4f.
Guam mean SLP) if it were to be accompanied with lower-troposphere cool anomalies without the subsidence warming. For example, a dropsonde released in the area of convection to the northeast of the pressure center at 0024 UTC 13 September (Fig. 8a) shows warm anomalies above 550 hPa but cool anomalies of $-1^\circ$C below (Fig. 8). This profile was nearly saturated from 900 to 600 hPa (Figs. 8b,c), and it is likely that the lower-troposphere cool anomalies of around $1^\circ$C cooler than the reference profile are due to the dominance of evaporative cooling in the heavier rain and convective downdrafts. Unlike in the onion profile cases, the MSE profile does not have as pronounced of a minimum in the lower troposphere, suggesting that subsidence warming is not a significant factor there (Fig. 8e).

In addition to the individual soundings described above, one common characteristic of the dropsondes released during F1 was a relative minimum in relative humidity at 700–800 hPa, collocated in the vertical with a relative maximum in temperature anomaly (gray curves in Figs. 6–8). The predominance of anomalously dry and relatively warm air at around 800–700 hPa with relatively low MSE suggests a history of subsidence below the melting layer. This thermodynamic structure is ubiquitous throughout the disturbance in part because heavy rain encompasses only a small fraction of the area of the disturbance. The subsidence warming has a mitigating effect on the low-level cool anomalies, which would dominate if evaporative cooling were dominant, such as in heavier stratiform rain and convective rain. If further lower-troposphere subsidence warming were to occur, the cool anomaly could be further eroded and/or transformed into a warm anomaly. Also, additional anvil warming above 500 hPa could help induce further surface pressure falls.

b. Tropical depression stage: Fanapi

The F2 mission into Fanapi was conducted several hours prior to the initiation of a TD in the JTWC best track. There was much more convection near the center during F2 compared with F1 (Figs. 4a,c). The dropsonde
data indicated a closed or nearly closed circulation center at 850 and 700 hPa with a wavelike structure at 500 hPa (Fig. 4d). Therefore, F2 is considered to have sampled a borderline TD.

Similar to F1, the onion thermodynamic structure is found near the incipient center during F2 with a distinct minimum in RH at 700–600 hPa (Fig. 9). In general, dropsondes during F2 had an RH minimum of ~80% between 600 and 800 hPa (gray curves in Fig. 9c). The minimum MSE corresponds with the onion feature, with a minimum at around 700 hPa consistent with a history of downward adiabatic transport from the midlevel MSE minimum (Fig. 9e). Note that the MSE was significantly higher than near the center in F1 (Fig. 7e), likely because of an overall increase in midlevel moisture near the incipient center (Wang 2012). A significant structural thermodynamic difference between the first and second flights into Fanapi was that the cool anomaly at around 600–700 hPa was significantly reduced throughout the system, with temperatures warming to near the reference profile there (Figs. 8d and 9d). This suggests that subsidence warming can erode the lower-troposphere cool anomalies from above, though the detailed processes involved are not fully resolved by the available data. Note that at the low levels (850 and 950 hPa), cool anomalies from −1.5°C to −1.0°C remained similar to F1 (Figs. 8d and 9d).

Another significant change between F1 and F2 is that the upper levels (above 500 hPa) became somewhat warmer (Figs. 8d and 9d). However, the largest upper-level warming occurred in the active convection area to the northeast of the low-level center (Fig. 10). The dropsonde released at 0109 UTC shows a nearly saturated profile with a warm, moist adiabatic profile above 500 hPa (Figs. 10b,c), suggesting the importance of latent heat release there. MSE was higher there than at the

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1 The dropsonde in Fig. 10 likely experienced sensor wetting errors, especially below 500 hPa.
center, likely because of upward transport of high-MSE air near the surface (Fig. 10e). The largest surface pressure falls and low-level circulation center during F2 were located not where upper-level warming was maximized, but where the upper-level warming was superposed with lower-troposphere subsidence warming and erosion of the lower-troposphere cool anomaly from above (Fig. 9). Also note that there is a clear deep cyclonic circulation extending down to 850 hPa as observed by the dropsonde data, which suggests a system-mean low-level convergence despite the presence of subsidence warming indicated by the onion-shaped soundings at some locations within the system.

Spatial maps of analyzed constant pressure level temperatures show that, between F1 and F2, the warm anomalies at 350 hPa increased by 0.5°–1.0°C (Figs. 11a,b). Recent modeling results from Cecelski and Zhang (2013) suggest that upper-level warming is important for TC genesis. However, at the early stage, the center is sensitive to the lower-troposphere subsidence warming. The temperatures below the melting level (~550 hPa) at 650 hPa increased by 1°–2°C (Figs. 11d,e). During F1, the incipient center was located where the cool anomalies were less pronounced. The cool anomalies at the center were 0.5°–1.0°C lower than the reference profile temperature, compared with 2.0°C cooler nearby to the southwest. Also, the warmest air at 350 hPa was to the southwest of the center (Fig. 11a), where the 650-hPa temperatures were ~1°C cooler than at the incipient center (Fig. 11d). By the time of F2, the upper levels were warmer (Fig. 11b), and the cool anomalies at 650 hPa were nearly eliminated (e.g., less than 0.5°C cooler than the Guam mean; Fig. 11e). Both of these thermodynamic changes played a role in the initial formation of the initial low-level closed circulation (Fig. 4d). Further intensification to a minimal tropical storm involved warming both at 350 and 650 hPa and vertical alignment as observed during F3 (Figs. 11c,e).

c. Minimal tropical storm: Fanapi and Megi

According to the JTWC best track, Fanapi became a tropical storm during F3, and Megi attained this status during the first mission into that storm (M1). Subsidence
warming was found below the melting level within the developing central dense overcast, especially at the periphery. The sonde deployed nearest to the center at 0042 UTC 15 September during F3 (Fig. 12a) shows subsaturated conditions both above 500 hPa and below 650 hPa (Figs. 12b,c). Adiabatic subsidence warming may play a role around 700 hPa, where both the relative humidity (Fig. 12c) and MSE are at a minimum (Fig. 12e). The profile was significantly warmer than the Guam mean above 750 hPa (Fig. 12d), where the temperature profile followed a nearly moist adiabatic profile (Fig. 12b). This is consistent with recent studies showing that latent heat release aloft plays a large role in surface pressure falls near the center (Zhang and Zhu 2012; Cecelski and Zhang 2013). However, the subsaturated conditions above 500 hPa (both with respect to water and ice) would not be consistent with warming solely because of latent heat release. Interestingly, at around 450 hPa there is a minimum in relative humidity of ~60% (Fig. 12c) and maximum in temperature anomaly (Fig. 12d), as well as a fatter appearance on the skew T diagram (Fig. 12b), similar to an onion subsidence warming feature. The height of this feature would be inconsistent with mesoscale unsaturated downdrafts below the melting layer and may be indicative of a descending inflow jet. Note that most sondes during F3 had a subsaturated RH minimum of around 85% and a broad MSE minimum between 850 and 650 hPa (gray lines in Figs. 12c,e), indicating that adiabatic subsidence warming is likely counteracting the evaporative cooling beneath the melting layer throughout the system—in particular, in the outer circulation portion of the storm. A sonde released north of the center but still beneath the central dense overcast (Fig. 13a) at 2208 UTC 14 September shows a narrow onion profile below the melting layer (Fig. 13b). There is a MSE minimum there (Fig. 13e), suggesting unsaturated adiabatic warming at the periphery of the developing circulation center.

In M1, the dropsonde released near the center beneath the central dense overcast (Fig. 14a) suggests the superpositioning of warm anomalies above 500 hPa and an onionlike feature just below 700 hPa (Fig. 14b). The
sonde was released beneath cloud-top temperatures below 200 K (Fig. 14a). Above ~650 hPa, the profile is saturated (with respect to ice) and nearly moist adiabatic. At 700 hPa, the relative humidity is reduced to 82% (Fig. 14c), and temperature anomalies are slightly positive (Fig. 14d). There is also an MSE minimum near 700 hPa (Fig. 14e). On the periphery of the system, the onion features were more pronounced: for example, to the southwest of the center (Fig. 15). Interestingly, the sonde released at 2318 UTC 12 October has a double onion structure with distinct onion features at 650 and 800 hPa (Fig. 15b). This may suggest subsidence warming due to both descending jet features and mesoscale unsaturated subsidence. Similar to TS Fanapi, the lower-troposphere warming due to subsidence may contribute to maintaining and intensifying the outer circulation of the storm (if it is widespread), although upper-level warming may play a more important role at the center.

Note that, in F3 and M1, the tight center of the minimal tropical storms was not adequately resolved by the dropsonde data. Also, distinct descending inflow jets are not resolved. The subsequent missions, however, were conducted at 700 hPa and resolve some features of the lower-troposphere component of the developing warm core in more detail, as will be discussed in the following section.

d. Intensifying tropical storms Fanapi and Megi

Fanapi and Megi were both sampled by the C-130 as intensifying tropical storms with an intensity of 55 kt. In both missions, the flight track passed through the central dense overcast and the circulation center (Figs. 4d and 5b), providing valuable flight-level data below the melting layer. The first transect considered here is from the northwest to the southeast during F4, from 0215 to 0330 UTC 16 September (Fig. 16). The circulation center at 700 hPa was located at the eastern periphery of the large cloud system (Fig. 16a). As the aircraft flew from the northwest approaching the center, it encountered mostly subsaturated air with the temperature within 18 C of the Guam mean flight-level temperature. The warm core near the circulation center consisted of air at 14°–18°C, compared with the Guam mean of 9.3°C. The dewpoints were around 7°–9°C. This warm air with large dewpoint depressions near the center suggests adiabatic subsidence warming. Dropsondes released there show dry air and a temperature inversion capping the nearly saturated boundary layer at the center (Fig. 16d), which
is similar to mature TCs. As the aircraft moved southeast away from the circulation center, it encountered cooler and saturated air.

The second transect considered here was taken during M2 from the northwest to the southeast during 0320–0500 UTC 14 October. At this time, convection had increased, and a distinct central dense overcast enveloped the circulation center (Figs. 17a, 18a). The aircraft made several loops at the center for a period of around half an hour (dashed vertical lines in Fig. 17b). Similar to the first transect, the warmest air was within the subsaturated part of the flight leg. The warmest air encountered was 19.1°C with a dewpoint of 2.9°C. Dropsondes released within the warm-core region suggest vigorous subsidence reminiscent of soundings taken within mature TC eyes (Figs. 17c,d). In contrast, a dropsonde deployed to the southeast of the warm core within the central dense overcast region was saturated and significantly cooler than the reference profile below 800 hPa (Fig. 17e).
Passive microwave satellite data from an overpass of the Advanced Microwave Scanning Radiometer for Earth Observing System (AMSR-E) during M2 suggest that the strongest convection was in an arc at the western and northern part of the cold cloud area displaced 50–100 km from the circulation center (Fig. 18). A large area of moderate rain extended to the southeast and south of the arc of vigorous convection, which is inferred from the moderate 85-GHz scattering signal (i.e., green; Fig. 18b) and cyan colors on the 37-GHz color composite (Hawkins and Velden 2011; Kieper and Jiang 2012). The low-level center and subsaturated warm air encountered by the C-130 around 0400 UTC was at the periphery of the moderate stratiform rain area, which may have evolved from a feature similar to the wake low in Chen and Frank (1993), as depicted in Fig. 1. The cyan spiral band features in Fig. 18c converge inward toward this center. This structure of the developing center displaced from the strongest convection and at the edge of the stratiform rain area is similar to that of Dolling and Barnes (2012).

5. Hydrostatic contribution to the SLP fall

To quantify the significance of the lower-troposphere subsidence warming for SLP falls relative to the Guam mean profile, the hydrostatic equation is used as described in section 2f. The vertical profiles in Fig. 19 show how much each vertical level would contribute to making the SLP be higher or lower than the Guam mean profile. Pressure fall contributions are negative, to the left of the vertical dashed line. The net SLP difference is the vertical integral of each level’s contribution (plus unsampled contributions above flight level). The thermodynamic changes related to the development of the storms are related to the variations from the left panels to the right panels, which are in order of the degree of storm development. Note that, for each individual flight, the pressure falls relative to the immediate storm environment—which may be cooler or warmer than the Guam mean—are not known. In the early stage of TC genesis (F1), SLP fall contributions are mainly from the upper levels (>4.5 km; Figs. 19a,b). These warm anomalies and associated contributions to surface pressure falls are likely caused by the concentration of latent heat release in the increasingly inertially stable environment (Chen and Frank 1993; Zhang and Zhu 2012). This is a preexisting feature for several days prior to genesis, and it becomes more pronounced as the system develops (Fig. 19b,c). Below 4.5 km, SLP contributions are mostly positive, which counteracts the SLP falls induced from the upper levels (Fig. 19a). However, at ~2 km (~800 hPa), there is a relative minimum within the low-level cool anomaly (Fig. 19a). The lower-troposphere cool anomalies are partially mitigated at around 800 hPa compared with 700 hPa, likely in part because of subsidence warming based on the onion-shaped profiles in Figs. 6 and 7. Nevertheless, the low-level cool
anomaly remained a road block for TC genesis in F1. It is unlikely that the observed increased anvil warming above the melting layer (550 hPa) in F2 and F3 could result in a well-defined TC if the positive pressure tendencies of up to $+0.5 \text{ hPa} \text{ km}^{-1}$ below 650 hPa were to remain. Indeed, these pressure rise contributions were mostly eliminated in F2 (Fig. 19b) and replaced by pressure falls of up to $-0.5 \text{ hPa} \text{ km}^{-1}$ by the time of F3 (Fig. 19d). Many of the dropsonde thermodynamic profiles for those flights were also consistent with subsidence warming (Figs. 9, 13, and 14), suggesting that subsidence warming may play a role in reducing the lower-troposphere cool anomaly, which makes it more favorable for TC genesis.

As TC genesis proceeds, the low-level cool anomaly is eroded from above. For the second flight into Fanapi, when the system was a borderline TD, most dropsondes show significant contributions to falling SLP above 2 km ($\approx 800 \text{ hPa}$), with some counteracting SLP rise contributions from the remaining near-surface/boundary layer cool anomaly (Fig. 19b). In terms of the contribution to the SLP perturbations, the most significant structural thermodynamic difference between the first and second missions into Fanapi is the transition from positive SLP contributions to neutral/SLP fall contributions between 2.5 and 4.5 km (Figs. 19a,b). The erosion of the low-level cool anomalies allows the net surface pressure perturbation to be negative, inducing a well-defined low-level pressure center (Fig. 4d). This erosion of the low-level cool anomaly roadblock from above results in the transition to a borderline TD. By the time of F3, the lower-troposphere contributions of $0.2$–$0.5 \text{ hPa} \text{ km}^{-1}$ over $\approx 6 \text{ km}$ can account for several hectopascals in surface pressure falls, as opposed to counteracting the upper-level contributions in F1. During M1, the pressure fall contributions were similar to F3. Note that the cool anomalies remained near the surface, but they were no longer sufficient to counteract the effect of the warming above, resulting in falling SLP.

For the F4 and M2, SLP fall contributions were present in most profiles above $\approx 1 \text{ km}$ (Figs. 19f,g), with pressure fall contributions of around $3$–$4 \text{ hPa} \text{ km}^{-1}$ over several kilometers in the lower troposphere. These large pressure fall contributions are related to the eyeclike low-level structures seen near the centers of these intensifying tropical storms (Figs. 17 and 18). Similar to the results of Dolling and Barnes (2014), the largest pressure fall contributions were from $\approx 2 \text{ km}$ altitude.

6. Conclusions

The aircraft observations from ITOP have revealed an important but previously underappreciated ingredient in TC genesis: namely, the subsidence warming within MCSs. The GPS dropsonde measurements from ITOP show that characteristic onion-shaped thermodynamic profiles are common in the lower troposphere prior to and during TC genesis. This observation helps shed some light on a previously unresolved issue in TC genesis. Prior to TC genesis, lower-troposphere cool anomalies associated with evaporative cooling within the stratiform regions in MCSs as shown in Bister and Emanuel (1997) counteract the upper-level warm anomalies described by Chen and Frank (1993), which inhibit the net hydrostatic surface pressure falls. The subsidence warming, especially from $600$ to $800 \text{ hPa}$ beneath the melting layer as shown by the dropsonde data in this study, helps erode the lower-troposphere cool anomalies. When superimposed with the mid-upper-troposphere warming induced by latent heating and enhanced by inertial stability in long-lived large MCSs (Chen and Frank 1993), it can help to reduce the surface pressure, which is critical for TC development. The lowering of the level of minimum MSE throughout the system, especially near the center in F1, F2, and F3, is consistent with a history of subsidence, while the system moistening may partially counteract the drying effect. The subsidence warming in the low troposphere is one of multiple key ingredients of TC genesis, and it would not be expected to cause TC genesis on its own—in particular, without upper-level warming (Zhang and Zhu 2012; Cecelski and Zhang 2013). The vertical alignment of the mid-upper-level warm anomalies with the low-troposphere warming below the melting layer is a crucial component of TC genesis for the cases observed in this study. The results of this study are summarized below and in Fig. 20:

- Onion-shaped thermodynamic profiles indicative of lower-troposphere subsidence warming were ubiquitous in pre-Typhoons Fanapi and Megi during ITOP from the disturbance to the tropical storm stage.
- TC genesis occurred when the subsidence warm anomalies eroded the lower-troposphere cool anomaly and superposed with the preexisting mid-upper-troposphere warm anomaly (Fig. 20a) in regions favorable for TC genesis.
- As Fanapi and Megi intensified into strong tropical storms, there was evidence of vigorous localized subsidence warming similar to the lower-troposphere structure of mature TC eyes (Fig. 20b).

The flight-level data during the intensifying TS phase of Fanapi and Megi suggest a deepening, intensification, and scale contraction of the subsidence warming during intensification to a more mature system, similar to that in Dolling and Barnes (2012), in which the center...
develops where there is enhanced lower-troposphere subsidence in the light stratiform rain area at the periphery of the deep convection. One remaining question is how the lower-tropospheric warming intensifies during the transition from a TD to a TS. A possible factor could be the interaction of the multiple MCSs or 8-h clusters (Figs. 3 and 4), similar to that shown in Kerns and Chen (2013). In addition to the interaction and intensification of the midlevel vortices (Ritchie and Holland 1997; Simpson et al. 1997), the lower-troposphere subsidence warming can potentially be enhanced if the MCSs are arranged in such a configuration that their areas of lower-troposphere mesoscale subsidence reinforce each other (Fig. 20).

It is important to note that the lower-troposphere subsidence warming only occurs in the context of large, long-lived, organized MCSs, which contain deep convective cores or “hot towers.” These deep convective cores can help maintain the system mean convergence and ascent (Riehl and Malkus 1958; Malkus and Riehl 1960; Wang 2014) and also provide a source of low-level relative vorticity (Hendricks et al. 2004). Therefore, the mesoscale subsidence need not cause the disturbance to dissipate as long as convective cores are maintained elsewhere within the system. Indeed, the observations obtained in the pre-TCs showed a clear deep system mean cyclonic circulation and low pressure center extending down to 850–900 hPa (Figs. 4d, 5b), while the subsidence warming is indicated by the onion-shape soundings within the system. The aircraft data available in this study cannot resolve these convective scale processes, including convective updrafts and downdrafts, which persist and may intensify throughout TC genesis. Similarly, the data presented in this study cannot definitely determine how widespread the mesoscale subsidence warming is. A high-resolution numerical modeling study has been conducted to help clarify the roles of the convective scale features in relation to the lower-troposphere subsidence warming. The results will be described in Part II of this study.

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REFERENCES


